

Characterization of catastrophic flood-related features in the English Channel

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Abstract

Megaflow flows are global-scale phenomena that can have a significant impact on Earth's landscapes and climate. The low frequency occurrence of these events prevents their direct observation. The analysis of ancient megaflow terrains and bedform associations are the primary resources for the understanding of catastrophic flows. Previous studies have proposed the palaeo-channel network present at the English Channel seafloor to have been carved by catastrophic flows.

This thesis investigates the origin of the English Channel palaeovalley network, testing the hypothesis of its catastrophic formation and aiming to reconstruct the relative timing and magnitude of the events that formed the palaeo-channel system. Seismic reflection data reveal the presence of depressions up to 100 m-deep carved into bedrock at the proposed spill-point. New high-resolution bathymetric data show kilometre-scale channels and associated erosional bedforms that extends for more than 200 kilometres from the Dover Strait to the Central English Channel. The stratigraphic and geomorphologic analyses indicate the observed features are similar to other flood-generated features from other well-established megaflow terrains on Earth. The quantitative characterisation of the bedforms presented in this work, together with interpretation of cross-cutting relationships of the mapped erosional surfaces, allow for the reconstruction of the relative history of the catastrophic events that carved the palaeovalley network and led to the separation of Britain from Europe through the breaching of the rock ridge present at the Dover Strait. The detailed interpretation of the 100 m-deep bedrock depressions located at the proposed breach point allows for a reconstruction of the rapid erosion of the Dover Strait. Geomorphologic analyses of the palaeo-channel network and its associated bedforms and relation to bedrock geology are presented in this work thanks to available high-resolution bathymetry and seismic reflection data. Finally, the magnitude of the palaeo-flows, estimated through palaeo-hydraulic calculations, gives further evidence of the catastrophic nature of the flows.

The proposed model and relative history of the English Channel megaflow is presented in agreement with palaeo-geographic and palaeontological studies previously carried out in the study area, contributing to the reconstruction of the Pleistocene palaeo-geography of north-western Europe.

I hereby declare that the research presented in this thesis is my own work and that all mention to work carried out by others is appropriately referenced.

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1 . Introduction

1.1 Motivations of this study

The first megaflood terrain discovered on Earth is represented by the Channeled Scablands megaflood complex, in the State of Washington US, described by J. H .Bretz (1923) as a landscape produced by “a flood in which the sudden release of a volume of water much larger than that which annually flows through the area produced the very large scale erosion”. Following Bretz’s discovery, several megaflood terrains have been identified on Earth, associated with the presence of melting ice-sheets draining into glacial lakes located at their margin (Teller, 2003). The flood-water, from catastrophically drained from the dammed glacial lakes, flows through the fluvial drainage network to the Ocean, eroding characteristic associations of flood-generated features (e.g. Bretz et al., 1956; Baker, 1978a, 2009a; Carling et al., 2009a, 2009b) that are present for hundreds of kilometres from the ice-sheet margin (Carling et al., 2009a). During megafloods, the river network undergoes strong erosion, as the dimensions of typical fluvial systems cannot accommodate the volume of water involved in flood processes (Baker, 2009b) that, as suggested by the dimensions of the observed bedforms, have discharges comparable to ocean currents (Baker, 1994). As a result, great quantities of freshwater are conveyed from the ice-sheet to the Ocean (Kourafalou et al., 1996; Mulder et al., 2003), altering water temperature, salinity and the ocean currents that regulate Earth climate (e.g. Barber et al., 1999; Clark et al., 2001; Teller and Leverington, 2004).

As megaflood processes are related to ice-sheet decay mechanisms that only happen every thousands of years, such events cannot be directly observed or predicted. The study of megaflood landscapes and bedform associations produced by ancient floods is therefore crucial for the understanding and reconstruction of the processes and mechanisms that formed these features. Despite a detailed knowledge of typical flood-associated landforms that has been gained through the study of ancient flood landscapes (e.g. Carling et al., 2009a, 2009b), many aspects of megafloods are still poorly known, as a result of the lack of direct observation. For example, although erosional and depositional flood features have been studied in detail by a number of authors in the last decades (e.g. Bretz, 1923, Baker and Nummedal, 1978; Rudoy, 1998; Meinsen et al., 2011), factors like erosional power

distribution throughout floods (Costa and O'Connor, 1995) or timing of deposition and/or erosion required to develop specific landforms (Carling et al., 2009a, 2009b) are poorly constrained. Intensive efforts have also been made in designing palaeo-hydraulic models (Partridge and Baker, 1987; O'Connor and Webb, 1988; Enzel et al., 1994; Wohl et al., 1994; Kehew and Teller, 1994; Carling et al., 2010) capable to reconstruct peak discharge and magnitude of ancient floods, with the aim of understanding their possible global-scale impact on Earth landscape and climate. These models suffer due to a lack of constraints. For example, little is still known about regional-scale deglaciation mechanisms and ice-sheet melt-water runoff paths (e.g. Teller, 1990, 2003; Marshall and Clarke, 1999; Toucanne et al., 2009a), as well as megaflood hydraulic behaviour and parameters (Carling et al., 2009a). A better understanding of flood mechanisms and evolution is therefore required for the reconstruction of the potential impact and mechanisms through which these events affected climate past changes and future evolution patterns.

All major megaflood terrains identified on Earth are associated with the presence of glacial dammed-lakes that formed at the margin of continental ice-sheets during Quaternary glaciations. Such lakes were present at the margin of North American Laurentide Ice Sheet (Kehew and Teller, 1994; Clarke et al., 2004) and Cordilleran Ice Sheet (Bretz, 1923; Baker, 1978a, 2009a), as well as Eurasian Ice Sheet (e.g. Baker et al., 1993; Carling et al., 2002; Rudoy, 2002). It has also been proposed that the coalescence of British and Scandinavian Ice Sheets (Oele and Schuttenhelm, 1979) resulted in a bedrock-dammed pro-glacial lake located in the Southern North Sea at the ice-sheet southern margin (Smith, 1985; Gibbard, 2007; Gupta et al., 2007). Failure of this dam caused a catastrophic flood that carved a system of incised channels known as Northern Palaeovalley into the English Channel seafloor (Gupta et al., 2007). This network of deeply incised channels which show great similarity to the Channeled Scablands flood channels (Gupta et al., 2007). Palaeo-geographic reconstructions indicate the Dover Strait was the flood spill-point (e.g. Smith, 1985; Gibbard, 1995; Gupta et al., 2007; Toucanne et al., 2009a; Meinsen et al., 2011) and suggest that the presence of the ice-sheet and associated pro-glacial lake would have had a strong impact on both the north-west European river drainage network configuration (e.g. Gibbard, 1988, 1995; Busschers et al., 2008; Gibbard and Clark, 2011) and early human occupation of south-eastern Britain during Middle and Lower Pleistocene (Stringer et al., 1998; Bridgland et al., 2006; Ashton and Hosfield, 2010).

This work builds on the results of Gupta et al. (2007) on the catastrophic genesis of the Northern Palaeovalley channel network. This thesis utilises new compiled datasets for the detailed characterisation of the interpreted flood-generated features, aiming for the understanding of the erosional mechanisms, magnitude and relative timing of the events that generated the observed bedforms. The results of these analyses results permit to test the megaflood hypothesis for the erosion of the palaeovalley network. Finally, this work develops new insights on the Quaternary palaeo-geographic and palaeontological evolution of the Dover Strait area, reconciling the inferred megaflood history with the evolution of north-western European river network configuration and landscape evolution during Pleistocene (e.g. Gibbard, 1985, 1995; Bridgland and D'Olier, 1995; Busschers et al., 2008; Toucanne et al., 2009a; Ashton and Hosfield, 2010), addressing the issue of global and regional-scale impact of megafloods on Earth landscape and climate.

1.2 Aims and Objectives

The research aims were defined by the following questions:

1) Does the morphology of the seabed show evidence for catastrophic erosional processes? The formation of large scale erosional features, such as the bedforms observed in the English Channel by Gupta et al. (2007), can hardly be explained by the action of common fluvial processes. Typical river processes cannot erode kilometre-scale features into bedrock. The analysis of new compiled high-resolution bathymetric datasets that extend from the Dover Strait to the downstream area allows for the investigation of the erosional features present in the Northern Palaeovalley, in order to test the proposed hypothesis of a megaflood-related origin of these bedforms. This was achieved through the comparison of the interpreted bedforms to typical megaflood-generated features, such as plunge pools, cataracts, streamlined islands and erosional lineations previously observed in other megaflood terrains on Earth and Mars (e.g. Komar, 1983; Komatsu and Baker, 1997; Burr, 2005).

2) Are the infilled bedrock incisions present in the buried topography consistent with the megaflood hypothesis? The enigmatic depressions located at the Dover Strait and previously named the “Fosse Dangeard” (Destombes et al., 1975) play a key role in testing the flood hypothesis as the process responsible for the erosion of the Northern Palaeovalley

channel network. Deep circular depressions are among the most distinctive features eroded by high-energy megaflood flows, forming at the base of waterfalls and giant cataracts headwall (e.g. Baker, 1978b; Carrivick et al., 2004; Lamb et al., 2007). A detailed analysis of the geometry and infill of the incisions was carried out in order to constrain the nature of the processes that carved these features into bedrock.

3) Where was the megaflood spill-point area located? Spatial and temporal evolution of megaflood events is still poorly known (Carling et al., 2009a). The analysis of the flood-related bedforms present at the Dover Strait is a primary step towards understanding the evolution of the events that led to the insularity of Britain. This was pursued through: 1) testing the hypothesis of the Dover Strait was the flood spill-point 2) understanding the mechanisms and evolution of the erosional flows from the spill-point to the downstream area 3) reconstructing the palaeo-geographic history of the Dover Strait (e.g. Gibbard, 1988, 1995) before and after the megaflood events.

4) How many events carved the palaeovalley system and what is the relative chronology and magnitude of the individual floods that formed the observed bedforms? The relative chronology of the mapped erosional surfaces and their cross-cutting relationship allows for the reconstruction of the number and relative timing of the erosional events that carved the palaeovalley. Building on this, a palaeo-hydraulic reconstruction was carried out using a series of channel cross-profiles from the Dover Strait to the downstream area, allowing for an estimation of the peak discharge and magnitude of the processes that carved the channel network, permitting comparison with other megaflood terrains (e.g. Teller 1990; O'Connor and Baker, 1992; Teller et al., 2002; Benito and O'Connor, 2003).

In order to address the above research questions, the following objectives were targeted in this work:

a) Analyse the influence of the inherited bedrock geology and tectonic framework on the geometry and distribution of the observed flood-related landforms (Chapter 4). Different hypotheses have been proposed on the possible influence of tectonic structures on the formation of the Fosse Dangeard depressions (e.g. Destombes et al., 1975; Smith, 1985; Van Vliet-Lanoe et al., 2004). In this work, the interpretation of the processes that formed the Fosse Dangeard and the observed bedforms is accompanied by the analysis of the influence

of regional geology on the flood-features geometry and distribution. To do so, seismo-stratigraphic analyses were carried out together with geological mapping of the seabed performed throughout the entire study area on the regional bathymetric dataset.

b) Map the buried incisions located at the Dover Strait, interpreting the depression morphology and infill from seismic reflection data (Chapter 5). The seismic reflection interpretation of the internal structure and stratigraphy of the Fosse Dangeard incision infill and basal surface is a primary objective in the understanding of the processes that led to the formation of the depressions. Detailed seismic reflection stratigraphic analyses permit an almost three-dimensional reconstruction of the geometry of the incisions, allowing for the analysis of the architecture and geometry of the infill seismic-stratigraphy and internal erosional surfaces. This allows for the identification of the number of erosional events that carved the depressions through the identification of distinct erosional surfaces. Finally, the detailed reconstruction of the incision geometry permits for a comparison with similar erosional features previously described in literature (e.g. Lamb et al., 2008; Pagliara et al. 2008a, 2008b).

c) Map the seafloor geomorphology from high-resolution and regional-scale bathymetry datasets, targeting the flood-generated landforms (Chapter 6). Deeply incised channels, knick-points and other typical flood features such as erosional grooves and streamlined islands are the targets of geomorphologic and geomorphic analyses carried out in this work. These interpreted features allowed for the comparison to other megaflood-related features described in literature. Finally, the analysis of flood palaeo-indicators such as streamlined islands and longitudinal lineations permit the identification and reconstruction of palaeo-flow path and direction.

d) Compare scale, morphology and spatial distribution of the interpreted flood-bedforms and the calculated flood peak discharge to other megaflood events described in literature (Chapter 6). Firstly, detailed morphometric analyses were carried out on the mapped streamlined islands and longitudinal lineations in order to compare the observed bedforms to similar features observed elsewhere. Morphometric parameters previously used for the characterisation of flood features (e.g. Baker and Kochel, 1978; Komar, 1983, 1984; Spagnolo et al., 2010) were employed in these analyses, in order to compare the obtained values with typical flood features observed in other megaflood terrains. Secondly, geometric

and hydraulic parameters were estimated from a series of cross-channel profiles that were employed for palaeo-hydraulic reconstructions from the spill-point to the downstream area.

1.3 Thesis structure

This thesis is composed of eight chapters, starting with a review (**Chapter 2**) of the existing literature of the main research themes treated in this work. The chapter begins with an introduction of megaflood events and description of major megaflood terrains known on Earth, followed by the description of the English Channel area, its geology and palaeo-geography. **Chapter 3**, which is divided into two sections, details data and methods used in this work. The first section describes the high resolution seismic reflection datasets available in this work and the methodology used for the seismic stratigraphic analyses. The second section is dedicated to the available bathymetric datasets, describing the methods used for data gridding and analyses of each dataset. **Chapters 4-6** describe the results obtained in this research, showing how the research aims and objectives have been addressed, each chapter targeting a different aspect of the research and including an introduction followed by results and discussions. In particular, **Chapter 4** consists of the work carried out on the regional and seabed geology using bathymetric and seismic reflection data. The chapter also shows how bathymetric data can be used to characterise the geology of the seabed at different scale (ranging from m to km), showing a very good correspondence between the character and texture of onshore outcrops and acoustic response of each lithology. **Chapter 5** consists of the seismic reflection analyses performed on the bedrock depressions located at the Dover Strait, illustrating their 3D geometry, distribution and analyses of their stratigraphic. The interpretation of the nature and formation of the mapped features is also detailed in this chapter and related to the formation of the megaflood channel network further examined in **Chapter 6**. This chapter describes the geomorphology of the Northern Palaeovalley channel network using bathymetric data. Geomorphologic and morphometric analyses of the mapped bedforms are detailed in this chapter, providing an interpretation and reconstruction of the relative timing of the events that formed the observed bedforms. In the final part of the chapter, palaeo-hydraulic reconstructions of the flood events are detailed, comparing the results obtained for the English Channel network to the events described in the literature review in Chapter 2. Finally, **Chapter 7** offers a final reconstruction of the interpreted flood model, discussing the obtained results in a wider perspective, with the aim of reconciling the

proposed model with the Middle and Late Pleistocene palaeo-geographic evolution of the study area. **Chapter 8** draws summary and conclusions of this work, proposing further steps to be carried out in future work.

2 . Literature Review

2.1 Introduction

The aim of this thesis is to characterise the geomorphology of the Northern Palaeovalley channel network and test the hypothesis of its catastrophic formation (Gupta et al., 2007), through the characterisation of the kilometre-scale erosional features mapped from bathymetric and seismic reflection data. The literature review proposed in this chapter is to 1) introduce the main background material 2) provide background information on the main aspects of this thesis, in order to give an overview of the main and most relevant studies that have been undertaken to date in the different fields of research 3) emphasize which aspect of the research requires further investigation and improvement and how this work can contribute.

In order to do this, the chapter has been divided into three sections describing different aspect of this work. The first section (Section 2.2) provides an introduction to terrestrial megafloods, starting with a description of these processes, and of the different sources and drainage mechanisms. This is followed by a description of the most typical features associated to megaflood terrains, subdivided into erosional and depositional bedforms. The chapter then provides a summary of the Pleistocene megaflood terrains recognised so far on Earth, concluding with a brief review of megaflood palaeo-hydrology. In this section, a description of the techniques used for the calculation of flood peak discharge is given, illustrating the most common limitations and pitfalls encountered by previous studies.

The second part of this chapter focuses on a regional characterisation of the English Channel geology (Section 2.3). In particular, the English Channel section describes the geology of the study area, starting with the characterisation of the geomorphology of the study area, followed by a brief history of the structural evolution of the area, illustrating the two main tectonic phases that characterise the formation of the basins that underlie the

Channel. The chapter continues with a discussion of the recent stratigraphy of the area. A brief summary of the tidal regimes and surficial sediments present on the English Channel seafloor is provided.

Section 2.4 opens with a detailed description of the Dover Strait area and morphology, illustrating the main depositional and erosional features observed at the seabed. The chapter continues with a focus on previous studies carried out on the Pleistocene palaeogeography of the Strait, reconstructing the ice-sheet limits during Middle and Late Pleistocene glaciations, and the influence of glacial isostasy.

The final section of the literature review chapter (Section 2.5) discusses the seismicity of the English Channel, reviewing the historical seismic events and discussing date and magnitude of the seismic events that affected the study area. A final consideration on the possible role of regional seismicity as a trigger for megaflood events in the English Channel is presented at the end of the section.

2.2 Review of Pleistocene megafloods

2.2.1 Overview of megaflooding on Earth

Megafloods are catastrophic events characterised by the sudden release of enormous volumes of water, with discharge comparable to ocean currents (Baker, 2001). A variety of megaflood terrains have been studied on Earth (Table. 2.1). These can be distinguished by different types of water bodies that were the water source. During Pleistocene glacial and inter-glacial times, the Northern Hemisphere was largely covered by ice: the presence of ice-sheets on the landscape changed the hydrological balance and the elevation of the continents (Marshall and Clarke, 1999). Water was stored both in the ice-sheets and in the glacial lakes located at the ice-body margin, causing dramatic reorganization of drainage network (e.g. Teller, 1990; Marshall and Clarke, 1999). Pleistocene glacial lakes are likely to have influenced regional climate (Hostetler, 1994) as well as the ice-sheet dynamics, through calving of wide regions located at the southern margin of the ice-sheet (Pollard, 1983) and acceleration of ice-sheet collapse. As a result, the evolution of the drainage system at late-glacial times was a function of melt-water generation, ice-sheet surface topography, continental topography, and ice-sheet retreat, usually characterised by a sequence of advance/retreat sequences (e.g. Mickelson et al., 1983; Clayton et al., 1985; Clark, 1994), which had strong impact on melt-water routing (Teller, 1990; Clark et al., 1996; Licciardi et al., 1999; Marshall and Clarke, 1999).

Ice-sheet – lake geometry

Costa and Schuster (1988) and Tweed and Russell (1999) have identified nine different types of lakes, depending on the configuration of the water body in relation to the ice, as shown in Fig. 2.1. Depending on the type of water body, different mechanisms for water outburst were identified. In particular, ice-dammed lakes will start to drain if critical thresholds are reached. These vary with lake topography and ice conditions (Herget, 2005) and are therefore difficult to classify. Although these are the most common type of dam bounding glacial lakes, none of these models apply to the case of the study area of this work, as illustrated in the result chapters. In relation to the water outburst mechanisms, Tweed and Russell (1999) distinguished eight types: some of these are related to external events, such as subglacial volcanic (jökulhlaup) or seismic activity, other are connected to variations in lake water level (overspill, Glen mechanism, ice dam floatation) or breaching of the ice (siphoning, cavity formation and sub-aerial dam breaches). A combination of the listed mechanisms is also possible, as conditions do not always remain stable and can change during lake drainage.

Letter (Fig. 2.1)	Type of ice-dammed lake	Typical drainage trigger	Typical duration of lake drainage
A	Tributary glacier dammed in trunk valley	Subaerial breaching	Hours/day
B	Pro-glacial lake	Overspilling, floatation, subglacial cavities	Weeks/months
C	Supraglacial lakes	Subglacial tapping	Month
D	Tributary glacier and trunk glacier dammed	Overspilling, floatation, subglacial cavities	Weeks/months
E	Ice marginal lake	Overspilling, floatation, subglacial cavities	Weeks/months
F	Converging glacier dammed	Overspilling, floatation, subglacial cavities	Weeks/months
G	Stream valley blocked by trunk glacier	Overspilling, floatation, subglacial cavities	Weeks/months
H	Subglacial caldera	Floatation	Hours/days
I	Volcanic waters	No storage	Weeks/months

Table 2.1: Description of the different types of ice-dammed water bodies shown in Fig. 1, from Tweed and Russell, 1999.

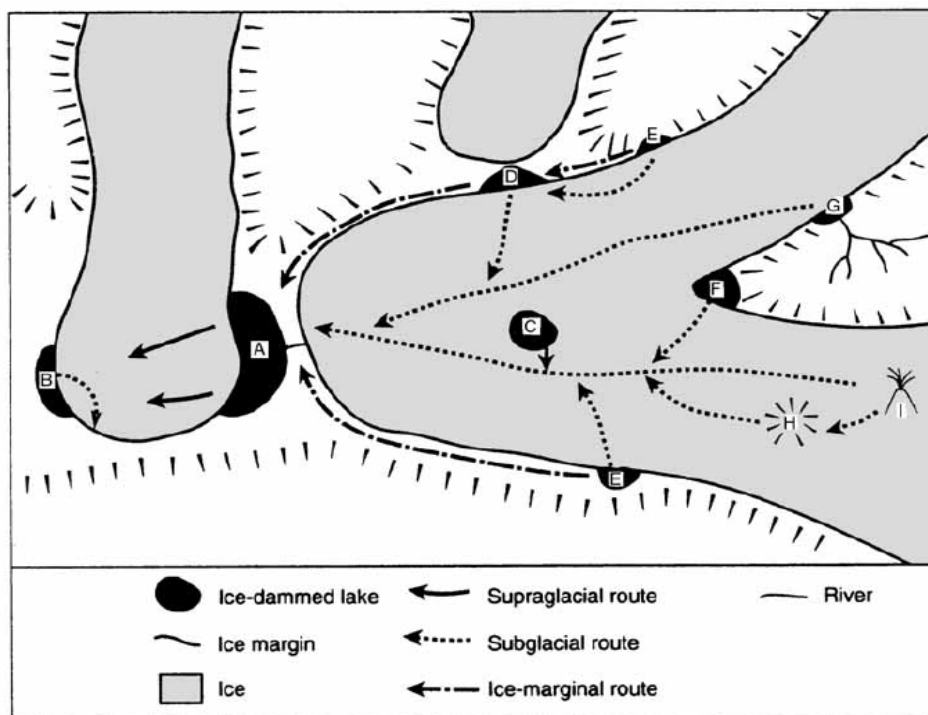


Figure 2.1: Settings of ice-dammed lakes and potential overburst routing explained in Table 2.1 (from Tweed and Russell, 1999).

Pro-glacial lakes (Blachut and Ballantyne, 1976) are the most important source of megaflood water. Pro-glacial lakes are dammed by the ice-sheet margin and form in the topographic depressions caused by ice loading (e.g. Marshall and Clarke, 1999). During Pleistocene times (Fig. 2.2), these lakes were present throughout much of the glacial period and occupied a narrow band around the southern margin of the ice-sheet. At ice retreat times, pro-glacial lakes filled even larger areas (Marshall and Clarke, 1999). Many factors control the formation of ice-marginal lakes: among the most important are location of the ice-sheet margin, elevation of outlets, differential isostatic rebound and the topography of the region adjacent to the ice-sheet (Teller, 2003).

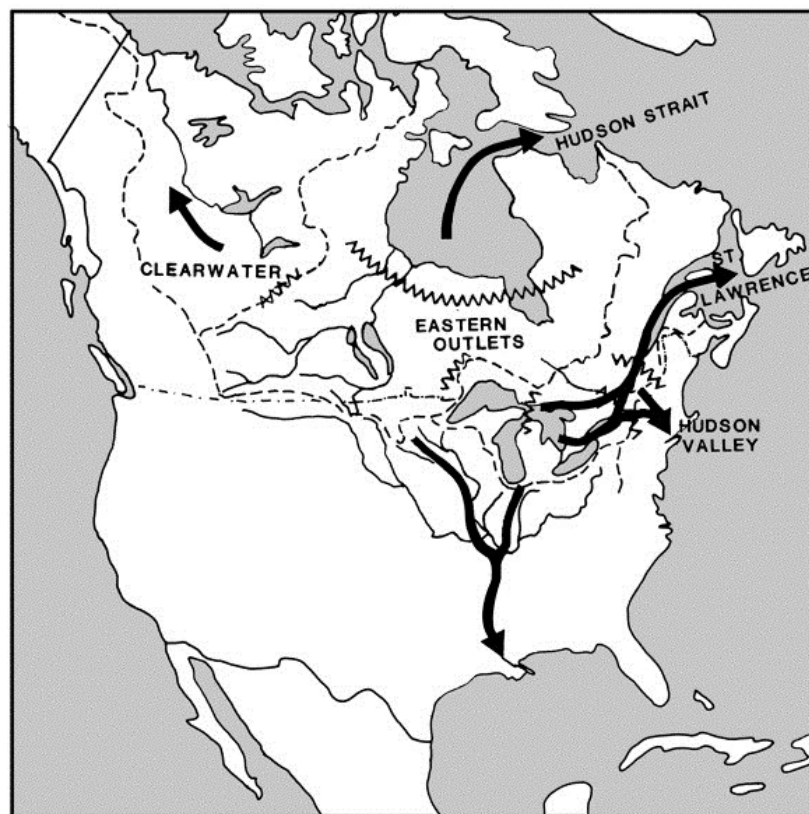


Figure 2.2: Palaeo-hydrological thresholds (thin lines) associated with the presence of Laurentide Ice Sheet and related pro-glacial lakes (inland grey areas) in North America during Late Quaternary times. Arrows indicate major overflow routes from pro-glacial lakes; dashed lines indicate modern hydrological divides (from Teller, 2003).

Derived from ice melt-water and from local input of the regional drainage river network, diverted by the presence of the ice body (Teller and Leverington, 2004). As pro-glacial lakes usually do not have efficient outlets, the level of the lake gradually rises until a sudden failure in the lake dam causes catastrophic over-spilling of the water (Walder and Costa, 1996). This typically occurs during ice-sheet melting, as the input of water is greater at these times, resulting in high lake-levels and sudden collapse or breaching of the lake dam. The collapse of the dam releases the water to flow from the lake to the ocean through the river drainage network, dramatically changing the landscape of entire regions, as enormous volumes of water flows through the river network in a very short time (Baker, 2002a). As normal fluvial channels cannot accommodate the volume of water involved in megaflood processes, the river network undergoes strong erosion during the flood. The effect of megafloods on the landscape is carved into areas of hundreds of square kilometres, preserved for hundreds to thousands of years, as common erosional processes do not have the necessary erosional power to modify the bedforms that are carved into megaflood terrains (Baker, 2009a; Baker, 2009b; Carling et al., 2009a).

A second type of glacial lake which can cause megafloods is supraglacial lakes: these lakes are known to be the largest ice-sheet related water bodies. They are located on the ice surface (differently from sub-glacial lakes, located under the ice surface) and their dimensions are extensive, occupying thousands of square kilometres. Because of their size, supraglacial lakes are ephemeral features that form when the ice sheet is in a state of transition. Diffusive relaxation of the ice surface topography is considered the process that produces topographic basins on the ice surface (Marshall and Clarke, 1999) where these lakes form. Because this process takes hundreds of years, while supraglacial lakes only exist for a short period of time, the existence and stability of huge supraglacial lakes is widely debated in literature. If on the one hand they have been proposed as the source of huge floods that triggered climatic changes such as the Heinrich event (e.g. Shaw et al., 1996), the majority of authors discard this possibility as large supraglacial lakes are unstable and would require too long time to be large enough to have an influence (e.g. Marshall and Clarke, 1999; Clarke, 2005).

Other types of lakes that can generate megafloods are represented by lakes ponded by structural dams, as for the case of the Bonneville flood (Jarrett and Malde, 1987), caldera lakes, landslides or Jokulhlaups, which are typical of alpine settings. Jokulhlaup, meaning

glacier burst, is the Icelandic term for a flood caused by the sudden and catastrophic release of water impounded within or behind glacier ice (Thorarinsson, 1953).

Megafoods events are characterised by damming of either trunks or tributary streams by glaciers (Fig. 1), advancing into valley systems and generating lakes that can catastrophically drain when the dam is breached. In particular, there are two main possible geometries for the lake formation, depending on the timing of the valley occupation by a tributary or by a glacier valley (Walder and Costa, 1996). The mechanism that leads to the dam breaching also depends on the geometry of the lake. When the glacier valley is dammed by a tributary, the dam failure mechanism happens by overtopping of the dam. This causes a feedback effect of ice melting that increases the size of the breach, up to the final failure and drainage of the lake by subaerial channels. Conversely, when a glacier valley dams up a tributary, the dam failure usually occurs sub-glacially, through the growth of a subglacial conduit (Clarke et al., 1984) or tunnel, where the melt water flows up to the terminus of the glacier (Nye, 1976).

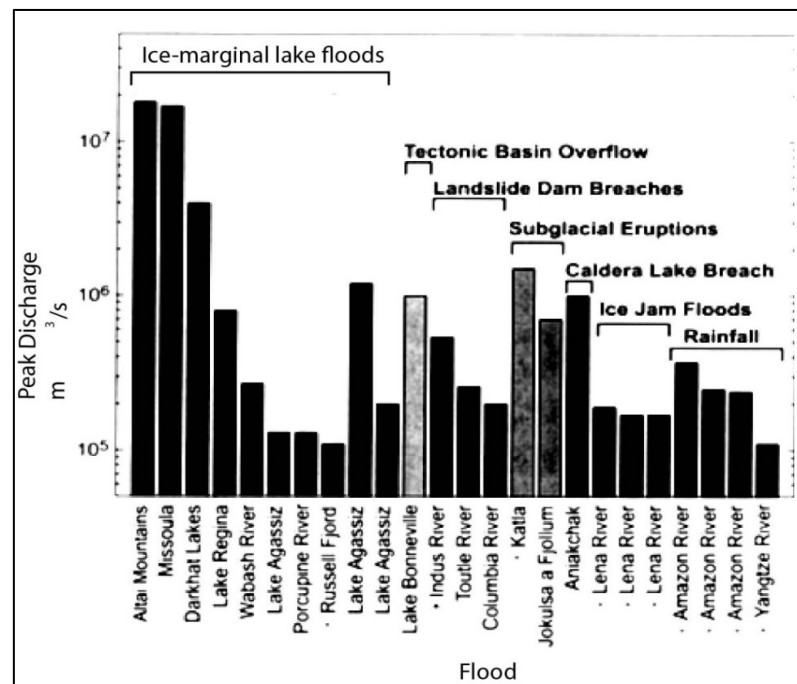


Figure 2.3: Graph showing the calculated peak discharge of the main megafood events known on Earth, grouped depending on the nature of the water drainage mechanism. The events with highest discharge are associated to floods originated from pro-glacial lakes (from Herget, 2005).

In summary, water bodies can be classified according to their spatial relation to the ice-sheet, as shown in Tab. 2.1 and Fig. 2.1. Depending on the position of the lake in respect to the ice-

sheet (e.g. pro-glacial lakes are located at the ice-sheet margin, sub-glacial lakes under its surface, supra-glacial lakes on the ice surface), different mechanisms can drainage the lake. In the case of the Dover Strait, possible mechanisms are discussed in Section 7.5 following the discussion and interpretation of the data.

Fig. 2.3 shows the peak discharge of some of the major floods identified on Earth in relation to the lake-drainage mechanism, compared to the discharge of Amazon and Yangtze rivers. As shown in the figure, events with higher discharge are usually associated to ice-sheet dammed lakes (Fig. 2.3), as these types of lake tend to accommodate bigger volumes of melt water originated from the melting ice-sheet (Herget, 2005).

2.2.2 A scale problem and global implications

The main characteristic of megaflood events is their high magnitude (with discharges of million of cubic meter per second), and exceptionally high-energy erosional power, and low frequency of occurrence, as they are related to processes of ice-sheet dewatering that only happen every tens of thousands of years (Baker, 2009b). The concept of megaflood was not accepted by the scientific community until 1940, following two decades of controversy on the origin of the Channeled Scabland of Washington, US, the first recognized megaflood terrain.. The “big flood” concept, introduced by H. J. Bretz (1923) in his first paper on the Channeled Scablands, was refused for decades by the scientific community that rejected the concept of cataclysmic events in geology.

As megafloods are characterised by low frequency, their direct observation have not been possible. The study of megaflood terrains and palaeo-indicators present on ancient flood landscapes is crucial for the understanding and reconstruction of the processes and mechanisms. The connection between ice-sheet and ocean are not well understood. There is evidence of a close connection between high discharge of ice melt-water, ice rafting and rapid sea-level rise. These combined effects could alter ocean equilibrium (Clark et al., 2001). In particular, in the last 30 kyr, a number of events of discharge of large volumes of water into the oceans have been related to periods of climate cooling. Among these are Heinrich events (that occurred at 17, 24 and 31 kyr ago) and Younger Dryas (12.8 kyr ago), interpreted to be related to rapid discharge of huge volume of fresh-water fluxes, coming from ice (Barber et al., 1999; Clark et al., 2001; Teller and Leverington, 2004). Heinrich

events, first described by marine geologist Hartmut Heinrich, occurred during the last glacial period. These events are known as movements of icebergs breaking off the ice sheet during its disintegration and flowing from the ice sheet into the North Atlantic Ocean (e.g. Heinrich, 1988; Grousset et al., 2000). As a consequence, rock mass eroded by the glaciers and transported by the icebergs was deposited onto the sea floor as ice rafted debris. Six distinct events can therefore be identified in cores, labelled H1 to H6 (e.g. Bond and Lotti, 1995; Hemming, 2004). In addition to debris, melting of icebergs caused huge volumes of fresh water to be added to the North Atlantic. Heinrich events, may have altered the thermohaline circulation patterns of the ocean, and are interpreted as indications of global climate fluctuations. During these events, huge volumes of fresh water flow into the ocean. For the biggest event the fresh water flux has been estimated to 0.05 Sv (1 Sverdrup corresponds to $1 \times 10^6 \text{ m}^3/\text{s}$) with a duration of 100-300 years (Roche et al., 2004), equivalent to a fresh water volume of about 2 million km^3 . The global extent of these records illustrates the dramatic impact of Heinrich events.

Megafoods can discharge as much as of several Sverdrups of fresh water into the ocean. The water enters the ocean as low density plumes (Kourafalou et al., 1996; Zuffa et al., 2000; Benito and O'Connor, 2003) or high density hyperpicnal flows (Mulder et al., 2003), affecting the salinity, the temperature and current equilibrium of oceans. This is due to a contrast in the temperature, and therefore density of fresh water. As the fresh water has lower temperature than the ocean, the plume sinks to the bottom layer of the water column, disrupting the natural circulation and upwelling of warmer water. As the heat in the ocean is distributed by currents such as the Gulf Stream flow through this upwelling mechanism, which moves volumes of water of up to 100 Sv (Baker, 2009b), acting as a global-scale heat conveyor system, a disruption in the ocean current equilibrium can cause global-scale alteration to Earth climate. This global-scale, long-term effect of megaflood flows represents one of the main reasons why the study of these events is crucial in the reconstruction of past climate changes and in the understanding and prediction of present and future Earth climate evolution patterns.

2.2.3 Review of channel-scale megaflood landforms

Megaflood terrains display a diversity of channel-scale erosional and depositional features that are characteristic of high energy flows (Carling et al., 2009b). Although most bedrock

channels are thought to be eroded over millions of years (e.g. Karlstrom et al., 2008) by steady flows with occurrence times of few years, kilometre-scale valleys that are observed on Earth and Mars, appear to have been formed rapidly in a single event (e.g. Lamb and Fonstad, 2010) or in a series of cataclysmic flood events (e.g. Baker, 1978a; Herget, 2005; Lamb et al., 2008). Reconstructing the flood discharge and duration of these flows is difficult, mainly because of the lack of tested morphologic metrics and models of bedrock erosion (Carling et al., 2009b). This is in part because floods capable of rapidly carving bedrock canyons occur infrequently; only a small number of examples exist on Earth and most have been inferred from geologic evidence rather than observed directly. The interpretation of the formation of such features is therefore a key point in the reconstruction of the palaeo-hydraulic and magnitude of the flows that carved them, although the formative condition of some of these features is not clearly understood yet (Carling et al., 2009b). The Channeled Scabland megaflood terrain, in the north-western, US, displays all the typical flood bedforms and morphologies that were firstly interpreted (Bretz, 1923) as generated by catastrophic flood. A review of the characteristic flood erosional and depositional features is given in this chapter using examples from the Channeled Scabland terrain.

In the case of the English Channel, despite having been carved in subaerial conditions, the megaflood terrain is nowadays submerged. For this reason, most of the depositional landforms described in the following paragraphs and typically preserved in the sub-aerial setting are not present as they have probably been altered or eroded by later processes (see Chapter 6).

Erosional features

Megafloods can erode bedrock and loose granular material. Analysis of erosional bedforms allows for the estimation of the magnitude of the erosional processes that carved the observed features (Carling et al., 2009a, 2009b). Some of the flood erosional features have dimensions that are proportionally related to flow depth, like scours (Curl, 1966), or to flow field width, like bedrock plunge-pools (Keller and Melhorn, 1978).

Concave features

Erosional features are usually distinguished between concave and convex features. The first class is negative features carved into bedrock: these can often be used as palaeo-hydraulic indicators, in order to reconstruct the direction of the flood flow as they typically develop parallel to the direction of the flow.

The main concave erosional features in megaflood terrains are represented by linear bedrock channels, showing cross section morphology of inner incised channels with lateral flat benches that extend for hundreds or thousands of meters (Carling et al., 2009a). Carved into the channels are knickpoints, usually located at the head of inner channels: knickpoint retreat is in fact considered as the mechanism that initiates the incision of inner channels (Baker, 1973a; Wohl et al., 1994; Howard et al., 1992). The geometry of megaflood channels is variable, as it ranges from relatively straight geometry to a complex network of channels (Fig. 2.4) usually separated by teardrop-shaped bedrock remnants (Baker and Nummedal, 1978). The main characteristic of flood channels is their kilometre scale dimension and depth that can reach hundreds of meters. These features often laterally cut smaller channels, called hanging valleys, which are found at higher elevation in respect to the main channel floor, as they are pre-existing channels that were eroded during floods.

Potholes and plunge pools are also concave erosional features that are incised into the channel floor and that consist of near-circular depressions carved into bedrock.

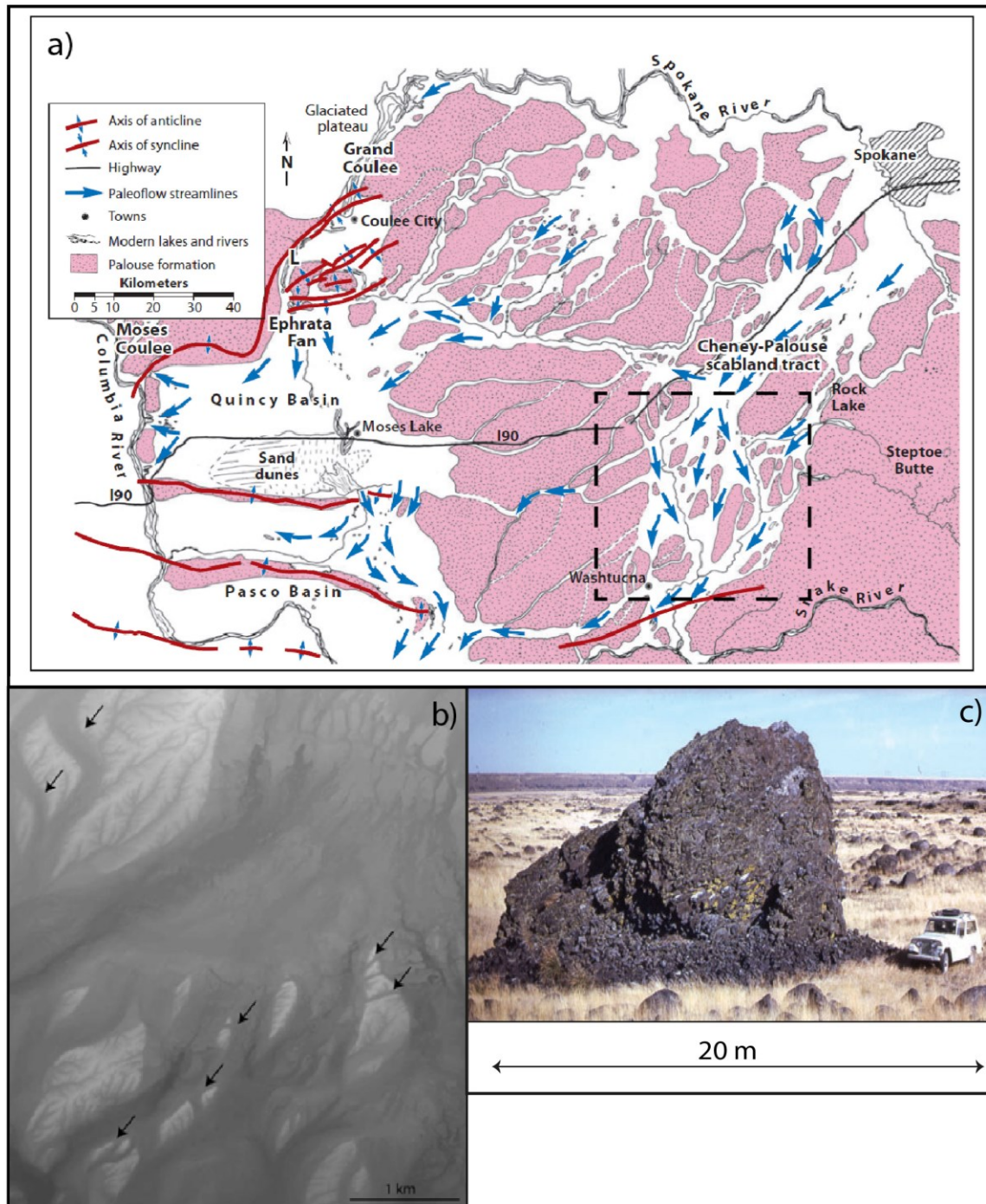


Figure 2.4: a) Example of megaflow channel and bedrock remnants of the Scablands terrain (from Baker, 2009a), showing braided to straight geometry of the channels separated by streamlined remnants. These are particularly well developed in the Cheney-Palouse area (dashed box) from which an example is shown in box b. This shows divide crossings (pointed by the black arrows) separating a cluster of streamlined islands in the Channeled Scabland of Washington. Elevation data from Seamless SRTM (USGS, 2006. <http://srtm.csi.cgiar.org>). Light grey represents elevated area while darker grey represents lower elevation. Box c shows a large boulder deposited by the flood in the Euphrata fan area of the Channeled Scablands. The major axis of the boulder measures ~18 m. From Baker, 2009a.

Potholes are carved by vortices and measure up to 250 m in diameter in the Scablands, as for the case of the Pothole Coulee (Bretz, 1923; Baker and Nummedal, 1978). Plunge pools also display circular geometry and are created by plunging water plucking these features into bedrock and forming near-circular depressions, located at the base of a hydraulic jump (Lamb et al., 2007). In the Channeled Scabland, deep plunge pools are located at the base of the Dry Falls cataract, measuring up to 1 km in width and 35 m deep (Bretz, 1923; Baker, 1978a).

Among concave erosional features that can be used as palaeo-flow indicators are also flutes, furrows and big scale lineations such as grooves. In the Channeled Scabland, sets of parallel megalineations are carved into bedrock just upstream of the Dry Fall cataracts. The lineations are up to 300 m wide and 2.5 km long: their origin has been explained by erosion by longitudinal vortices in the flow (Baker and Milton, 1974; Baker, 1978b). Interestingly, these features show striking resemblance to glacial megalineations associated to sub-glacial melt-water (e.g. Bradwell et al., 2005; Shaw et al., 2008; Graham et al., 2009). A detailed description and discussion of longitudinal megalineations is given in Chapter 6 of this work.

Convex features

The most common erosional convex features in megaflood terrains are represented by residual landforms (Carling et al., 2009b). Among these features, the most typical bedforms are bedrock or sediment remnant islands (Fig. 2.4) characterised by “airfoil” type geometry (Baker, 1978a; Patton and Baker, 1978a, 1978b; Komar, 1983). These are termed residual as they consist of remnants of a sedimentary or bedrock body that filled the valley before the flood event (Carling et al., 2009a). These streamlined convex features show a strong similarity to the lemniscate forms in glacial drumlins (Baker, 1973b), revealing a relationship between length, width and area. Baker (1979) and Komar (1983) demonstrated that the teardrop-shaped, streamlined geometry of these islands reduces resistance to the flow. In the Channeled Scabland terrain these features are present as clusters of hundreds of loess streamlined, kilometre-scale islands that rise up to 50 m above the surrounding scabland terrain (Patton and Baker, 1978b). The islands are separated by braided channels within the overall scabland complex. The presence of bars located in the area where the islands are mapped, useful to diagnose the flood pathway, suggests that they were deposited in an area that was protected from reworking by following flood events

after their deposition (Baker, 1978a). The islands are often severely modified or destroyed by further flood flows and display typical cross-over small-scale channels called divide crossings (Fig. 2.4) that are used as high water mark indicators (Baker, 1978b). A detailed description and analysis of streamlined islands is given in Chapter 6 of this work.

Together with residual landforms, cataracts are also classified as convex erosional features. They are near vertical drops carved at the head of the kilometre-scale megaflood channels (Bretz, 1923; Baker and Nummedal, 1978; Lamb et al., 2008; Baker, 2009b). Retreat cataracts typically form by sub-aerial erosion and knick-point headward migration during flow retreat (Wohl, 1998). In the Scablands, cataracts are up to 120 m above the underlying channel floor, as in the case of the Dry Falls (Baker and Nummedal, 1978), forming a horseshoe-shaped vertical wall whose head is up to 5.5 km wide.

Depositional features

Identification of alluvial bedforms requires the interpretation of the nature of the flow and of the depositional conditions under which the landform has formed (Carling et al., 2009a). In particular, factors like dimension of the landform and geometry of the bedform association can be used to interpret changes in the erosional/depositional character of the flow that formed them (Carling et al., 2009a). The kilometre-scale geometry of bedform and unusual amplitude may indicate high water-depth, as their crests develop close to maximum water depth (Costa, 1984). Accepting this assumption, these features can be used as high-water indicators for palaeo-hydraulic reconstructions (O'Connor and Baker, 1992; Burr, 2003; Herget, 2005). Finally, the stratigraphy and sedimentology of these landforms give indication of their formation mechanisms (Carling et al., 2009a).

Gravel bars and giant current ripples are the most common depositional features found in megaflood terrains. Thick, up to 10 metre gravel bars are generated by deep flows, showing internal cross stratification and foreset bedding (Bretz et al., 1956; Baker, 1973a). In particular, two main typologies of bars were recognised in the Channeled Scabland terrain: longitudinal bars, elongated in the direction of the flow, and eddy bars, located at tributary valley mouths (Baker, 1973a). Giant current ripples (GCR) are usually present in groups of dozens of bedforms; unpublished works by Bretz indicate the presence of more than 100 GCR in the Lake Missoula flood area. Baker (1973a) defines the typical distance between GCR in the Channeled Scabland as varying from 20 to 200 m, displaying a height that ranges from 1 to 15 m and occurring in groups of 20 or more features. The widespread preservation of these structures indicates the presence of a lower flow regime during megaflood flows (Baker and Nummedal, 1978). Pardee (1942) related a progressive change in their heights to a reduction in the flow speed linking landform geometry to the flow pathway.

Finally, boulders can also be considered as megaflood sedimentary features (Baker, 1973a). Large boulders, with up to 18 m in diameter for the Channeled Scablands boulders (Baker, 1978, 2009a), clearly indicates very high transport capability of the flood flow (Fig. 2.4c)

2.2.4 Megaflood terrains on Earth

A brief description of the well studied megaflood terrains is given in this section. These will be used as analogues for the analysis of the English Channel megaflood bedforms, comparing the observed morphology, distribution and association of flood features to the main examples of flood-generated, erosional features described in literature.

Channeled Scablands terrain

The Channeled Scabland region is a portion of the basaltic Columbia Plateau and Basin area that occupies more than 100,000 square miles between the States of Washington, Oregon and Idaho. In this area, a pro-glacial lake named Missoula (e.g. Bretz, 1969; Baker, 1973b; Smith, 2006) formed during the Pleistocene (Pardee, 1942), located at the margin of the Cordilleran ice-sheet. Reconstructions indicate that the lake was up to 150 m deep at the lake dam site, and had a volume of 2700 km³ (Clarke et al., 1984; O'Connor and Baker, 1992). The multiple flood events that carved the Scabland terrain are associated to the periodic overspilling of Lake Missoula into two smaller adjacent lakes, named Spokane and Columbia. Floodwaters surged over the south bank of the lakes into three major spillways, crossing the Columbia Basin towards the ocean and carving the typical scabland butte-and-basin topography into the

bedrock and loess (Baker, 1973b). This particular morphology of the landscape has now become one of the main distinctive characteristics of megaflood terrains.

The estimated number of events that occurred during the megaflood events (late Pleistocene) ranges from 40 to 100 over a period of 3000 years between 17.5 and 14.5 kyr before present (Waitt 1985; Atwater 1986). These dates were constrained dating lake varve layers at the bottom of the sequence and an ash layer originated from Mt. St. Helens eruptions, located at the top. This was dated to 13 kyr (Moody 1987; Bunker 1982; McDonald and Busacca, 1992). However, evidence of flooding during earlier Pleistocene was also found from sedimentological analyses (Bretz et al., 1956; Patton and Baker, 1978b; McDonald and Busacca, 1992). Varve layers present in the lake fill give evidence of multiple, periodical events with interval of 40 to 60 years between the floods (Brunner et al., 1999).

Many palaeo-hydraulic calculations have been performed in order to estimate Lake Missoula megaflood peak discharge: different techniques have been used to estimate a discharge of the order of magnitude of $10^6 \text{ m}^3/\text{sec}$. Bretz (1925) and Pardee (1942) used Chezy uniform flow equation for the palaeo-hydraulic reconstructions, while Baker (1973a) used Manning's slope-area formula, obtaining up to 30% higher discharge (Baker and Bunker, 1985). More recent palaeo-hydraulic reconstructions (O'Connor and Baker, 1992) indicate discharges up to 17 Sv for the biggest of the flows, with average values of 13 Sv and estimated duration of the floods of 5 to 7 days.

The Bonneville flood

The Bonneville flood, in the Great Basin region, US, is the best constrained megaflood event in terms of palaeo-hydraulic reconstruction. It consisted of a single flood event, apparently caused by the breaching of a structural dam that bounded Lake Bonneville. For this reason, the Bonneville flood has been used for the calibration of dating and hydraulic techniques (e.g. Jarret and Malde, 1987; Oviatt et al., 1994; Cerling and Craig, 1994; Wohl, 1998). The flood was generated by water overflowing from Lake Bonneville, the largest of the late Pleistocene lakes in the North American Great Basin (Benson et al., 1990; Currey, 1990). The lake rose in level between 15 and 14.5 kyr BP, reaching a maximum at which the water drained over the basin rim, eroding the alluvium barrier that bounded the lake (Malde, 1968). As a consequence of the flood, the lake level dropped of up to 100 m and about 4750 km^3 of water was released (Richards, 1995). The flood water escaped the basin through a passage called Red Rock Pass, successively joining the Snake River and

flowing for more than 800 km in 12 days. The peak discharge for this flood has been calculated to be 0.6 Sv, with an estimated duration of three months (O'Connor, 1993).

Lake Agassiz megaflood

Lake Agassiz is a pro-glacial lake that was located at the margin of the Laurentide ice-sheet in North America. The Lake Agassiz flood is of particular interests in the analysis of megaflood impact on climate change: the fresh water input generated by Lake Agassiz flood is thought to have been the cause of a cold event that affected Earth climate, known as the 8.2 kyr climate event (Alley et al., 1997). The flood, dated to 8.4 kyr (Barber et al., 1999), is associated to the catastrophic drainage of pro-glacial lakes Agassiz and Ojibway (Leverington et al., 2002; Teller et al., 2002), located at the margin of Laurentide ice-sheet during the ice body retreat that followed Last Glacial Maximum. At that time, the ice-sheet covered approximately 10 % of North America surface, with an ice mass of $15.9\text{--}37 \times 10^6 \text{ km}^3$ (Licciardi et al., 1998). Given the enormous dimension of the ice body, the load exerted by the ice-sheet created wide depressed areas into which pro-glacial lakes formed. These lakes were located at the ice-sheet southern margin, were filled by huge volumes of melt-water originated by ice-sheet ablation and melting (Teller and Clayton, 1983). The size of the lakes at that time was dependent on the location of their outlet in respect to the ice-sheet, with water catastrophically overspilling above the lake divide and eroding new outlets. This caused either further drainage of the lake or water-lake lowering, depending on the outlet configuration and on the volume of the melt-water input into the lake.

Sediments deposited by the flood were found in sediment cores located at the Hudson and Labrador Strait, indicating that the sediments were transported for more than 700 km from their source (Barber et al., 1999). Hydraulic modelling (Clarke et al., 2004) suggests that the peak discharge of the flood was up to 5 Sv, with duration of less than one year. The water coming from the lakes entered the ocean, forming a fresh water cap may have inhibited North Atlantic deep water formation and circulation, originating the cold spike registered in Earth climate at 8.2 kyr (Teller and Leverington, 2004).

Big Lost River megaflood

The Big Lost River flood took place in South Idaho 16.9 kyr, according to cosmogenic measurement on flood boulders and scour features (Cerling et al., 1994). Although this flood is smaller than other megafloods that occurred around the same time in North America, it

represents further evidence for widespread warming and ice-sheet melting at the end of Pleistocene. The flood path ran along the Big Lost River into the Snake River Plain, depositing boulder bars and loess hills with streamlined geometry and eroding scour features as well as plunge pools and cataracts (Cerling et al., 1994). The estimated peak discharge of the flood is 0.06 Sv (Rathbourn, 1993).

Altai Mountains megaflood

A series of catastrophic floods occurred in the Altai Mountain region, Siberia, between 46 and 13 kyr BP (Rudoy, 1988; Rudoy et al., 1998), generating one of the largest megaflood terrains known on Earth. It has been subject of numerous sedimentological studies (e.g. Rudoy and Baker, 1993; Rudoy, 2002; Carling et al., 2002; Herget, 2005). Several floods occurred, with one major event, named Chuja-Kuray flood, that reached a peak discharge of 18 Sv (Baker et al., 1993). This event represents the biggest known single event ever occurred on Earth

(Rudoy and Baker, 1993), although more recent calculations suggest a peak discharge of 9 Sv (Carling et al., 2010). The flood water was catastrophically released from a series of ice-dammed lakes with capacity up to 600 km³: the biggest flood events occurred through the collapse of an impounding ice barrier bounding the lake that reached its highest volume and drained over the dam (Baker et al., 1993). The flood water flowed towards the West Siberia plains, transporting sand and gravel rich sediments that were entrained along the course of the flood path. Sediments were deposited over a distance of more than 70 km from the source (Herget, 2005): analyses of the flood fill materials indicate a Marine Isotopic Stage (MIS) 2 (see Fig. 2.11b for reference) age of the series of main floods, although analyses performed on lake sediments indicate that Chuja-Kuray lake already existed during MIS 4 (Carling et al., 2002; Herget, 2005; Carling et al., 2009c).

2.2.5 Introduction to megaflood palaeo-hydrology

Megafloods are generally characterised by peak discharge in excess of one million cubic metre per second, from which the prefix “mega” is derived (Baker, 2002a). The unit used for the measurement of high magnitude peak discharge events also corresponds to one million cubic meter per second, named Sverdrup (Sv), and traditionally used for the measurement of oceanic currents that involve the same volumes of water as megafloods and which are the other major short-term water movement mechanisms on Earth surface (Baker, 2001). The

analysis of palaeo-flood hydrology supports the estimation of flood magnitude, calculation of peak discharge and estimated duration of the flood.

The exploration and analysis of the land-forming capacity of ancient floods can be performed using a variety of techniques (Carling et al., 2010). One-dimensional (1D) steady-flow techniques have been widely used in the last thirty years in order to reconstruct the magnitude of megafloods and the maximum reached water level (O' Connor and Webb, 1988). 1D models mainly require the geometry of the channel cross-section and evidence of maximum flooding levels (Pardee, 1942; Baker, 1973b; Dorava and Meyer, 1994; Tomasson, 1996), assuming instantaneous peak discharge at all sections of the floodway. In channels with stable cross-sectional geometry during and after the flood, step-backwater models (Partridge and Baker, 1987; Enzel et al., 1994; Wohl et al., 1994) can be used to reconstruct water surface-profiles (Wohl, 1998). These profiles are then best-fitted with profiles derived from landscape palaeo-indicators: the most employed model for these calculations is HEC-2 model (HEC-2, 1990). HEC-2 assumes steady-flow condition in time, gradually varied in space (Wohl, 1998). Thanks to the advances in hydrodynamic modelling, in the last few years, 1D models have been replaced by more complex 2D and 3D unsteady-flow models. These can integrate multiple factors, such as velocity variations along the cross-section (Carling et al., 2010), bed erosion and deposition along the floodway (e.g. Cao and Carling, 2002; Cao et al., 2004).

Examples of terrestrial catastrophic palaeo-flood discharge calculations include analyses on the major megaflood terrains on Earth described in Section 2.2 of this Chapter, such as the Channeled Scablands (e.g. Clarke et al., 1984; O' Connor and Baker, 1992), Lake Bonneville megaflood (e.g. O'Connor, 1993), Altai Mountains (e.g. Baker et al., 1993; Herget, 2005; Carling et al., 2010) and Lake Agassiz (e.g. Clarke et al., 2004).

Peak discharge calculations

Regardless of the complexity of the model, palaeo-hydraulic analyses can only be performed when palaeo-flow indicators such as high-water marks (Baker, 1978b) and channel cross-section geometry are available (e.g. Benito and Thorndycraft, 2005; Baker, 2008) as peak discharge is calculated using the palaeo-channel cross-sectional area (A) and mean velocity (U) of the flow, through the equation:

$$Q = U * A$$

Cross sectional area is calculated assuming bankfull surface of the channel (e.g. Glenn, 1911; Longwell et al., 1948; Leopold et al., 1964; Tuttle, 1975; Williams, 1978), represented by channel bankfull depth and width (Fig. 2.5). Bankfull depth represents the average vertical distance between the channel bed and the estimated water surface elevation required to completely fill the channel. Bankfull width is the measurement of the lateral extent of the water surface elevation perpendicular to the channel at bankfull depth. Finally, mean velocity represents the mean velocity within a vertical profile or the time-averaged velocity at a fixed point, and is usually calculated using three different equations, described in the following section.

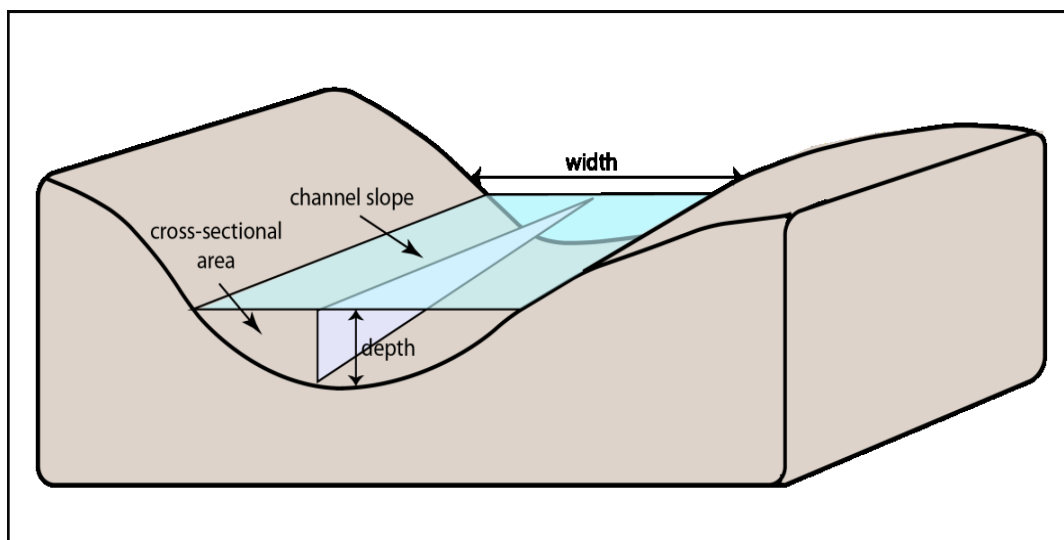


Figure 2.5: Cartoon of channel cross section parameters used for the peak discharge calculation. Slope is indicated by the triangle showing the dip of the channel-floor, assuming dip of the channel increasing towards the reader.

Manning's equation

The most common formula to calculate flow mean velocity is Manning's equation, represented by the formula:

$$U = a (R^{2/3} * S^{1/2}) / n$$

with **a** is dimensionless constant that corresponds to 1.0 when SI units are used, **S** is the slope of the channel in degrees, **n** is Manning's roughness coefficient, sensitive to the channel geometry, flow rate, flow depth and channel roughness, (Chow, 1959; Barnes, 1978; Bathurst, 1993). As these parameters and their relationship with **n** cannot be easily quantified, empirical and theoretical formulae were calculated and

tabulated for the estimation of **n** in a wide range of different channels (Cowan, 1956; Chow, 1959). When possible, **n** can be calibrated using observations of flow velocity and depth (Kidson et al., 2006). **n** generally ranges from 0.01 for smooth laboratory channels to 0.1 in rough step-pool streams, with natural channels usually having values of 0.02 to 0.08 (Wohl, 1998). Finally, **A** is the cross sectional area of the channel and **R** is the hydraulic radius, given by the equation:

$$R = A/P = (W*H)/(W+2*H)$$

where **P** is the wetted perimeter and **W** and **H** are respectively the channel bankfull width and depth. For very wide channels where $W \gg H$, as in the case of megaflood channels, **H** is taken as a proxy of hydraulic radius **R**.

Chezy formula

Chezy formula is another formula for the calculation of mean velocity. It was empirically calculated by Chezy following his work on pipe flows and canals in Paris. The formula is the following:

$$U = C*(R*S)^{1/2}$$

where **S** is the slope of the channel in degrees, **R** is the hydraulic radius, **C** is known as Chezy coefficient as is derived by the equation:

$$C = R/n$$

where **n** is Manning's roughness coefficient. **C** is dependent on surface roughness, Reynolds number, and hydraulic depth. Tabulated values of **C** and the related **n** are given by Chow (1959).

Darcy-Weisbach equation

The third formula is represented by Darcy-Weisbach equation, given by the formula:

$$U = (1/f) * (gHS)^{1/2}$$

with **S** representing the slope of the channel in degrees, **H** is channel depth, **g** is gravitational force and **f** is Darcy-Weisbach friction factor.

This parameter accounts for the loss of energy in natural channels, due to the friction between the flow and the wet perimeter. The value of f is estimated for natural channels and can be theoretically calculated with the formula:

$$f = (8 \cdot \tau_0) / \rho \cdot u^2$$

where ρ is fluid density, u is flow velocity and τ_0 the basal shear stress inside the channel. However, the formula is not suitable for natural channel application and, as for the case of n , the value of f is usually empirically measured taking measurement of channel flow velocity and depth from the channel to which the formula is applied (Bathurst, 1978; Heritage et al., 2004). Another way of deriving f is through Manning's coefficient (Eq. 7 of Kleinhans, 2005), suggesting that the value of f may range between 0.3 and 1.1 for braided gravel flood rivers, although it does not remain constant but changes with depth of the flow, declining as the flow thickens.

Limitations

One of the main limitations of palaeo-hydraulic reconstruction of megafloods is the fact that calculations are based on hydraulic formulae which were derived for use in streams whose discharges are three or four orders of magnitude smaller than megaflood flows (Baker and Nummedal, 1978). Channel geometry and roughness parameters are also difficult to constrain for megaflood channels. The absence of direct and detailed observations prior and during the flood means that specific values of n and f are difficult to estimate (O'Connor and Baker, 1992; Benito and O'Connor, 2003; Burr, 2003; Wilson et al., 2004; Kleinhans, 2005; Clayton and Knox, 2008; McIntyre et al., 2012), as well as the precise measurement of channel geometry.

Secondly, palaeo-hydraulic calculations are based on bankfull channel geometry, implying that no subsequent flows or long recession of the flood hydrography might have deepened the channels, enlarging their capacity beyond what is implied by high water surface (Baker, 1978a). In order to overcome this problem, high water marks are used to reconstruct the maximum flood surface related to the megaflood. The exponential relationship between flow and channel depth (Bathurst, 1993) implies that small errors in the measurement of water level lead to larger errors in the calculation of peak discharge. In the case of megaflood palaeo-channels, being the direct measurement of channel geometry not possible, it is necessary to use palaeo-indicators such as benches, crossing divides and inner channels

(Baker, 1978b), in order to define the channel geometry, flow depth and high water surfaces (Baker, 1973b), and therefore distinct erosional events, that carved the channel.

Resolution of the data used for the measurement of the channel parameters, and in particular, of the extracted channel profile is therefore a primary factor that has to be taken into account when analysing the channel geometry (McIntyre et al., 2012). Poor resolution of data prevents from distinguishing multiple surfaces present in the channel, generated by multiple flows, leading to the calculation of higher peak discharge. This problem particularly affects palaeo-hydraulic calculations for limited-resolution Martian data (e.g. Burr, 2003) imaging extremely wide and deep megaflood channels where the assumption of one single event flood for the formation of the channels can lead to overestimated calculated peak discharge.

Poor resolution of data also affect measurement of the channel slope, that can also be affected by later processes that alter the original geometry (and slope) of the channels. Limited resolution of the data can lead to errors in the measurement of the slope that can significantly vary from along the channel profile but not be resolved by data with low resolution. In the case of the English Channel, measurement of slope was strongly affected by glacial isostasy movements that affected the Lobourgh Channel area, as explained in more detail in Section 6.5.

2.3 Geology of the English Channel

A description of the main aspects that characterise the geology of the English Channel is given in this section, with particular interest on the geomorphology, tectonic evolution, stratigraphy and brief description of the tidal mechanisms present across the Channel and strongly affecting surficial sediment distribution.

2.3.1 Geomorphology of the Channel

The English Channel is an ENE–WSW trending seaway that connects the Southern North Sea to the Celtic sea (Fig. 2.6), measuring ~600 km in length and ranges between 30 to 200 km in width, respectively at the Dover Strait and at its south-western reaches, where the Channel is much wider. Depth varies between 30 and ~200 m at the Channel's deepest point in the south-western sector. Seafloor morphology is dominated by a network of channels, known as

Channel River system (Hull, 1912) (Fig. 2.7) carved into its seabed in both sectors. The Channel River forms a braided network of channels that flow with a NE-SW direction for more than 200 km, from the Dover Strait to the Hurd Deep. The transported sediments are discharged into the Celtic Sea fans, located in the Atlantic Ocean at the mouth of the English Channel, on the west of the Hurd Deep (Fig. 2.7).

The British sector of the Channel is dominated by the presence of the Northern Palaeovalley (Fig. 2.7), a network of channels with ENE-WSW direction that measures more than 150 km of length and that displays variable morphology throughout the Channel. Depth ranges between 60 and 80 m, locally reaching 170 m of depth in correspondence of a deep carved into the seafloor known as Hurd Deep (Hamilton and Smith, 1972). The Northern Palaeovalley is joined by the Lobourg Channel from the north-east (at the Dover Strait) and from the north by the palaeo-Arun and palaeo-Solent rivers, in the area of the Isle of Wight.

The Hurd Deep (Fig. 2.7) is the termination of the Northern Palaeovalley and the deepest point in the English Channel. The deep is located at the palaeovalley south-western reaches and shows a depth of up to 170 m, length of 150 km and width of 2 to 5 km (Stamp, 1927; Dangeard, 1929; Kellaway et al., 1975; Smith, 1985). Although Smith (1985) proposed the Northern palaeovalley and the connected Hurd Deep to be the result of a catastrophic flood, the Hurd Deep was formed by different processes with respect to the Palaeovalley. Many interpretations have been proposed, ranging from karstic processes (Boillot, 1963) to tidal scouring (Hamilton and Smith, 1972) to a tectonic origin, proposed by Lericolais et al. (1996).

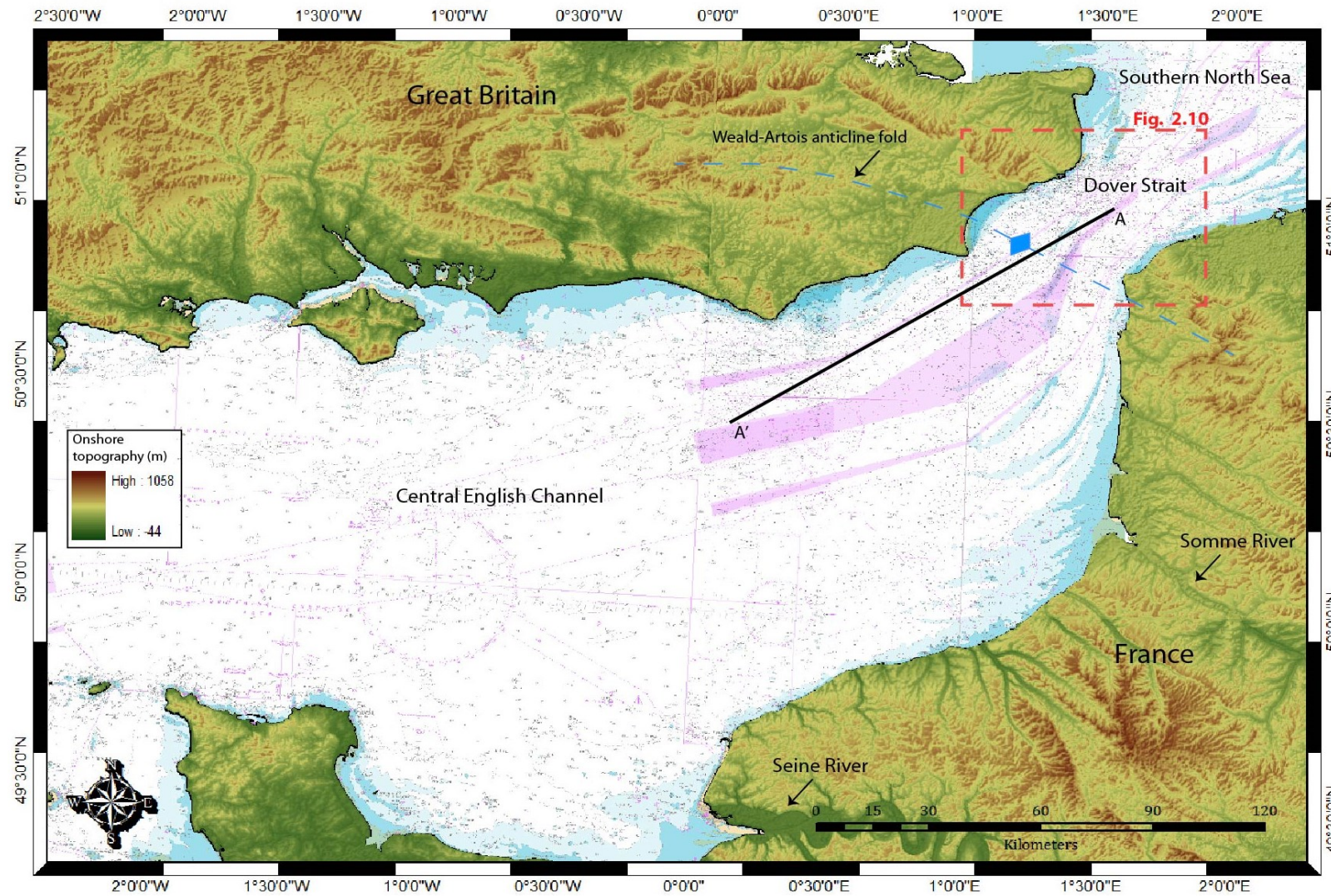


Figure 2.6: Topographic map of the present-day English Channel onshore (SRTM Topography) and offshore (Admiralty Chart from UKHO) areas, showing the location of the profile (A-A') in Fig. 2.9. The dashed red box shows the location of Fig. 2.10. The blue dashed line represents the Weald-Artois fold axis.

In the French sector of the Channel (Fig. 2.7), a south-east trending network of channels joins the Northern Palaeovalley, representing the offshore continuation of Somme and Seine rivers (e.g. Auffret and Larssonneur, 1971; Larssonneur and Walker, 1982; Lautridou et al., 1999). The first channel (Somme) is 70 km long and flows with a south-west direction. This palaeo-channel average water depth is of 70 m. The second channel (Seine) measures more than 20 km and joins the palaeovalley in the Cherbourg area (Fig. 2.7). There is no direct evidence of the age of the incisions of these two channels: these are interpreted as the result of multiple incisions during regressions, as a consequence of increased run-off from north-east of France at periglacial times (e.g. Auffret and Colbeaux, 1977; Hamblin et al., 1992; Lericolais et al., 2003).

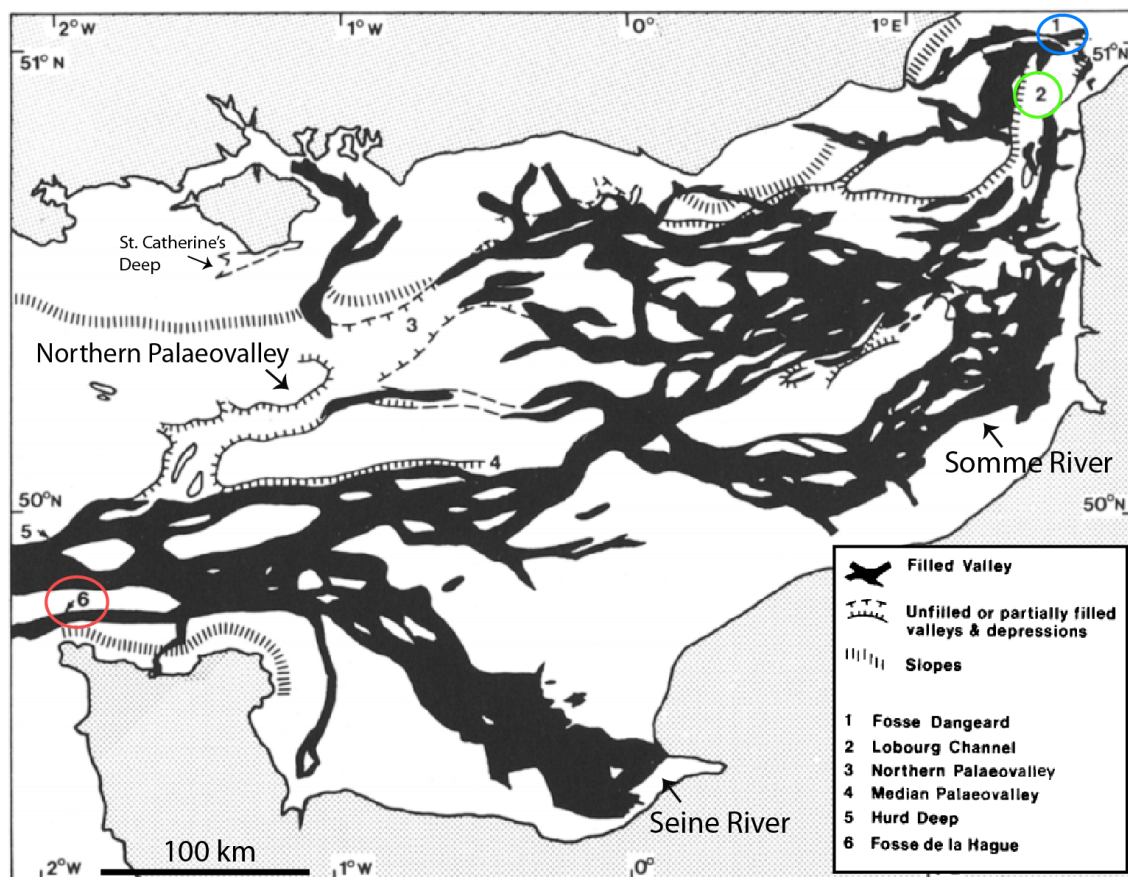


Figure 2.7: Map of the Channel River system (black areas) and of the deeps carved into the seabed. The coloured circles in the upper-right corner represent the location of the Fosse Dangeard (blue), Lobourg Channel (green) and Hurd Deep (red). From Smith, 1985.

2.3.2 Structural evolution of the English Channel

Basin evolution

A number of sedimentary basins underlie the British continental shelf. The English Channel is underlain by smaller basins with distinct character and geology. The basins are product of two main tectonic phases (e.g. Ziegler, 1978; Chadwick, 1993) and are, from north-east to south-west the London Basin, Weald Basin, the Hampshire Basin, the Channel Basin, the Central Channel Basin and the westernmost Western Approach Basin.

Two major tectonic events affected NW European continental shelf: rifting phase, with fragmentation of Eurasian and North American plates (Chadwick, 1986) and Alpine orogenesis. The first produced extensional features, while the second produced compressional structures (Ziegler, 1978, 1987, 1994; Dewey, 2000; Gibbard and Lewin, 2003) with the reactivation of old tectonic structures that were mainly concentrated along reactivated major low angle Variscan thrust zones (Cheadle et al., 1987).

During early Paleocene, rifting of Greenland and European plates caused thermal uplifting of Scotland, attributed to magmatic underplating resulted from a mantle plume located underneath northern Britain in early Tertiary times (Brodie and White, 1994). The opening of North Atlantic determined a westward migration of the plume, leaving Britain uplifted at the present elevation (Blundell, 2002). At this time, connected to activity of Variscan structures, depo-basins started to form in southern Britain, successively deformed by lithospheric stress generated by Alpine orogenesis (Chadwick, 1993; Hawkes et al., 1998; Dewey, 2000). Following the continental rifting, crustal relaxation gave origin to continued and uniform basal subsidence (Ziegler and Louwerens, 1979).

Four progressive deformation stages were recognized for the inversion of the English Channel basins (Blundell, 2002), consisting in a first early-mid Paleocene inversion that affected Celtic Sea, Western Approaches and Channel region (Ziegler, 1978, 1987, 1994). This was followed, in mid Eocene, by uplifting of Weald Basin and Isle of Wight (Gale et al., 1999) and subsidence of London and Hampshire basins. Alpine compression resulted in further uplift and inversion during post Eocene, affecting all the Mesozoic basins (Whittaker, 1985; Lake and Karner, 1987; Mascle and Cazes, 1987) and continuing intermittently into Pliocene and Pleistocene (Gibbard and Lewin, 2003).

The Portland-Wight High and Weald-Artois anticline formed at this time (Ziegler, 1978, 1987, 1994; Blundell, 2002), representing the result of compression and folding phase with major crustal flexures that display axial uplifts of over 1000 m (Chadwick, 1993).

Late Tertiary to Quaternary evolution

The existence of the Channel as a marine embayment is supported by the occurrence of Pliocene to early Pleistocene marine sediments and by a transgressive surface crosscutting the described geological structures (Lautridou, 1985; Gibbard, 1995).

From late Neogene times, the Channel area was an essentially fluvial basin, connected to the Atlantic Ocean (Gibbard et al., 1988; Lericolais, 1997). The drainage area of the fluvial basin included northern Brittany, Normandy, the Paris Basin, the Hampshire Basin, Devon, Cornwall and, episodically, the London and Rhine basins (Gibbard, 1985). A network of channels was carved into the seabed of the English Channel, known as the Channel River system, and composed of multiple branches that represented the offshore continuation of the main rivers on land. The Northern palaeovalley, located in the British sector of the Channel, represents the tract of the Channel River system that flows from the Dover Strait to the Hurd Deep. The Palaeovalley and its connected channels and deeps (called fosses), appear distinct from the traditional fluvial valleys that compose the Channel River system as there is no evidence of any connections to the onshore river system. The origin of the fosses connected to the Palaeovalley, such as St. Catherine's deep (Central Channel), Hurd Deep (Western Channel) and the Fosse Dangeard were studied in the past by several authors (Kellaway et al., 1975; Destombes et al., 1975; Smith, 1985; Lericolais et al., 2003), individually they suggested tidal, glacial or tectonic origins of these features, although their origin is still debated.

2.3.3 Stratigraphy

The bedrock succession of the study area is composed of Mesozoic to Palaeogene rocks (Figs. 2.8 and 2.9), resting on 30 km of crust that can be subdivided in three sections (Whittaker and Chadwick, 1984; Whittaker and Penn, 1986). The first 4 km of crust are composed of the sediments, underlain by 10-15 km of deformed Precambrian and Palaeozoic rocks with igneous intrusions (Hamblin et al., 1992). These rocks are cut by the 20 to 30°

dipping Variscan thrusts (Chadwick, 1993, 1986; Lagarde et al., 2003) that correspond with the boundary between brittle/ductile rocks. The deepest of the three zones extends from 20 to 30 km of depth, representing the lower crust: its upper part is composed of crystalline basement rocks while the lower part is made of granulite gneisses (Hamblin et al., 1992). A summary of the rocks that outcrop at the English Channel floor will be given in the following section, with particular interest to the upper part of the succession as Upper Jurassic to Palaeogene rocks outcrop in the study area.

Deep stratigraphy

Permian and Triassic: The rocks composing the deepest part of the succession only outcrop in the western part of the Channel. Permian breccias lay at the base of the succession and were formed in hot and semiarid climate, with evidence of volcanic activity (Knill, 1969; Cornwell et al., 1990) during lithospheric extension that caused accumulation of thick sediments in the restricted basins. The breccias become fine-grained distally, interfingering with aeolian deposits (Hamblin et al., 1992).

Lower Jurassic rocks: This part of the succession is predominantly composed of calcareous mudstones and shales with bituminous and ferruginous layers, intercalated with argillaceous limestones. At certain levels, layers of sandstone are present, sometimes sufficiently prominent to form a member or a formation of the succession (Hamblin et al., 1992). Generally, the rocks are present in a rhythmical succession that opens with a basal shale resting with sharp contact on the underlying formation and passing up to mudstone and, finally, to limestones. This sequence indicates an increase in depth (deposition of shale) followed by subsequent pulses of sedimentation and decrease of water depth (presence of sandstone) reaching a starvation of sediment that resulted in the sedimentation of limestones (Haq et al., 1987; Hamblin et al., 1992).

Middle Jurassic rocks: This series comprises the Inferior Oolite and Great Oolite groups. These are composed of carbonate platform sediments deposited in a subtropical region that, at the time of deposition, was starved of terrigenous and clastic material (Hamblin et al., 1992). Quartz and sand rich sediments are in fact rare and the grain size ranges from fine-grained argillaceous limestones to oolite grainstones, packstones and wackestones. Cyclic sequences are observed at multiple scales in the series (Arkell, 1956; Rhys et al., 1982), indicating deepening or shallowing of water depth.

Upper part of the succession and outcrops at the seabed

The upper part of the stratigraphic succession (stratigraphic column in Fig. 2.8) is the portion of the succession on which this work was focused, as it represents both the rocks into which the flood-related erosional features are carved and the rocks outcropping at the seabed in the study area. The shallow stratigraphy of the English Channel area is mainly characterised by Upper Jurassic to Upper Cretaceous rocks, with the exception of Palaeogene rocks outcropping in the Hampshire Basin. The geology is particularly affected by the presence of the reactivated folds, formed during the inversion phase of the Channel structural evolution and deforming the upper part of the succession (e.g. Ziegler, 1978; Chadwick, 1993; Lagarde et al., 2003).

Thanks to the Channel Tunnel investigation works (see Appendix A), wide knowledge of the geology of the Dover Strait area has been gained. The main objective of the investigations was the geologic and the geotechnical characterisation of the different formations present along the proposed tunnel route across the Strait. This work provided a database for investigating the geology of the Dover Strait area. A brief description of the main formations composing the Jurassic to Palaeogene succession on the Central and Eastern English Channel, including the Dover Strait area, is given in this section starting from the Upper Jurassic. The thickness of the formations, unless otherwise specified, is based on Jukes-Browne and Hill (1904) and Kennedy (1969) for the English side and on Robazynski and Amedro (1986) for the French side.

Base of upper Jurassic series: The Upper Jurassic series includes a wide range of lithologies that vary from argillaceous, calcareous to evaporitic strata. In particular, the series is composed, starting from the base, of the Kellaways Beds and Oxford Clay, represented by up to 175 m thick (Hamblin et al., 1992) micaceous and variably calcareous mudstones. These are overlain by up to 70 m thick Corallian Beds (Arkell, 1947), grey and fossiliferous sandstones interbedded with bivalve-rich and shell fragmental mudstone layers that were cyclically deposited during Corallian times.

Kimmeridge Clay: This formation, which is up to 510 m thick, overlies the Corallian Beds and is composed of a complex series of sedimentary rhythms (Cox and Gallois, 1981). They are composed, at the base, by thin siltstones to silty mudstones overlain by mudstones that become more calcareous upwards. The upper part of the rhythms is represented by

bituminous shales. Ammonite-bearing strata are used for correlation of this formation over wide areas, together with the variation in kerogen and lime content (Hamblin et al., 1992).

Portland Beds: This formation is composed of two contrasting units and is characterised by a finely interbeds: at the base it is composed of the 60 m thick Portland Sand, overlain by the 40 m thick Portland Limestone. The sandy strata contain argillaceous siltstone layers interbedded with mudstone, capped at the top of each cycle by limestone. The limestone unit is composed of massive limestone that forms strong competence contrast with the overlying Purbeck beds (Arkell, 1947; Townson, 1975; Melville and Freshney, 1982).

Purbeck Beds: This formation, which is up to 100 m thick in the Isle of Purbeck area (Arkell, 1947), marks the passage between Upper Jurassic and Lower Cretaceous, resting on the Portland beds with rapid sharp transition that results from the difference in composition of the two formations. The Purbeck Beds are characterised by a fine vertical alternation (50 to 100 cm) of laminated limestones and mudstones, whose alternation gives rise to an overall morphology which is diagnostic throughout southern England (Townson, 1975). The mudstone layers are highly eroded while the carbonate (100% calcite) beds show small scale fractures (10-30 cm). These beds were deposited in a low energy depositional environment (lagoon) as also suggested by the presence of stromatolite layers (Whittaker, 1985).

Hastings Beds (Valanginian) and Weald Clay (Hauterivian): These two formations compose the Wealden Beds Supergroup. The lower part of the Wealden Beds is composed of Hastings Beds (mainly sandy) (Lake and Karner, 1987) while the Weald Clay represents the upper part. The boundary between the two formations corresponds to the transition from sandstones to mudstones, which is generally related to a transgression in the marine environment. The Weald Clay is in fact composed of silty mudstone and muddy siltstone, with clay-ironstone beds at the base. The total thickness of the Hastings Beds in the Hastings area reaches 385 m while an almost complete succession of the Weald Clay is recorded in East Kent with a thickness of 122 m (Hamblin et al., 1992).

Lower Greensand (Aptian): This formation, composed of sandstones, is thicker on the UK side (25 m) than on the French side (15 m), and is composed of glauconitic clayey sand, with cross bedded stratification. The sediments were deposited in a tidally influenced, shallow marine environment with estuarine facies (Casey, 1963) and take the name from the green-coloured sands of the overlying Upper Greensand Formation. BGS boreholes and geophysical

studies (BGS, 1988) allowed for a detailed mapping of four units into which the Lower Greensand is subdivided in the Wealden area, up to the Dover Strait (Hamblin et al., 1992): the Atherfield Clays (shales and mudstones), the Hythe Beds (hard, greyish sandy limestone and calcareous sandstones with glauconite), Sandgate Beds (argillaceous sandstones or glauconitic silty mudstones) and Folkestone Beds (poorly consolidated quartzose sands with pebbles and thin phosphatic nodules). This last unit corresponds to an irregular morphology of the seabed where it outcrops, due to the harder and more cemented bands (Dingwall and Lott, 1979).

Upper Greensand and Gault Clay (Albian): These Albian formations rest on a regional unconformity that formed during Early Albian times (Hancock, 1969). This resulted in a deeper-water marine depositional environment in the basinal areas, where more than 90 m of Gault were deposited (Ruffell, 1991). The Upper Greensand Formation represents the shallow-water equivalent of the Gault Clay on the west (Hamblin et al., 1992). These consist of dark to light grey soft mudstones, although the basal part shows some glauconitic and calcareous layers with phosphatic nodules, representing a good seismic marker. The top of the Gault, which represents the boundary between Gault and Lower Chalk, is also a very good seismic marker (Arthur et al., 1997) as a strong difference in the lithology marks the passage from marine shales to post-rift chalk. Although the Gault maximum thickness is 90 m, BGS and Channel Tunnel boreholes show great variability in the thickness of this formation, ranging from 40 m in the eastern English Channel and then reducing to 11 m at Wissant, on the French coast. Strong variability in thickness is also observed in the Upper Greensand, whose thickness varies from 26 to less than 2 m.

Cenomanian rocks: Late Cretaceous sea level rise (Haq et al., 1988) and regional tectonic subsidence led to the Cenomanian transgression and deposition of the Chalk Formation, throughout the entire study area and part of Europe, in a 100 to 600 m deep marine environment (Hancock, 1975). Seismic reflection profiles in the English Channel show that the Chalk Group rests on the underlying formation with angular discordance of 5 to 10° (Auffret and Colbeaux, 1977), with maximum thickness reached in the Hampshire Basin where the Chalk is up to 400 m thick. The rocks composing the Chalk Group are made of micritic limestone, consisting in a matrix of calcite and debris from planktonic algae. Despite the low magnesium content of these rocks, deep-burial or tectonic stress have hardened the Chalk, also reducing its porosity (Robinson, 1986). Nodular beds are present, indicating

shallow water deposition whereas marls indicate an increase in terrigenous material (Hamblin et al., 1992). Flint levels are also present within the Chalk, filling crustacean burrows and derived from biogenic sources (Bromley, 1967). The flint levels are mainly present in the upper part of the Chalk, particularly in the Turonian Chalk, forming a bed that can be traced up to northern France and which represents a contemporaneous sedimentary event on both sides of the Channel (Mortimore and Wood, 1983). On the basis of the lithology, the Chalk is subdivided in three Formations: Upper, Middle and Lower Chalk.

Lower Chalk: This is the formation that has the highest content of terrigenous material. It is subdivided in four main units, its base represented by the Glauconitic Marl and its top by the Plenus Marl: the total thickness of Lower Chalk in the Dover area is 78 m (Jenkyns et al., 1994) thinning to 68 m at the French coast (Shephard-Thorn et al., 1972).

Glauconitic Marl: this unit is up to 7 m in thickness and comprises dark grey marls with abundant glauconite grains and a few phosphatic clasts (Bristow et al., 1997). This unit is intensely bioturbated and less rhythmically stratified in the upper part (Kennedy and Garrison, 1975). This interval is highly reflective, making a strong impedance contrast with the overlying and lower units (Arthur et al., 1997).

Chalk Marl: this unit is characterised by rhythmically stratified beds, each bed comprising basal dark bioclastic marl with an upward decrease of clay in favour of carbonate, forming a spongiferous limestone (Destombes and Shephard-Thorn, 1971). In the middle part of the unit the rhythmicity is less marked because of an increase in the clay content. This unit is 35 m thick on the English side against 25 m on the French side and is the unit in which the Tunnel was drilled (Arthur et al., 1997); the diffraction hyperbolae of the Tunnel can be used as seismic markers of this formation (see Section 4.3). Chalk Marl is subdivided into Upper and Lower units.

Grey Chalk: this unit is paler in colour respect to the Chalk Marl, less rhythmically bedded and fossiliferous than the underlying unit (Bristow et al., 1997). Its thickness is 21 m on the English side and 26 m on the French side.

White Chalk: this unit is characterised by an extremely soft, non-bioturbated and homogeneous, pale grey, marly chalk (Destombes and Shephard-Thorn, 1971). Its thickness ranges from 19 m in England to 15 m in France.

Middle Chalk (Upper Cenomanian to Mid-Turonian): This formation is characterised by massively bedded strata and is up to 90 m thick, with a higher concentration of flint in the few top meters. It is subdivided into Holywell Chalk and New Pit Chalk members (Mortimore, 1986). The upper member is composed of shell detrital and intraclastic chalks with a detrital maximum in the higher part of the succession. It contains the Melbourn Rock hardbed, consisting of hard, white, nodular chalk (Destombes and Shephard-Thorn, 1971). The lower member lacks of fossils and shell-detritus.

Upper Chalk (Turonian to Campanian): This formation is composed of micro-porous coccolithic massively-bedded limestone with beds of flint, nodular chalks and hard-grounds (Bromley and Gale, 1982). The Upper Chalk is the thickest of the three formations composing the Chalk Group, with up to 400 m of thickness preserved in Hampshire and 260 m in Dorset (Melville and Freshney, 1982).

Reading Formation: This formation represents the base of the Palaeogene series of the English Channel and is 50 m thick (Hamblin et al., 1992). It is composed of non-marine, reddish, silty clays that are visibly slumped where outcropping on the cliffs (Buurman, 1980). The sediments show a series of lenticular mottled clays and sands, locally containing pebbly beds and fine sand converted into sandstone (Bone, 1986). These beds are generally unfossiliferous and are found in the north and west portions of the London Basin and in the Hampshire Basin.

London Clay: This formation, whose thickness of 161 m has been recorded on the east of the Isle of Wight in the BGS boreholes (Hamblin et al., 1992), is composed of brown-grey clays and contains aragonitic shells, as well as brown-red sandstone with a cemented iron layer which represents a good marker in this formation (Curry, 1962; Edwards and Freshney, 1987).

Bracklesham Beds: The Bracklesham Beds are composed of grey glauconitic, clayey sands alternated to layers of sand and clay. This alternation gives this formation a characteristic layered character: the lower layer of the rhythm is made of clay, usually laminated and brown to lilac tinted (Hamblin et al., 1992). The middle layer, containing fossils, consists of highly glauconitic sand of deep green colour with seams of variegated clay. The younger, upper layer, is made of sand, loam and clay, locally a pebble band is observed at its base (Plint, 1988). The total thickness of the formation ranges between 13 to 20 m. Because of the mix of

sands and clays this formation is popular for brick making. In addition, the presence of an ironstone band may have been the source for prehistoric iron industry (Hamblin et al., 1992).

Barton Beds: This formation, which is up to 185 m thick, is composed of fine-grained yellow sands, frequently located at the top of a lower pebble bed. Marine fossils are found only occasionally in this formation (Auffret, 1973).

The BGS stratigraphic column and geologic section across the Dover Strait and eastern English Channel (Figs. 2.8 and 2.9) show the thickness of the described formations and the variation in their dip in correspondence of the Weald-Artois anticline (Fig. 2.9).

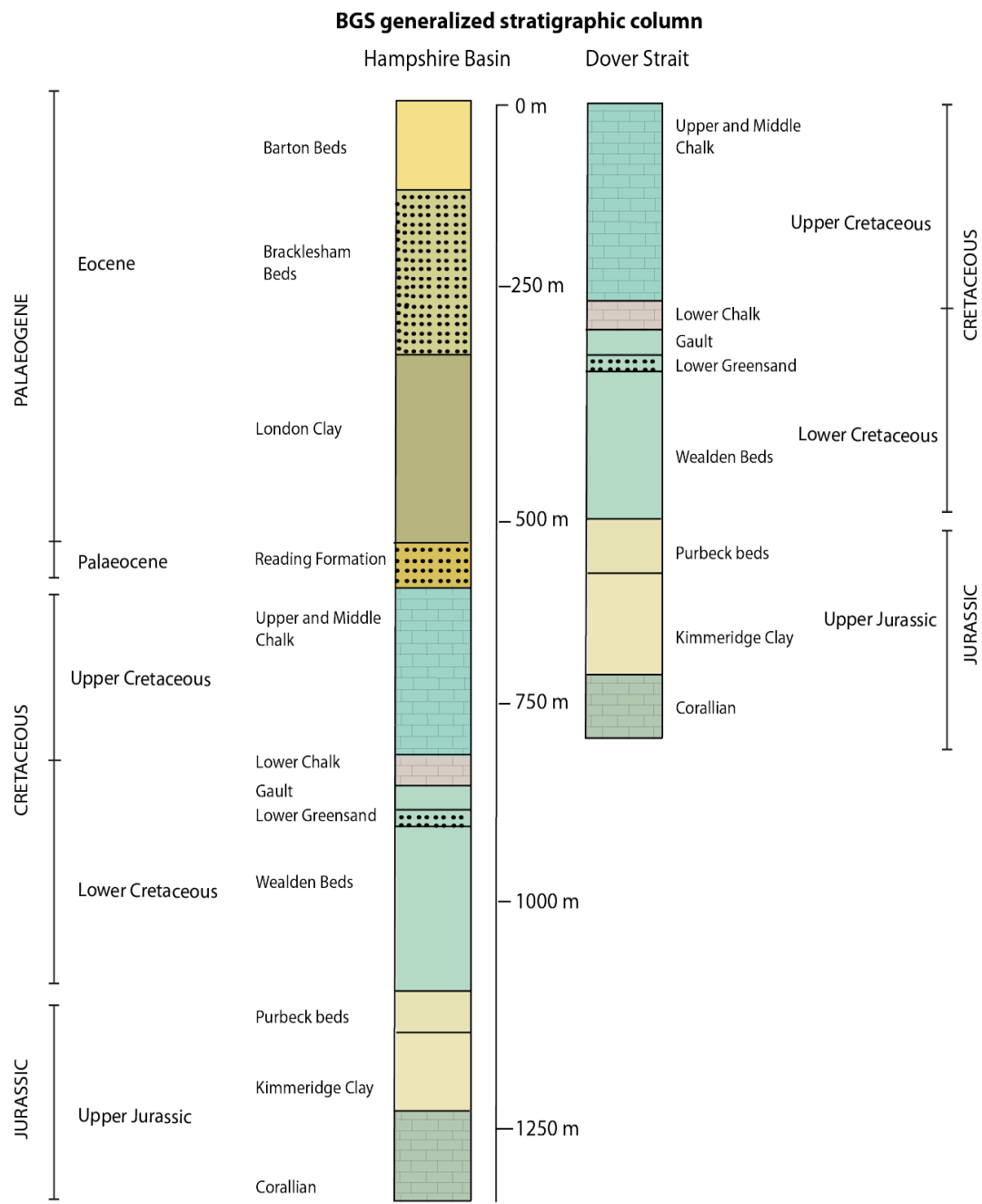


Figure 2.8: BGS stratigraphic columns of the bedrock succession in Hampshire Basin and Dover Strait areas (from BGS, 1988, 1995).

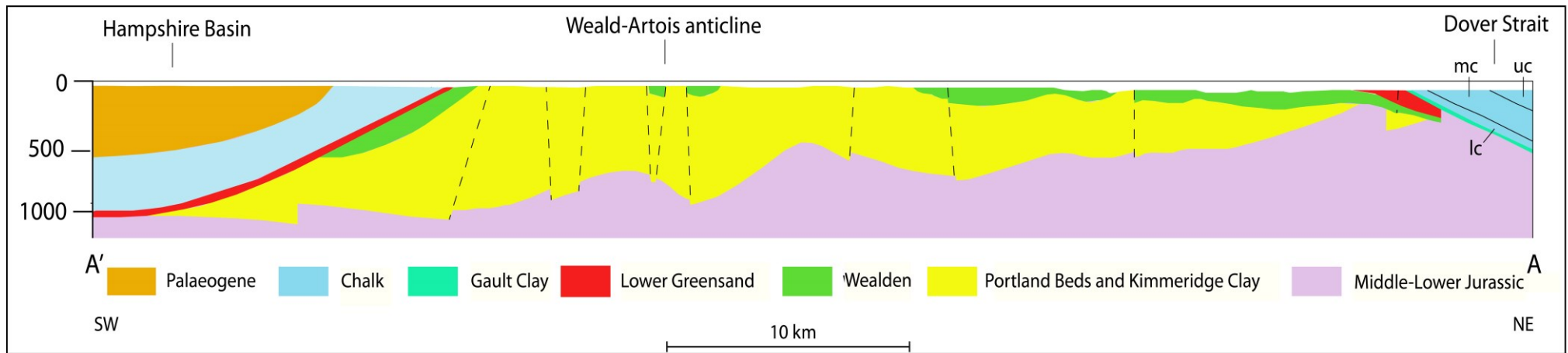


Figure 2.9: Section across the English Channel, showing the stratigraphic succession from the Dover Strait to the Hampshire Basin, crossing the Wealden Artois anticline (adapted from BGS, 1998). Dashed lines represent major faults. Location of the profile is shown in Fig. 2.6.

2.3.4 Tides and modern surficial sediments

The surficial sediments of the English Channel seabed are divided in two categories: the coarser lithoclasts with flint fragments (10 - 60 mm) and a finer (0.15 - 0.5 mm), weathered mature quartz fraction (Larsonneur et al., 1979). The coarser fraction is a residual lag deposits, overlain by the finer mobile fraction, subjected to the effect of tide. There is no significant relationship between the distribution of seabed sediments and palaeovalley infill. Where the channel infill is overlain by seabed sediments, their contact can be identified in high resolution seismic reflection data by a reflector (see Chapter 4) (Hamblin et al., 1992). The lag deposit is present everywhere in the area, with the exception of the portions of seabed where rock outcrops are present. Its typical thickness is lower than 50 cm throughout the entire area: given its low thickness, the contact is only visible on high resolution reflection seismic. The maximum grain size of the lag indicates the local velocity of tides at the moment of its formation and is represented by boulder of diameter of 50 up to 80 cm constituting blocks of bedrock (Hamblin et al., 1992). On the other hand, the main input that composes the finer mobile fraction of the seabed sediment is composed of fluvial material, mainly transported during interglacial times after Last Glacial Maximum, as the contemporary fluvial input in the area is limited (Laignel et al., 2008). The biggest river discharging into the English Channel is in fact represented by the Seine River with 500,000 to 1,000,000 t/yr (Avoine et al., 1986) of which the vast majority is deposited in its estuary (Avoine et al., 1981) and does not reach the Central English Channel.

Sediment transport in the area is strongly affected by the tidal regime: at spring tide, tidal speed varies from 1 m/s up to 4.6 m/s with value of 2 m/s in the Dover Strait area (Hamblin et al., 1992). The bottom stress driven by tidal stream is a function of water depth, seabed roughness and tidal velocity. Wave height, due to high wind speed, is a secondary factor, as it does not seem to have a direct effect on sediment transport and distribution as wave in the Channel have much lower energy than tidal currents. In general, sediments are patchily distributed in areas dominated by weaker currents. In particular, thickest sediment cover is observed off the coast of Devon, south-west of the Dover Strait and in the Beachy Head area. These areas are dominated by highly mobile sediment banks that have formed during early Holocene transgression, due to the effect of the currents. A bed load parting area is located between the Isle of Wight and the Cotentin where the effect of ebb and flood currents remove medium-grained material (Hamblin et al., 1992).

Conversely, a bed load convergence area is located south-west of the Dover Strait, where the kilometre scale sand banks are located. The crest of the sand banks is overlain by sand waves and ripples, showing deposition from both the north-east and south-west, as indicated by the variable orientation of their asymmetric sides and growth of the sand banks in both directions (Hamblin et al., 1992). A map of the surficial sediment, drawn using the new data is presented in Chapter 4 of this work and compared to the described tidal sediment distribution.

2.4 The Dover Strait

The Dover Strait area (Fig. 2.10) is the offshore corridor that connects the North Sea to the English Channel. Most importantly, it is the proposed breach-point of the catastrophic outburst that led to the isolation of Great Britain from Europe (Smith, 1985; Gibbard, 1995; Gupta et al., 2007). For this reason, the Dover Strait is a key area for the understanding of the Quaternary palaeo-geographic history of north-west Europe and its river drainage system, as well as the variations in Pleistocene human occupation and migration in England and north-western Europe (e.g. Gibbard et al., 1988; Stringer et al., 1998; Bridgland et al., 2006; Ashton and Hosfield, 2010). A brief summary of the literature review on the Dover Strait geomorphology and Quaternary palaeo-geographic history is given in this section.

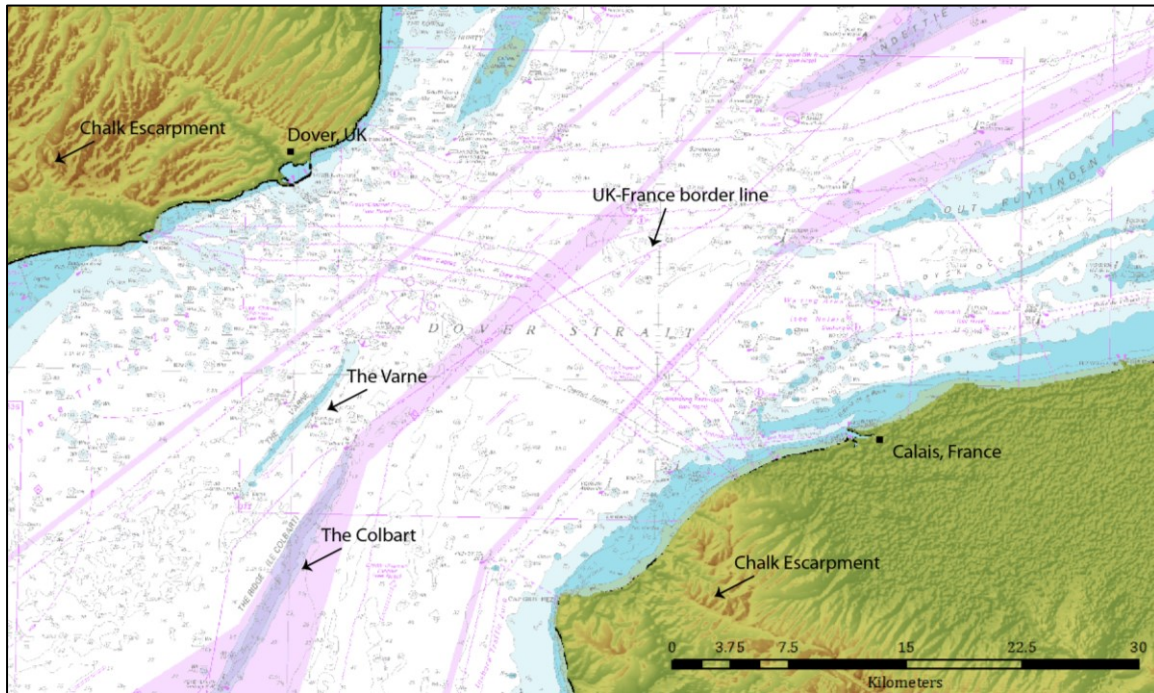


Figure 2.10: Image extracted from the topographic map in Fig. 2.6 showing the Dover Strait Admiralty Chart (UKHO) showing the UK-France border and the areas dominated by major surficial sediments and sand banks (blue areas). For onshore topography (STRM topography) colour-scale see Fig. 2.6.

2.4.1 Geomorphology of the Dover Strait

The Dover Strait area is characterised by depths ranging from 30 to 65 m: its Quaternary palaeo-geographic history has therefore been strongly affected by the changes in the sea level, cyclically changing from submerged to emerged areas. The Dover Strait reached a maximum height above sea level of about 5 m higher than present 120 kyr ago and a low level of -130 m about 18 kyr ago, during glacial times (Lambeck et al., 2002). The most characteristic feature of the Dover Strait geology is the presence of the Weald-Artois anticline (Figs. 2.6 and 2.9), whose axis crosses the Dover Strait with a WNW-ESE orientation, separating the area into a north-eastern and south-western limb of the anticline. At the eroded core of the anticline fold (Fig. 2.9) outcrops the Wealden Formation, while the north-eastern flank is dominated by the presence of the Chalk, forming an up to 100 m high step that marks the passage between the Lower Chalk and the underlying formations, known as Chalk Escarpment. This feature, clearly visible both onshore (Figs 2.10) and offshore, is one of the key elements in the evolution of the ice-marginal lake. More detailed description of the escarpment and bedrock geology is provided in Chapter 4.

Erosional features

Two erosional features are present in the Dover strait area: the Lobourg Channel and the Fosse Dangeard (Fig. 2.7). The Lobourg Channel is an up to 70 km long, 8 km wide and up to 25 m deep sediment free channel that crosses the Strait with a NE-SW direction, flowing from the Southern North Sea into the English Channel. The Fosse Dangeard, on the other hand, is a system of sediment-filled depressions carved into the bedrock and scattered across the Dover Strait. These depressions were first discovered and mapped during the Channel Tunnel investigation works (see Appendix A) by Destombes et al. (1975), who proposed the channels to be the product of glacial erosion during the Warthe phase (0.14 Ma) of Saalian glaciations (0.4 to 0.14 Ma). A detailed study of the upper part of the infill of the Fosse Dangeard was also made through the analysis of a core sampled during the seismic reflection survey. From palynological data analysed on the core, the age of the infill is dated to Eemian (0.11 Ma) and Broup (0.1 Ma) interstadial (Morzadec-Kerfourn, 1975) (ages are referred to Global chrono-stratigraphic correlation table for the last 2.7 million years, ICS-SQS, 2010). These dates were obtained from the upper part of the infill only and are probably not representative of the age of the entire infill and of the Fosse. The detailed analysis of the Lobourg Channel and Fosse Dangeard is described in Chapters 5 and 6 of this work.

Holocene depositional features

As for the English Channel sediment cover, the Dover Strait is a relatively sediment-free area dominated by high-energy currents, with the exception of the convergence area located on its south-western portion. The sediments that cover the seabed are swept and driven by tidal currents redistributing sedimentary bedforms, deposited on top of erosional bedforms carved into the bedrock (Hamblin et al., 1992). Strong tides dominate the Dover Strait area, showing a 7 m range and velocities around 3 knots (Hamblin et al., 1992). The only large-scale surficial depositional features present in the area, other than the coarse grained seabed sediments, are sand banks. Their formation postdates the flooding of the Strait (Smith, 1989), after which these features developed through the reworking and sedimentation of glacial and fluvio-glacial deposits during transgressive times (e.g. Mellet et al., 2013). The dimension of the sand banks varies from 10 to 20 km in length and 1 to 1.5 km in width, with a 20 to 25 m height. The Varne and The Colbart sand banks are present at the Dover Strait (Fig. 2.10): these structures are currently stable in size, indicating stable hydrodynamic conditions, with their crests characterised by well sorted medium grained sand (Hamblin et al., 1992).

2.4.2 Quaternary history of the Dover Strait

A brief summary of the previous research carried out on the palaeo-geographic history of the study area is summarised in this section, with the aim of illustrating the main factors that must be kept into account when performing palaeo-geographic reconstruction and analyses that represent one of the main targets of this work, as illustrated in the final model presented in Chapter 7.

Middle and Late Pleistocene ice-sheet limits in the Southern North Sea

British glacial history has been divided into four glaciations, separated by interglacial times (Geikie, 1894) until, after 1970's the number of glaciations was reduced to three (Shotton, 1968, 1976, 1977). The first Quaternary glacial sediments known on land in Britain are represented by glacial diamicton, regarded as of Anglian age (Gibbard and Clark, 2011), and related to the Anglian glaciation (0.43 Ma at MIS 12), corresponding to the Elsterian event (Fig. 2.11b) in eastern Germany (Eissmann, 2002). This event was the first and major Quaternary glaciation: at that time, the British Ice Sheet reached its maximum extension during Quaternary and, for the first time, coalesced with the Scandinavian Ice Sheet (Oele and Schuttenhelm, 1979). Importantly, the southern advance of the ice-sheet in southern England and Dover Strait area during Anglian glaciation strongly affected the palaeo-geography of the area, as the presence of the ice-sheet margin dramatically changed the landscape, relief and river drainage configuration (e.g. Gibbard, 1995; Busschers et al., 2008; Gibbard and Clark, 2011), as described in the following section where the opening of the Dover Strait is discussed. The second Pleistocene glaciation is dated to Wolstonian times, equivalent to German Saalian glaciation (0.16 Ma at MIS 6), although glacial sediments of this age are only recognised in the Midlands (Gibbard et al., 2009) and poorly represented in the Pleistocene glacial record.

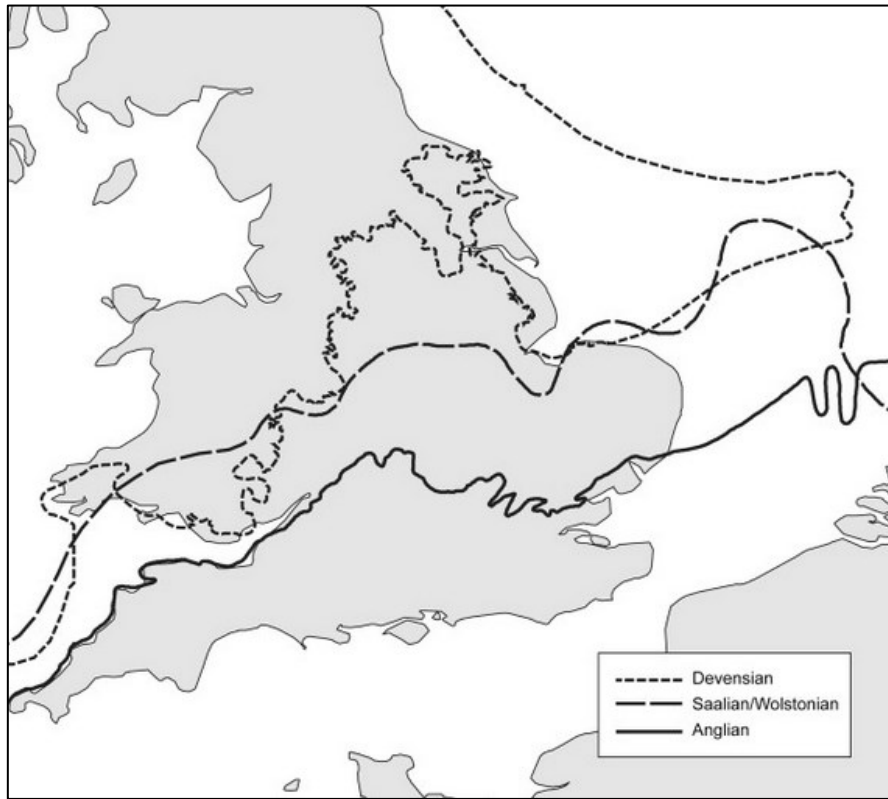


Figure 2.11a: Maximum ice-sheet extent for the three major glaciations recorded in southern Britain (from Gibbard and Clark, 2011).

Associated to the ice-sheet margin is the formation of pro-glacial lakes (Figs. 2.12 and 2.13), by water filling the depression created by the ice-sheet isostatic load (Marshall and Clarke, 1999). The principal periods of glacial-lake formation in Britain relate to glaciations during MIS 12, 6 and 2 (Murton and Murton, 2012). In the Dover Strait area, following the Anglian glaciation (MIS 12), a pro-glacial lake was present in the Southern North Sea (Fig. 2.13), bounded on the north by the coalescent ice-sheets (Smith, 1985; Gibbard et al., 1988; Gupta et al., 2007; Gibbard and Clark, 2011). These extended from eastern East Anglia across the North Sea into northern Belgium, Germany and the Netherlands and were fed by north-western Europe drainage network. The lake is thought to have measured 550 km in width (east-west) and 250 north-south, with a total area of 140,000 km² (Murton and Murton, 2012). A detailed description of Middle Pleistocene sediments (Cameron et al., 1987, 1992; Catt et al., 2006) indicate the Swarte Bank Formation (SBF) (Long et al., 1988) as composed of fluvial sediments derived from prograding delta of major north-west European rivers such as Thames, Rhine and Meuse (Cameron et al., 1987). This evidence is used for the reconstruction of the glacial palaeo-geography of the area as the middle member of SBF is

composed of glacio-lacustrine clay and is interpreted to have been deposited during Anglian glaciation (Long et al., 1988).

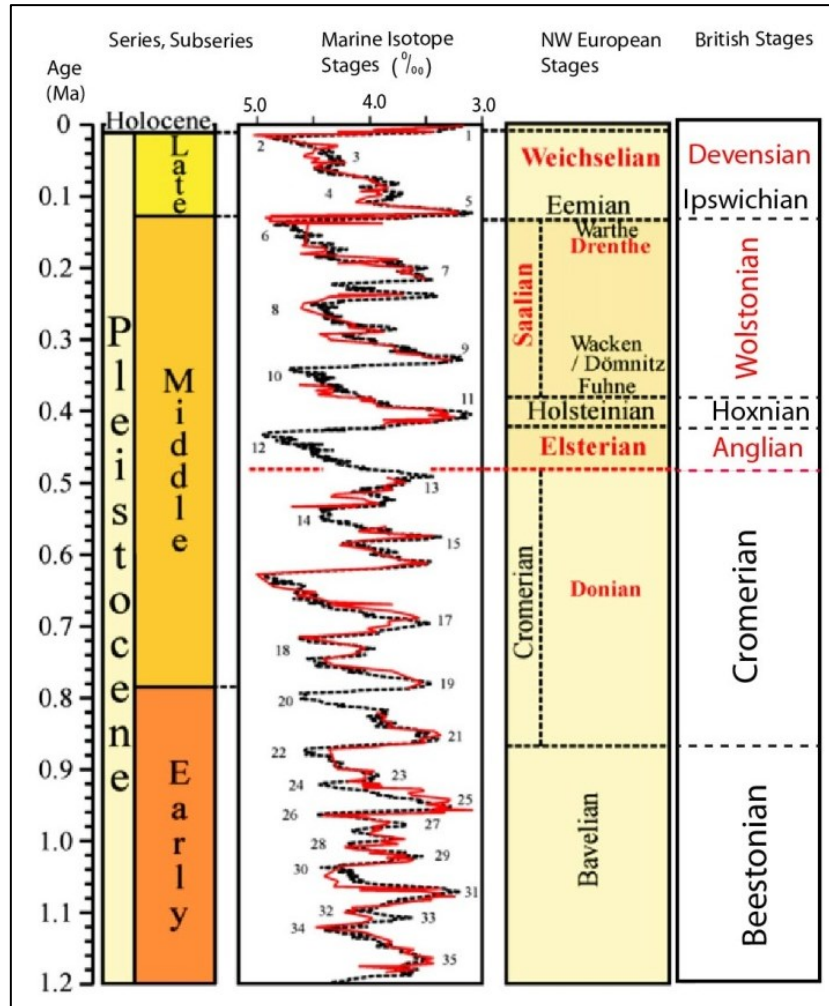


Fig. 2.11b: Nomenclature and ages of Pleistocenic NW European and British glacial and interglacial stages. Cold stages are indicated by red colour, interglacial stages are black. (Modified from Toucanne et al., 2009b).

With regard to the Saalian glaciation, the presence of a pro-glacial lake in the Southern North Sea, and in particular offshore of the Netherlands, requires the coalescence of British and Scandinavian Ice Sheets as a prerequisite. This has been supported by several authors (Rappol, 1987; van der Berg and Beets, 1987; Ehlers 1990; Kluiving et al. 1991; Murton and Murton, 2013). Cohen et al. (2005) and Busschers et al. (2008) also support the hypothesis of the presence of a pro-glacial lake offshore Netherlands during MIS 6 and bounded by the coalescent ice-sheets, suggesting overspilling of the lake during Drente ice-

sheet disintegration. This is supported by sedimentological evidence represented by prograding delta sediments from Rhine and Meuse rivers, deposited at an unusual high elevation corresponding to present sea level (Busschers et al., 2008). These are interpreted as sediments deposited by rivers Rhine and Meuse, diverted by the presence of the ice-sheet, depositing in a marginal-lake setting (Unit S4 of Busschers et al., 2008). The observed drop of 10 m in the sediment base level (Unit S5 of Busschers et al., 2008) is interpreted by the authors as incision of River Meuse as result of lake level drop, following the drainage (Fig. 2.12). The presence of a Saalian lake is also supported by the presence of varved clays in the same area (Cameron et al., 1986; Joon et al., 1990; Laban, 1995) and subglacially scoured basins in the northern Netherlands that are interpreted to have formed one interconnected lake, named Holland Glacial Lake (Beets and Beets, 2003).

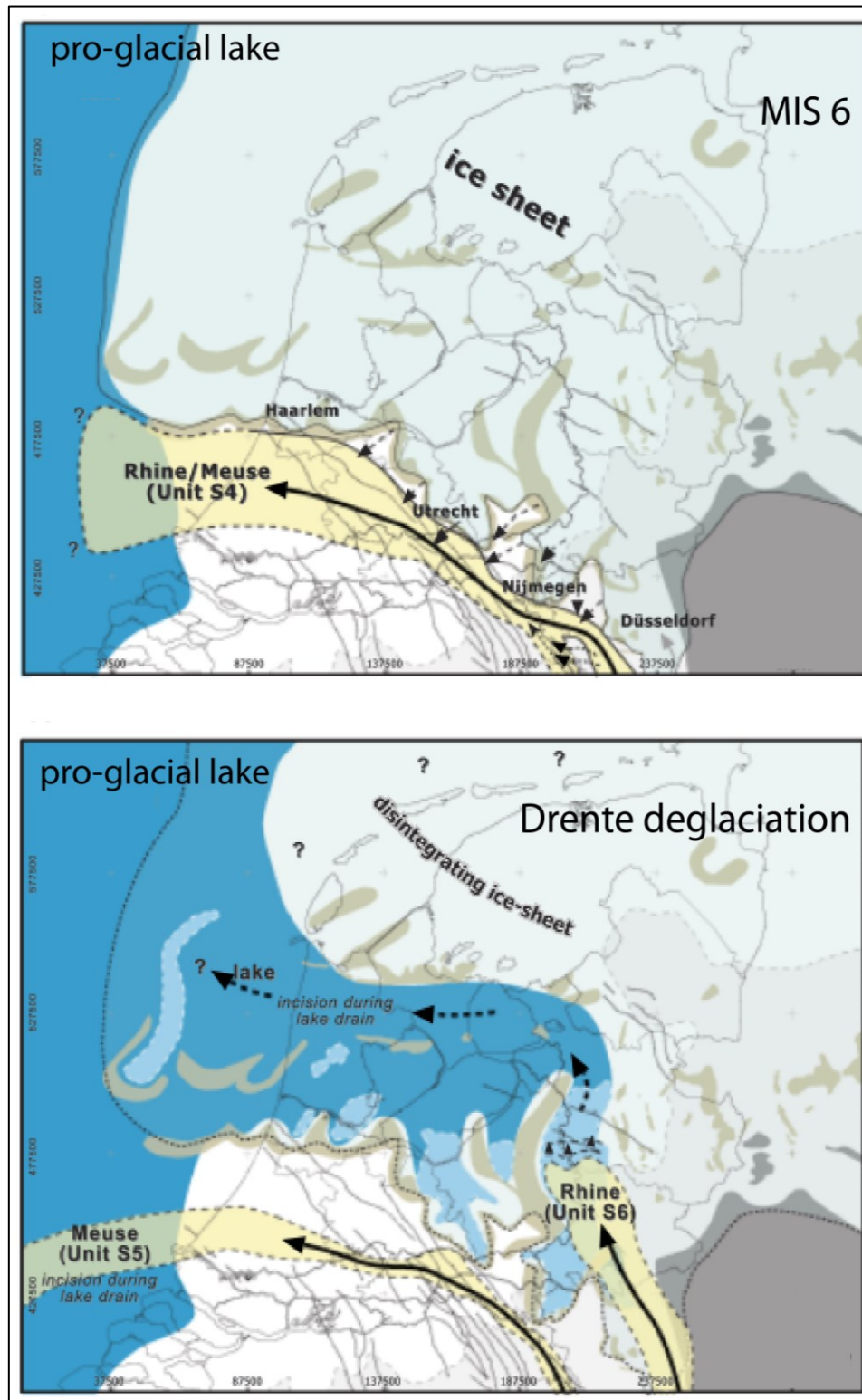


Figure 2.12: Location of the sediments (yellow area) interpreted by Busschers et al., 2008. The blue area represents the extension of the pro-glacial lake, the light blue area is subglacial basins. At MIS 6 (upper box) the Rhine-Meuse System (Unit S4) entered the pro-glacial lake located offshore the Netherlands. During Drente deglaciation phase (lower box), the Rhine deposited sediments in the former ice-sheet area while the Meuse (Units S5) dissected former Unit S4 during and after lake drainage. From Busschers et al., 2008. See Fig. 2.11b for age of NW European and British stages.

Glacial isostasy during Middle and Late Pleistocene glaciations

A brief summary on previous research carried out on glacial rebound effects in the study area is given in this section, as glacial rebound is one of the factors taken into consideration for the palaeo-geographic final model presented in Chapter 7. Glacial-related lithospheric isostatic adjustments result from ice-sheet loading and unloading during glaciations, involving compensatory crustal depression (in response to loading) and elevation (to unloading) as illustrated by Lambeck (1993, 1995) for the Southern North Sea and Dover Strait areas. Crustal movements due to glacial rebound have a strong impact on relative sea level changes and must therefore be taken into account when performing palaeo-geographic reconstructions (e.g. Maddy et al., 2000, 2001; Westaway et al., 2002; Busschers et al., 2007, 2008; Lewin and Gibbard, 2010; Cohen et al., 2011) on those areas occupied by, or adjacent to ice-sheet, during glaciation times. As the study area is located south of the ice-sheet margin during Anglian and Saalian glaciations, glacial-related crustal deformations are observed from variation in channel slopes (negative in the Lobourg Channel) in relation to the three main Middle and Lower Pleistocenic glaciation events described in the previous section.

Geophysical models of glacio-isostatic movements during ice-sheet advance and retreat phases of late Weichselian age are well time-constrained, thanks to the availability of data from raised beach systems in Scandinavia and Scotland, where altitude and chronology of a sequence of raised shorelines have been analysed by a number of authors (e.g. Dawson, 1984; Gray, 2008; Sissons, 2008). These give evidence of crustal movements due to isostatic depression of lithosphere under loading ice-sheet, and subsequent isostatic rebound since last glacial maximum (Lambeck, 1995; Peltier, 2004). Subsidence of non-glaciated areas parallel to the ice-sheet margin suggests collapse of peripheral up-warped zones in areas adjacent to former ice sheet (forebulge crest zone) (e.g. Lambeck 1995; Kiden et al. 2002; Peltier 2004; Steffen et al, 2006; Busschers et al., 2008).

With regard to Anglian and Saalian glaciations, although the impact of isostatic glacial rebound on lithosphere is more difficult to reconstruct. It is particularly difficult to separate glacial rebound from non-glacial tectonic and fluvial effect as these processes can affect relative sea level through deposition and incision of sediments that have the potential to affect sediment relative base level in the stratigraphic succession, e.g. Maddy and Bridgland, 2000; Maddy et al., 2001). Due to limited sedimentological evidence, it is only possible to

identify crustal adjustments related to ice-sheet advance and retreat, in the areas located in correspondence or vicinity of the British and Scandinavian ice-sheet margin. In relation to Anglian glaciation, isostatic rebound may have consisted of an initial phase of rapid glacio-isostatic rebound during the late phase of glaciation (Maddy and Bridgland, 2000), followed by an adjustment during a subsequent phase of less rapid rebound. Maddy (1997), Bowen (1994), Preece et al. (1990) and Maddy et al. (1999) have calculated the isostatic rebound rate for Anglian glaciation, which appears to be of the order of $0.07\text{--}0.10\text{ m kyr}^{-1}$. Although the isostatic model may have been affected by forebulge effects (Walcott, 1970) causing progressive uplift-depression-uplift phases (Maddy and Bridgland, 2000) in the immediate post-Anglian, although it is not possible to exactly constrain this effect due to a lack of temporal resolution in the sedimentary sequence.

The forebulge effect has also been recognised to have affected glacial rebound deformations during late Saalian glaciation times (whose palaeo-geography is illustrated in Fig. 2.12) in those areas situated in the direct vicinity of the ice-sheet, which is estimated to have been at least several hundred metres thick (Schokking, 1998). The study area may therefore have been located between the peripheral forebulge and peripheral depressed region, suggesting moderate isostatic depression during late Saalian times (Busschers et al., 2008). Sedimentological data from the Rhine-Meuse river system offshore the Netherlands, presented by Busschers et al. (2008) indicate relative high position of sediments that may have represented accommodation phase during a glacio-isostatic subsidence (suggested by the authors to have been of relative minor magnitude, $\sim 10\text{ m}$). This would have occurred at times maximum ice extent, with subsequent incision of the river system, indicating possible response to glacio-isostatic rebound of the area. As detailed in the previous section, elevation of Rhine-Meuse river sediments (units S4 and S5) presented by Busschers et al. (2008) represent one of the main available indicators of the elevation of the pro-glacial lake located offshore the Netherlands at late Saalian times. It is therefore crucial to consider the effect of glacial rebound on the elevation of the sediments and possible implications on the performed palaeo-geographic reconstructions.

During Pleistocene glaciations, the study area experienced significant glacio-isostasy-driven crustal warping (forebulge build-up and collapse) due to its particular distance to the center of ice-sheet that formed in Scandinavia and Britain (Morner, 1979; Peltier, 2004; Lambeck, 1995; Kiden et al., 2002; Lambeck et al., 2002). Forebulge effect formed as a

consequence of simultaneous crustal suppression below the Scandinavian and British ice-sheets (e.g. Steffen, 2006). These movements occurred as a response to strong lithosphere suppression beneath the ice sheets, causing the foreland areas around the ice-margin to host an uplifted zone (forebulge) (Busschers et al., 2007). This effect would be followed by anomalous subsidence of the forebulge areas during deglaciation, in opposition to uplift of formerly glaciated areas where the centre of the ice-sheet was located. This is confirmed by reconstructions of relative sea-level rise for Late-glacial to Middle Holocene times, testifying for the occurrence of anomalous high subsidence rates along the southern North Sea coasts. On the other hand, the deglaciated Scottish and Norwegian coasts show strong uplift (Kiden et al., 2002). The study area would therefore have experienced peripheral glacio-isostatic uplift during glacio-isostatic subsidence in the beginning of the interglacials (Busschers et al., 2007; Vink et al., 2007, Cohen et al., 2012). The most recent generation of glacio-isostasy models, calculated on the the Late Pleniglacial glaciation, indicate an area of maximal uplift to be situated in the northern Netherlands with relative upwarping amounts along the Rhine of 5-10 m. This resulting glacioisostatic effects would have further differentiated the sea-level signal between, for example, the Bay of Biscay, the English Channel and the southern North Sea sub-regions (Cohen et al., 2012).

The opening of the Dover Strait

The opening of the Strait was the first time the Southern North Sea and the Atlantic Ocean being connected through the English Channel (Gibbard, 1995; Smith, 1985; Toucanne et al., 2009a, 2009b). Despite this, the exact timing and mechanisms of the formation of the Strait is unclear.

During Early and Middle Pleistocene the Southern North Sea basin was a dry, emerged area, incised by the network of major north-west European rivers flowing in the Dover Strait area (Gibbard, 1995). Although some authors propose that the Dover Strait formed by extensional movements at the end of the Early Pleistocene, no sedimentary support was given to this interpretation (Colbeaux et al., 1980). At this time, during lowstands, the Strait is likely to have been a land-bridge (Stamp, 1927; Stamp, 1936; Keen, 1995; Gibbard, 1995; Hijma et al., 2012), going through cyclical variations of exposure and submersion following the changing sea level. On the other hand, during high sea level at Middle Pleistocene times, evidence of marine environment at the Strait was found in Southern

England (Gibbard et al., 1991), suggesting that a possible sea passage may have existed, although there is no sedimentological evidence for an opening of the Dover Strait during that time. This would indicate that throughout most of the Pleistocene, including the early Anglian/Elsterian Stage, the Strait did not exist (Gibbard, 1995).

Several authors agree on the theory that the Dover Strait was the outlet point for the overspilling of the Elsterian pro-glacial lake located in the Southern North Sea (Belt, 1834; Ussher, 1913; Stamp, 1927, 1936; Gullentops and Huyghebaert, 1974; Roep et al. 1975; Jones, 1981; Smith, 1985; Ehlers and Rose, 1991; Gibbard, 1995). Different mechanisms were addressed in order to explain the breaching of the Strait: Smith (1985, 1989), Gupta et al. (2007) and Gibbard (2007) suggest a catastrophic overflow from the North Sea lake (Fig. 2.13), also supported the presence of coarse conglomerate in the succession found in Wissant, France (Roep et al., 1975), interpreted by the authors as deposited by high energy flows. Destombes et al. (1975) suggested glacial processes to have carved the Strait, while Hamilton and Smith (1972) addressed the polycyclic alternation of low and high sea-level as the driving process responsible for the erosion of the palaeovalley system connecting the Southern North Sea to the Channel. This theory was supported by evidence of glacio-lacustrine conditions persisting until the end of Elsterian/Anglian Stage (0.48-0.40 Ma), according to the authors. This age is widely accepted as the time of the overspilling (Gibbard, 1995), with following minor events identified in the channel-fill sequence in Wissant. Smith (1985) and Gibbard (1995) suggest that the drainage of the ice-dammed lake and the following uplift of the Strait area, due to isostatic recovery, could have encouraged the river Thames and Scheldt to be confluent, flowing into the Channel River System.

The evolution of the Strait during the following Holsteinian/Hoxnian Stage (Fig. 2.11b) is unclear, although palaeontological evidence (Meijer and Preece, 1995) suggests that a sea-passage might have been open during the period of maximum interglacial sea-level. However, if the Strait was open, it would probably have been relatively narrow, perhaps 1-4 km wide (Keen 1995; Gibbard, 1995), as there is no evidence of complete erosion/opening of the Strait at that time.

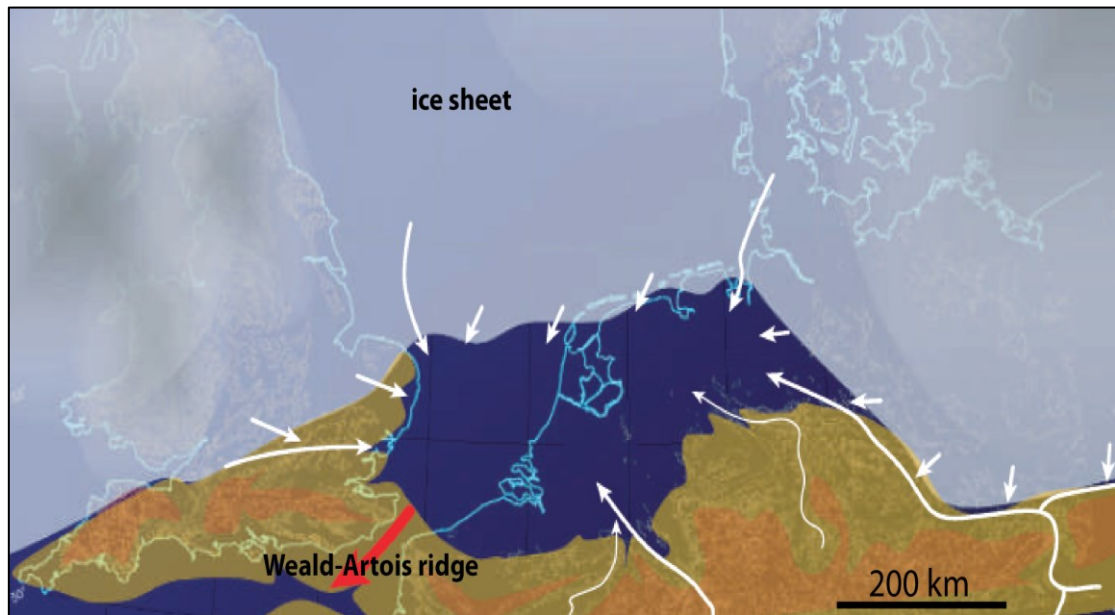


Figure 2.13: Palaeogeography of the Southern North Sea and English Channel areas at MIS 12. The white arrows represent the path of the river drainage system into the lake, the red arrow shows the direction of the proposed drainage of the lake, with water catastrophically flowing into the English Channel (from Gibbard, 2007, modified).

A rerouting event of the north-west European rivers took place at MIS 6 (e.g. Gibbard, 2007; Busschers, et al., 2008; Toucanne et al., 2009a, 2009b) in response to the Saalian/Drente ice-advance. At this time the river Rhine and Meuse draining southwards (Bridgland and D'Olier, 1995; Jelgersma et al., 1979; Oele and Schuttenhelm, 1979). This is considered as the time for the second breaching of the Strait after a first erosion at MIS 12 (Gupta et al., 2007; Gibbard, 2007; Busschers et al., 2008; Meinsen et al., 2011) and is supported by a sudden increase in the presence of terrigenous material in cores located at the Bay of Biscay (Eynaud et al. 2007; Toucanne et al. 2009a, 2009b). Although the peak in volume of terrigenous material does not prove the opening of the Strait, it certainly indicates a strong increase in sediment discharge through the Channel River at MIS 6. Meijer and Preece (1995) also give evidence of marine environment at the Strait at Eemian Stage on the basis of the occurrence of Mediterranean mollusca in the southern North Sea region. The proposed mechanisms that led to the opening of the Strait at MIS 6 are still debated: some authors suggest a second catastrophic event (Gupta et al., 2007; Gibbard, 2007; Busschers et al., 2008; Meinsen et al., 2011) while other authors favour a more “seasonal” behaviour of the Channel River discharging non-catastrophically into the Atlantic Ocean (Eynaud et al, 2007; Toucanne et al., 2009a, 2009b).

Geochemical and Palaeontological data

A variety of data were used in order to constrain the times and number of events for the opening of the Dover Strait, as well as the nature of the processes that led to its breaching. In particular, sedimentological (palaeo-shorelines), geochemical and paleontological data indicate at least two episodes of breaching of the Strait (e.g. Gupta et al., 2007; Toucanne et al., 2009a; Ashton and Hosfield, 2010; Meinsen et al., 2011). Despite this, the reconstruction of the Dover Strait palaeo-geography remains affected by the poorly constrained timing of the events and sea level history and by the lack of clear sedimentological evidence. Although there is wide acknowledgement on the timing of the first breaching of the strait at MIS 12 (e.g. Gibbard, 1995; Gupta, 2007; Toucanne et al., 2009b), the following phases of isolation and connection of Britain to the continent remain unclear. Despite palaeo-shorelines and palaeo-beaches in Southern England that can be used to reconstruct the Pleistocene sea level history, a number of factors affect the reliability of these indications. Firstly, the displacement of the deposits can be primarily due to long-term tectonic effect that affected the North Sea area since Mesozoic times (Keen, 1995), depressing only the marine deposits. Another factor that must be considered is the glacio-isostatic forebulge effect, as well as the hydro-isostasy related to transgression and regression of the sea, causing depressions and lifting of the crust, as detailed earlier in this section. Finally, although the ice-sheet limit is thought to have been located in the Southern North Sea (Clarke et al., 2004), the forebulge effect would have affected areas up to 400 km from the ice sheet limit (Walcott, 1972), causing uplift during cold stages and subsequent depression during interglacial times. Finally, the effect of tidal regime that can affect the displacement of marine deposits, depositing sediments up to 14 m (Carr and Blackley, 1973) above sea level.

Geochemical analyses of sediments discharged by the Channel River during Pleistocene are an important source for the reconstruction of the Quaternary palaeo-geography of the Dover Strait. Data from the Bay of Biscay (Eynaud et al., 2007; Toucanne et al., 2009a, 2009b) were analysed in order to reconstruct the Quaternary history of the interaction between the ice-sheet and the fluvial network sediment discharge in north-western Europe. In relation to the timing and history of the opening of the Dover strait, the cores located at the mouth of the English Channel give evidence of a first connection between the Atlantic Ocean and the southern North Sea, through the Dover Strait, at MIS 12. This is indicated by a steep increase in the mass of terrigenous sediments in the cores (Fig. 2.14), suggesting a strong increase in the fluvial discharge at this time. The increase is partly

interpreted as the result of the presence of increased volumes of water flowing through the English Channel as the result of the diversion of the river network caused by the advance of the southern margin of the British and Scandinavian ice-sheet (Gibbard et al., 1988; Sejrup et al., 2000, 2009) that were coalescent for the first time during the Elsterian glaciation (e.g. Stokes and Clark, 2001; Evans et al., 2005). Together with the input of the diverted river network, another factor that contributed to the increased discharge into the English Channel was the presence of ice-sheet melt-water flowing into the Channel for the first time (Gibbard et al., 1988; Busschers et al., 2007).

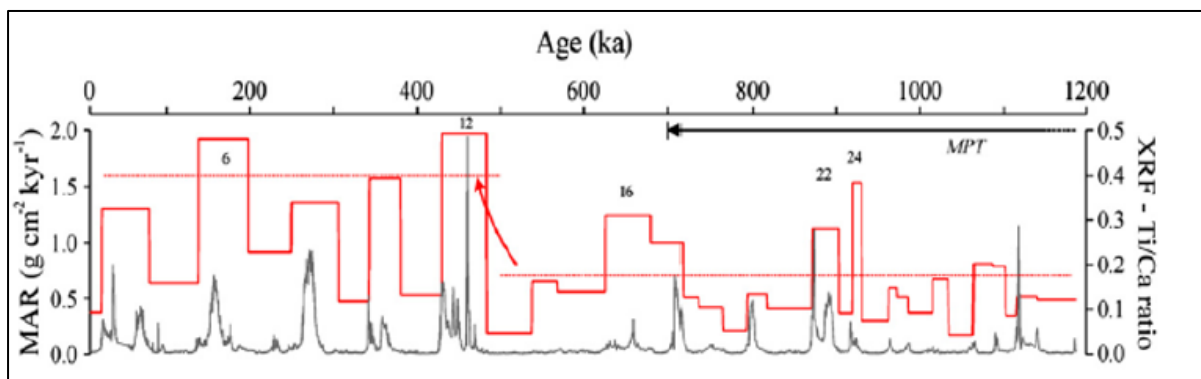


Figure 2.14: Peaks at MIS 12 and 6 indicate two episodes of increase in the terrigenous materials in the cores located at the Bay of Biscay, at the mouth of the English Channel. MAR (mass of terrigenous elements) is indicated in red while Ti/Ca ratio in black. The two parameters are strictly related as their increase generally corresponds to low-stand post glacial time with abundance of terrigenous material during climate cold phases. From Toucanne et al. (2009b).

Toucanne et al. (2009b) give evidence of two peaks in the mass of terrigenous elements (Fig. 2.8): the first at MIS 12 and the following at MIS 6. Toucanne et al. (2009b) do not interpret the second increase in terrigenous mass as caused by a catastrophic event but as consequence of increased volume of water running through the English Channel due to ice-sheet melting and normal fluvial drainage across the Strait. Nevertheless, other authors support the hypothesis of a second catastrophic event breaching the Strait at MIS 6 and fully connecting the southern North Sea to the Atlantic Ocean (Gibbard, 1995, 2000; Gupta et al., 2007; Busschers et al., 2008; Meinsen et al., 2011).

In addition to geochemical data, numerous palaeontological studies have been conducted over the last twenty years, in order to reconstruct the patterns and oscillations in the records of human presence (and absence) in ancient Britain during Pleistocene (Waitt,

1985; Gamble, 1987; Roebroeks et al., 1992; Roebroeks, 2001; Ashton and Lewis, 2002; White et al., 2006; Ashton and Hosfield, 2010; Ashton and Lewis, 2012). Patterns of human colonization and settlements in Southern Britain are strictly related to phases of insularity/connection to the Continent. Records from primary context sites and archives of stone artefacts available in river terraces (Bridgland et al., 2001; Bridgland and Schreve, 2001; Maddy et al., 2001; Bridgland et al., 2006; Westaway et al., 2006) are used to reconstruct the chronology and variation in human appearance and changes in artefact technologies and density. The variation of these factors during Lower to Middle Palaeolithic, can provide key indications and constrains on the changes in Britain palaeo-geography, times of insularity and periods of marine conditions at the Dover Strait (Ashton and Hosfield, 2010).

Palaeontological data show that changes in the density of human artefacts found in river terraces can be used as proxy for human population in Britain (Ashton and Hosfield, 2010). These data indicate a decline in the number of inhabitants from MIS 11 to MIS 8 and strong decrease or absence of population between MIS 6 and 4 (Fig. 2.15). The first decline event has received less support, with different interpretations proposed in order to explain the decrease in human artefacts. Ashton and Lewis (2012) suggest that the decrease in the number of artefacts can be used as proxy for population, indicating a possible migration of Neanderthal population from Britain to Eastern Europe. Other authors (Hosfield, 2005; White et al., 2006; McNabb, 2007) do not support this interpretation, explaining the decrease in artefacts as due to the variation of factors such as regional differences in archaeology, effects of collecting history or changes in technology and landscapes. On the other hand, wide agreement is found on the decline of human inhabitants between MIS 6 and 4 (Stuart, 1976; Currant, 1986; Wymer, 1988; Currant and Jacobi, 2001; Ashton, 2002; Ashton and Lewis, 2002). These two episodes of decline in human population would be consistent with the proposed breaching of the Strait at MIS 12 and 6 as the breaching of the Strait would have led to the isolation of Britain from Europe, preventing human migration from the continent across the land-bridge of the Wald-Artois anticline at the Dover Strait. Despite this, further analyses are required to constrain a more precise timing of human migration, in order to test its full consistency with the reconstructed palaeo-geography of north-western Europe.

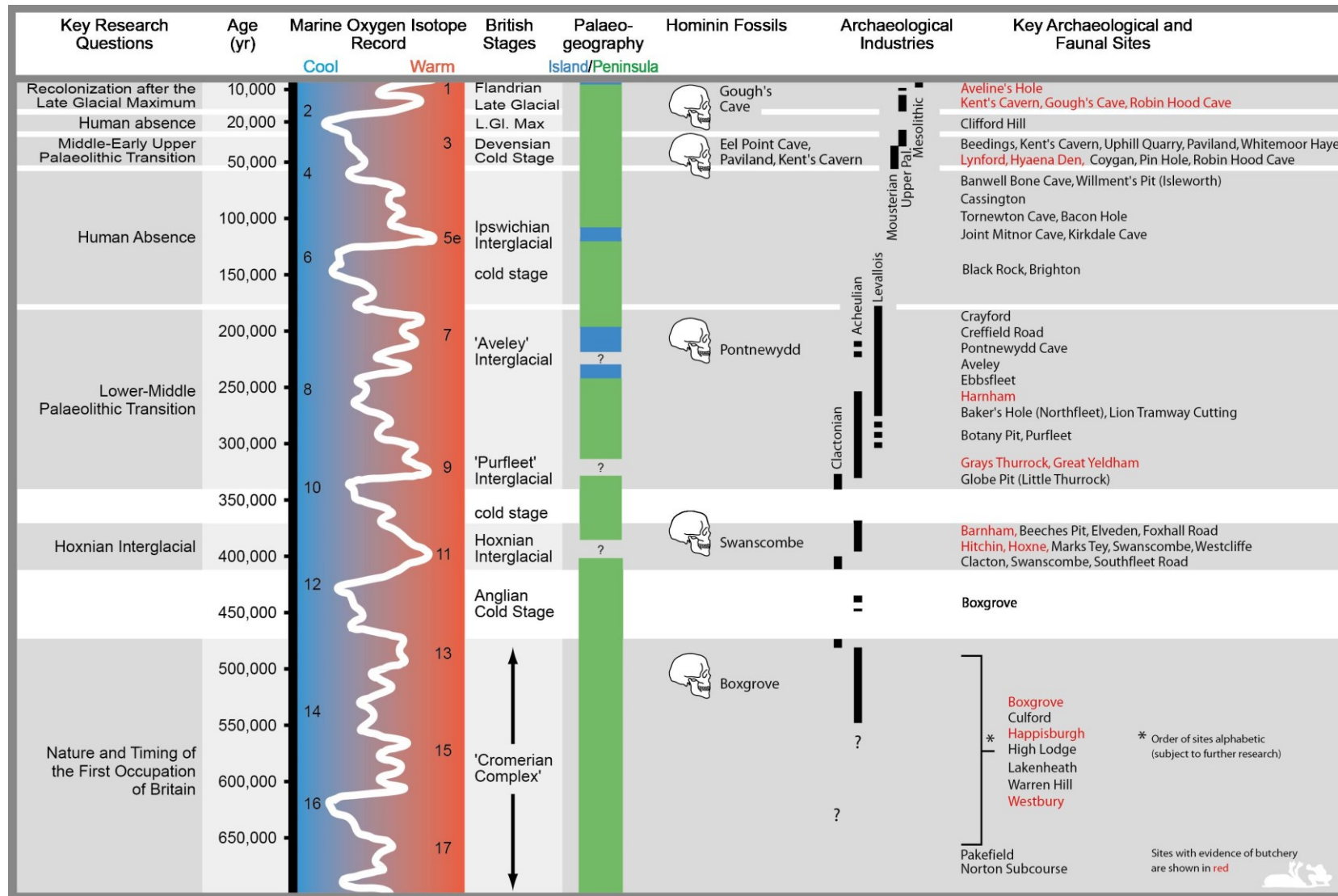


Figure 2.15: Chart from Ancient Human Occupation of Britain Project (AHOB Project) chart illustrating the oscillations in human occupation of ancient Britain in relation to the sea level and climate change, represented by the Marine Oxygen Isotope record from AHOB Publications (<http://www.ahobproject.org>).

2.5 Seismicity of the English Channel

A brief review and analysis of the Seismicity of Britain has been carried out in this study with the aim of understanding if a seismic event could have represented a possible trigger for the proposed failure of the pro-glacial lake bedrock dam and consequent breaching of the Dover Strait.

Despite Britain being considered a low seismically active area, a number of events that took place in the past hundred years in Great Britain have been a source of risk. The catalogue of the seismicity of the British Isles (Fig. 2.16) illustrates the major seismic events in the UK, defined as events with a magnitude equal or greater than 3 M_L (Musson, 1994). A total of 502 events are contained in the catalogue and are subdivided in three groups. The first group is composed of pre-1700 events; all the events are assumed to have magnitude greater than 4 M_L . Because of the lack of instrumentation at that time this section is probably incomplete. The second group is composed of the historical earthquakes recorded between 1700 and 1970, all having a magnitude between 3 and 4 M_L . The last and most complete group is represented by those earthquakes that occurred after 1970 and that have a magnitude greater than 3 M_L , and greater than 2.5 M_L from 1981 onwards.

The distribution of British earthquakes in time indicates an average recurrence time of 1 year for 3.7 M_L earthquakes and of 100 years for 5.6 M_L , with a maximum magnitude of 6.2 M_L expected for the UK. The geographical distribution of the earthquakes shown in the BGS catalogue (Fig. 2.16) is non-random. In particular, the western margin of the UK shows a higher concentration of events, together with the offshore area in the North Sea and Strait of Dover. The higher occurrence of events in this area indicates that the English Channel area represents a key area has had two of the largest earthquakes ever occurred in Britain took place in this area, respectively in 1382 and 1580 (Melville et al., 1996, 1994; Musson, 2004, 2007, 2012; Baptie, 2007).

2.5.1 English Channel earthquakes

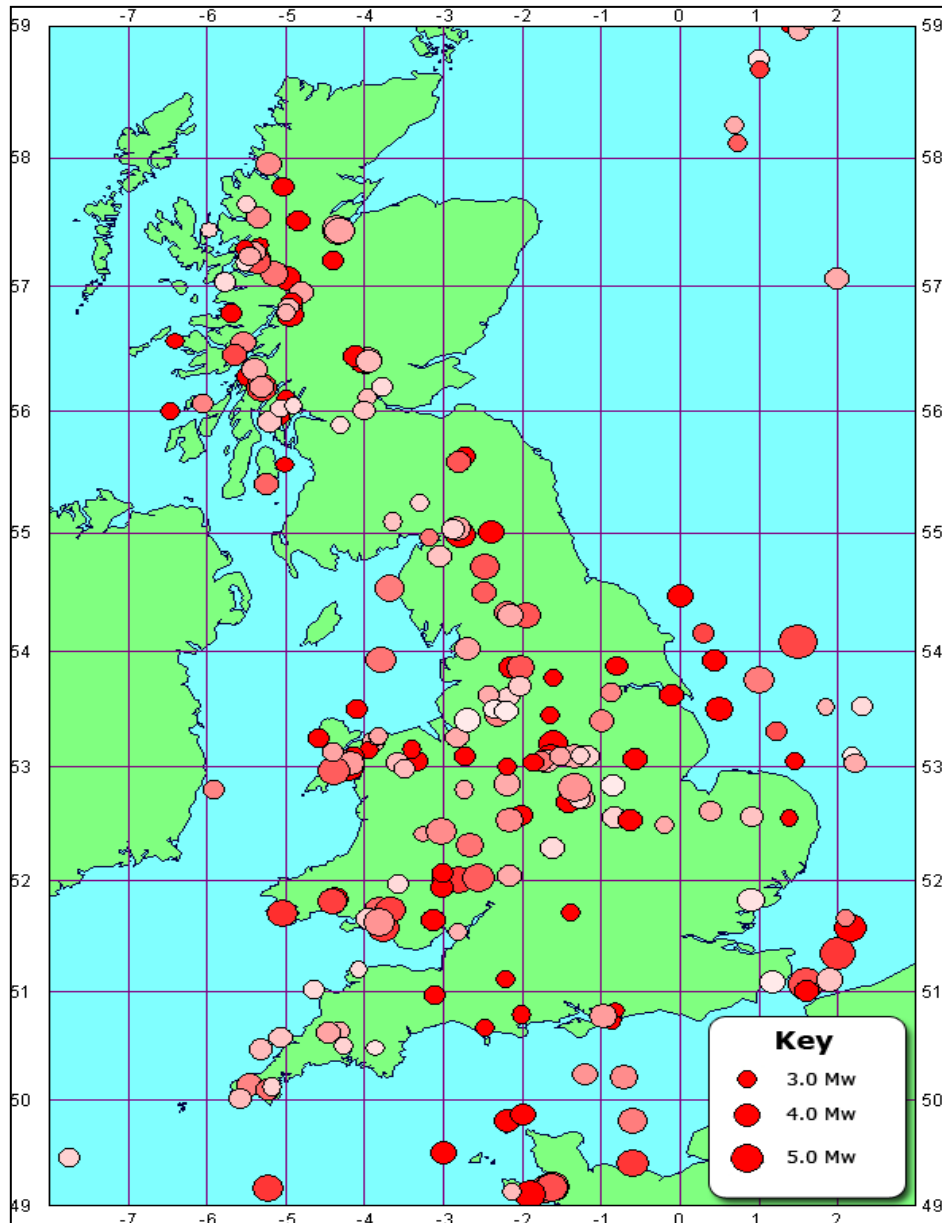


Figure 2.16: Seismicity of the UK from 1700 to 1990 (all events ≥ 3.0 Mw). Lighter colouring indicates shallow events from light pink, shallow (<15 km) to red, deep events (>15 km). From Musson, 1994.

The first of the two main English Channel earthquakes was recorded on the 3rd of May 1382 at 3 pm in an area that extended from Kent to the Flanders. The epicentre location is shown in Fig. 2.17, although this is a speculative interpretation as it is based on the felt quake effects in different areas. The Mercalli-scale intensity of the event was about VIII in England and VI in

Belgium, with an equivalent estimated magnitude of 5.5-6. The possible radius of isoseismal VII is approximately 70 km wide, with a focal depth of 25 to 30 km (Varley, 1996).

The 1580 earthquake felt an intensity of up to VIII was registered in Dover while the worst affected town in France was Calais, on the other side of the Channel. The effects related to the earthquake were reported in areas that are located very far from the assumed epicentre, identified in the eastern English Channel (the radius of perception of the event is estimated to be about 550 km). This implies a deep focus at a depth of 25 km, probably located in the lower crust, and connected to the reactivation of older faults (Neilson et al., 1984).

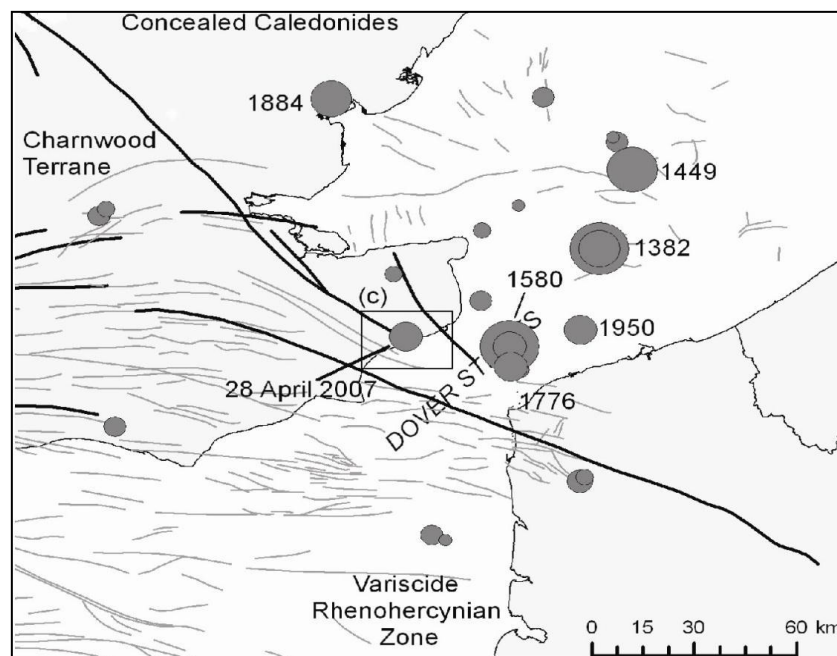


Figure 2.17: Fault map (main structures indicated by bold black lines, minor faults indicated by thin grey lines) of south-eastern England and Dover Strait area, showing with the proposed location of the epicentre of the major earthquakes that affected the area (from Musson, 2007).

Among the post 1700 events, two earthquakes affected the Dover Strait. A first quake affected the Dover Strait in 1776 on the 28th of November, with an intensity of V-VI recorded in Dover (Melville et al., 1996). The second and most recent event, although characterised by a lower magnitude with respect to the older quakes, was registered in the Folkestone area on the 28th of April 2007. This earthquake was the strongest event occurred in recent times in the UK; the event had a magnitude of 4.3 M_L (Ottemoeller et al., 2009) with a proposed focal

mechanism associated to a N-NW-S-SE striking strike-slip fault plane with normal component, possibly associated with a fault of the Variscan front (Sargeant et al., 2008).

In relation to the distribution of the described major earthquakes to the tectonic terranes (Musson, 1994, 1997) boundary, the high-level of seismic activity in the Southern North Sea area can be explained by to the presence of the Viking Graben area in the North Sea, showing high seismic activity, together with the Witch Ground Graben located to the south. As shown in Fig. 2.17, the epicentres of the Southern North Sea quakes are all located in the Strait terrane (Musson and Sargeant, 2007), while the Folkestone earthquake is located in the Variscan Zone on the south of the Variscan front that separates the two terranes.

2.5.2 Implications

The distribution of earthquakes in Britain indicates that the Dover Strait and the Southern North Sea represent the most seismically active locations in southern Britain, representing the area where two of the strongest earthquakes ever occurred in the UK took place (1382 and 1850 quakes). Applying the calculated average recurrence times to the past thousands year event record, assuming the main stress fields remained the same, the results indicate that events such as the 1382 earthquake have the potential to occur every 100 years (Musson, 1994). It is therefore possible that events similar to the Dover Straits earthquake could have happened in the same area multiple times in the past. An earthquake is one of the possible triggers of the pro-glacial lake dam failure that cause the drainage of the pro-glacial lake located in the Southern North Sea.

2.6 Summary

As a summary of the literature review proposed in this chapter, four main points can be drawn:

- 1) Although megafloods have been intensely studied in the last decades, many aspects of these global-scale events are still poorly constrained, as megaflood low frequency prevents direct observation and measurement. Further sedimentological, geomorphologic and hydraulic analyses are therefore required in order to better understand the mechanisms and consequences of megaflood events on Earth. The interpretation and reconstruction of the English Channel megaflood contributes to a deeper understanding of catastrophic processes and of the elements that characterise them, analysing the analogies and differences of the English Channel flood terrain in respect to those reviewed in literature. This includes the reconstruction of the source of the flood, represented by the suggested pro-glacial lake and bedrock dam located at the Dover Strait, and their relationship to the downstream portion of the flood terrain.
- 2) Accurate analyses are required for the reconstruction of the catastrophic flows that carved the Northern Palaeovalley, carving the flood-related features identified by Gupta et al. (2007). In particular, a detailed comparison to other megaflood terrains on Earth and to the typical megaflood-related feature associations is required, in order to test the catastrophic origin of the Northern Palaeovalley and its associated features. This also makes it possible to estimate the flood magnitude and calculate palaeo-hydraulic parameters for the English Channel megaflood and compare them to the Pleistocene flood events on Earth.
- 3) The Quaternary palaeo-geography of north-western Europe, and in particular of the Dover Strait area, requires further and more detailed investigation, particularly in relation to the relative and absolute timing of the events that led to the isolation of Britain from Europe. The analysis of the flood spill point, located at the Dover Strait, is therefore crucial in reconstructing the flood history and contributing to constrain the palaeo-geographical evolution of north-western Europe during Pleistocene.
- 4) To date, the available marine geophysical data and scientific research undertaken in the Dover Strait area are almost limited to the Channel Tunnel geo-engineering

investigation and construction works (Appendix A). The availability of regional-scale high-resolution datasets throughout the eastern and central English Channel allows for a re-evaluation of the geological characterisation of the area and in the understanding of its influence on the flood-related features present at the seabed. Importantly, this also allows for the distinction of flood-related features from fluvial and tidal bedforms previously mapped in the English Channel (see Section 2.3) whose origin has been widely debated.

3 . Data and Methods

3.1 Introduction

This chapter describes the dataset used in this work, focusing on the spatial extent and main research targets that were set for each dataset during the analyses. Details on the acquisition, processing and editing and interpretation phases of each dataset that was used for the analyses carried out in this thesis are provided.

In particular, the data used in this work (Fig. 3.1) is composed of two types of datasets: high resolution reflection seismic data and bathymetry. The chapter is divided in two sections: the part first illustrates the methods for the editing and analysis of the available seismic, while the second part of the chapter focuses on the bathymetric datasets.

The section illustrating the three seismic reflection datasets describes the acquisition parameters of each of them, the editing phase and, in particular, the different analyses that were performed on each dataset depending on their location, extent and quality. This section also describes the calibration of the seismic, detailing the available wells and logs that were used.

The second part of the chapter focuses on the bathymetric data, starting with an overview on the different type of bathymetric data, their resolution and extent. A description of the acquisition, editing and analysis phase of each bathymetric dataset is provided. In particular, for Maritime and Coastguard Agency (MCA) MBES data, a detailed discussion of the quality and standard of the data, together with its main advantages and limitations, is presented. The last section of the chapter details the gridding of the merged DTMs, illustrating which datasets were used and the techniques that were used for gridding the merged dataset. Finally, a comparison between the existing datasets and the ones gridded for this work is discussed, showing the main improvements and limitation of the gridded data.

3.2 Seismic reflection data

Three seismic reflection datasets were analysed in this work (Table 3.1): an analogue dataset collected for the Channel Tunnel investigations, a boomer seismic dataset collected in the

central English Channel and a digital sparker survey located at the Dover Strait. A description of the acquisition and analysis process of the datasets is given in this section.

Name of the survey	Colour on map (Fig. 3.1)	Details	Main Targets
4 1986-88 5 Analogue reflection seismic data (Channel Tunnel seismic)	Green	Shallow target seismic – stack and migrated data available	Calibration of Dangeard I 2002 seismic Constrain the extension of the Fosse Dangeard
	Pink	Shallow target seismic – line drawings available	
	Black	Deep target seismic – stack data available	
2005 MALSF seismic (Central Channel seismic)	Black	Segy unmigrated data available	Seismo-stratigraphic analyses Megaflood-related bedform analysis
Dangeard I 2002	Red	Segy unmigrated data available	6 Seismo-stratigraphic analyses Fosse Dangeard analysis

Table 3.1: Summary: Summary of the available datasets, their format, target and colour in the map in Fig. 3.1.

3.2.1 Analogue reflection seismic

Data acquisition

Extensive efforts were made in order to compile all the Channel Tunnel related seismic, collected pre 1986. Unfortunately the dataset used by Destombes et al. in their 1975 study is assumed lost (D. Blundell, pers. comm.). Despite this, data from the final campaign, here called “Channel Tunnel seismic data” (Figs. 3.1-3.3) were kindly supplied by Mr. John Arthur, formerly of John Arthur Associated who were subcontracted to interpret the seismic data. The data were collected in 1986-88 for geotechnical and geophysical studies for the

English Channel Tunnel project by TML (TransManche Ltd.). The supplied material was accompanied by detailed acquisition, processing and interpretation reports and key acquisition parameters (Table 3.2). During acquisition, lines were divided in “shallow target” and “deep target”: their acquisition parameters are summarised in Table 3.2.

Shallow target lines are numbered according to KP (“cable point”) position, mean distance from Dover, expressed in km, and were collected every 250 m.

	Shallow target	Deep target
n. of channels	24	48
record length	500 msec	1000 msec
sample rate	0.25 msec	0.5 msec
band pass low cut	750 Hz	18 Hz
band pass high cut	1200 Hz	60 Hz
streamer length	150 m	600 m

Table 3.2: Acquisition parameters of the Channel Tunnel seismic data.

Data analysis

As the seismic lines were available on paper only, the lines were hand digitised from a paper chart and imported into an ArcGIS database. The recorded sections were then scanned and loaded into Fledermaus v.7 to aid the interpretation of the gridded bathymetry.

The analysis of the Channel Tunnel dataset had two main purposes: the calibration of the high-resolution reflection seismic and a regional-scale analysis of the geology of the Dover Strait. The correlation of the main seismic reflectors present in the seismic datasets was made between the data interpreted by J. Arthur during the Channel Tunnel investigation and the new uninterpreted high-resolution data, defining a detailed seismic stratigraphy of the area (described in Chapter 4 of this work) as the Tunnel investigation had provided very high precision interpretation and positioning of the main geologic units (Fig. 3.3).

The main target of the analysis of the geological structures present in this area were the bedrock incisions mapped by Destombes et al. (1975) during Channel Tunnel investigations

and identified as Fosse Dangeard. As Destombes' dataset is no longer available, it is not possible to infer whether the area adjacent to the Fosse Dangeard has been surveyed in 1975 study or not. However, the same bedrock incisions are visible in the Channel Tunnel reflection seismic (Fig. 3.3) that was used to define the area of the Dover Strait where the incisions are present, as it covers a wider area throughout the Strait in comparison to the high-resolution Dangeard I 2002 reflection seismic which is restricted to the French sector. All seismic reflection lines available for this project and located across the Dover Strait were inspected with the aim of constraining the lateral extension of the Fosse Dangeard throughout the dataset.

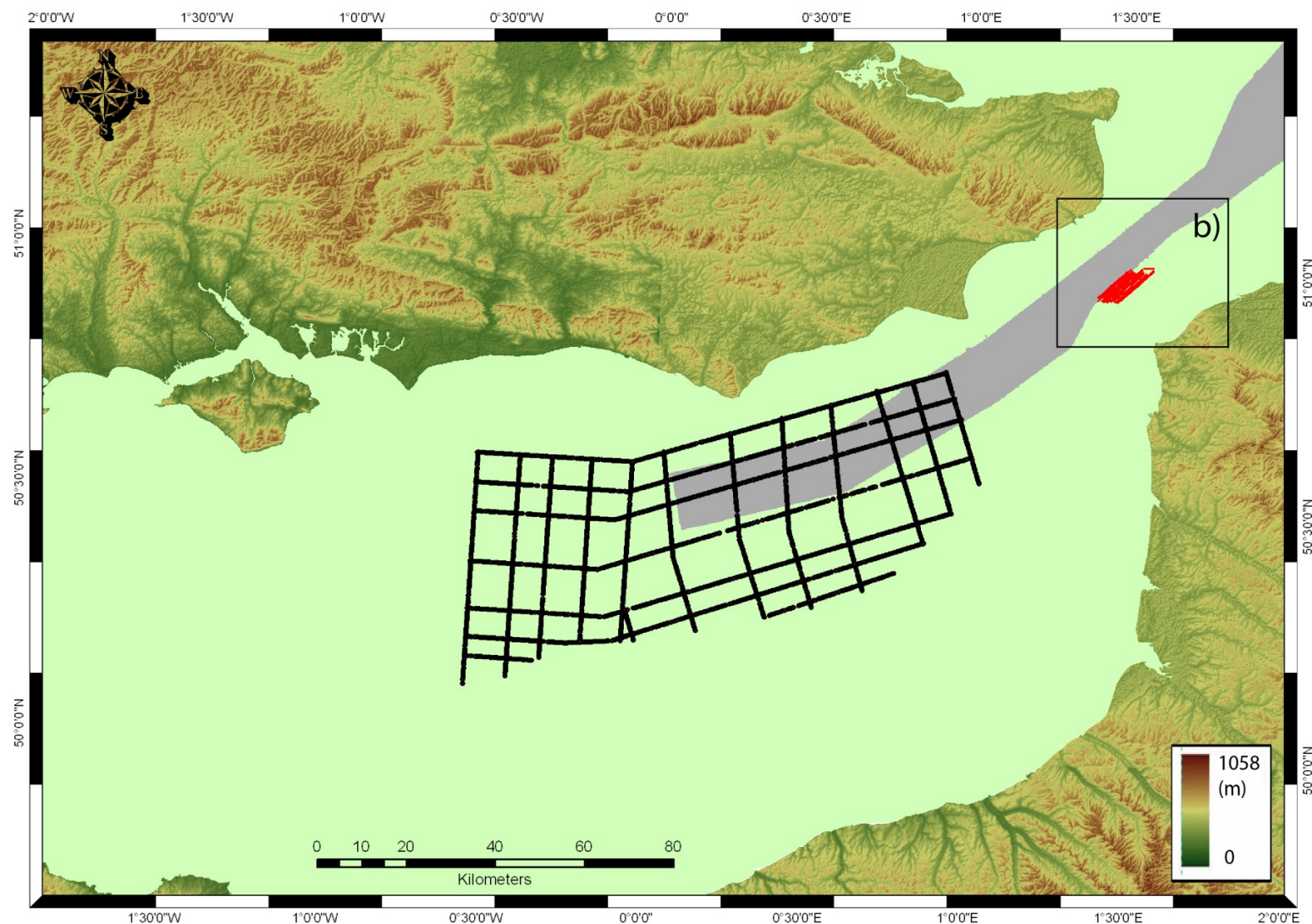


Figure 3.1: Map of the available datasets used in this work. The grey polygon represents the area covered by the high-resolution MBES data, the green area in the background represents the area covered by the merged bathymetry. The seismic lines are represented by the Central Channel dataset (black lines) and by the seismic datasets at the Dover Strait (in red, see Fig. 3.1b).

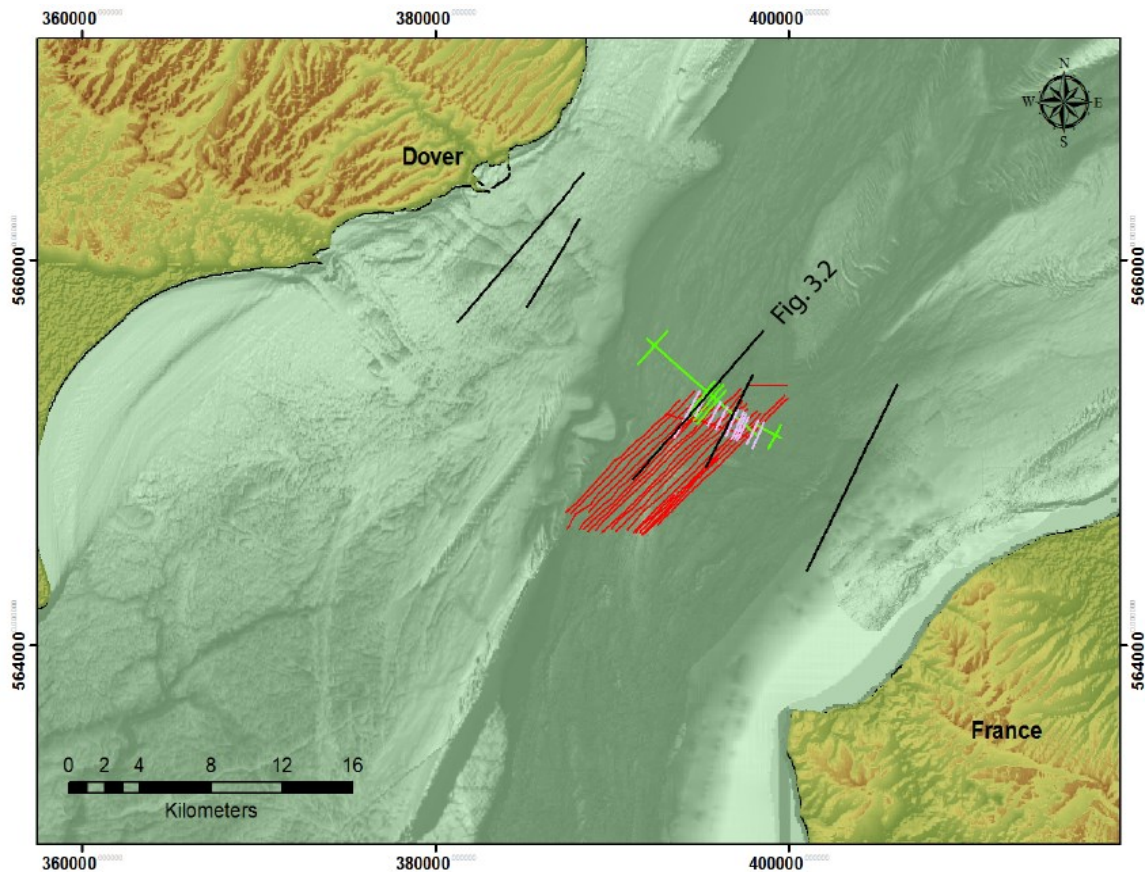


Figure 3.1b: The coloured lines represent the employed seismic reflection dataset: the Dangeard I 2002 Seismic lines (red lines) and the Channel Tunnel lines with three different colours. In particular, the Channel dataset is represented by green “shallow target” (stack and migrated available paper record), and pink “shallow target” lines (line drawing available). The black lines are “deep target data”. Bathymetric data are displayed in the background.

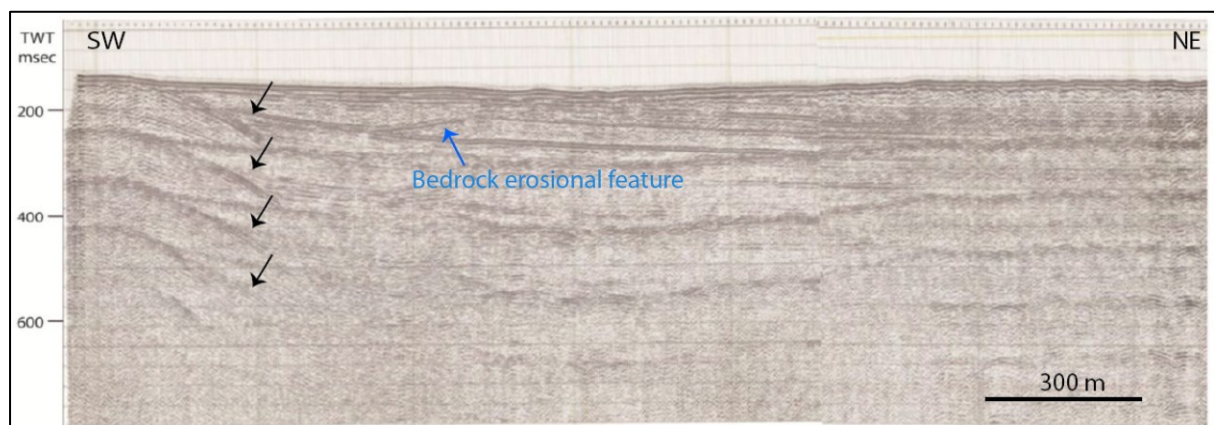


Figure 3.2: Example of one of the analogue seismic lines (Line KP16.25) from the Channel Tunnel dataset. The data is covered by the seabed multiple at a depth of 200 msec TWT (indicated by the black arrows). Vertical exaggeration (VE) is 1.5. For location of the line see Fig. 3.1.

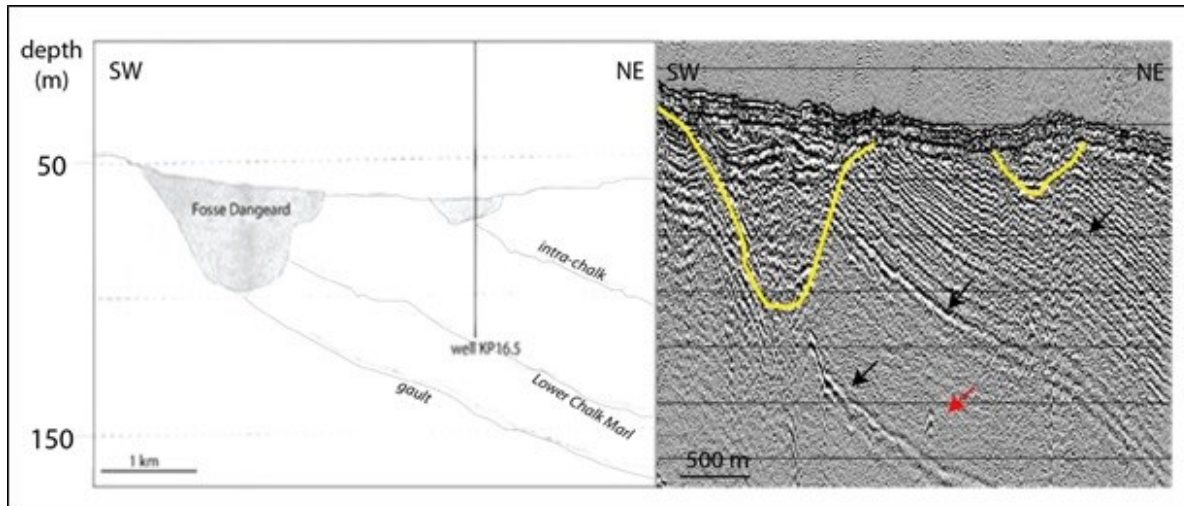


Figure 3.3: Example of line drawing of line KP 16.5 from Tunnel dataset (left) and high-resolution seismic line (right). The black arrows in the image on the right indicate the reflectors labelled in the line drawing (left) while the red arrow (right) indicates the diffraction hyperbolae of the Tunnel itself. The yellow lines (right) represent the basal surface of the two incisions carved into the bedrock reflectors (respectively 35 and 5 m deep) and represented by the grey area in the line drawing in the left image.

3.2.2 Dangeard I 2002 Digital reflection seismic

Data acquisition

The high-resolution seismic data, also called Dangeard I 2002 reflection seismic in this work (Fig. 3.1), was collected in 2002 onboard of the Sepia II (CNRS/INSU) vessel with 1 kHz frequency sparker (Centipede) with source energy of 300 Joules and 1 s shooting interval. For positioning, dGPS was used. DGPS uses a network of fixed, ground-based reference stations to calculate the difference between the positions indicated by the satellite systems and the known positions. The stations compute the difference between the measured satellite ranges and actual ranges. DGPS accuracy ranges from the 15-meter nominal GPS accuracy to about 10 cm. In the case of the English Channel, accuracy is better than 1 m. The data consist of 19 parallel, ~10 km long lines with ~200 m spacing and the acquisition software that was used is iXBlue DELPH SeismicSystem (A. Trentesaux, pers. comm.). The vertical penetration ranges from ~80 m up to ~150 m. Multiple seafloor reflections covers the primary signal.

Data analysis

The data were provided in a Kingdom Suite format; the SEG-Y files were successively exported in order to plot the data in GMT and Seismic Unix (Fig. 3.4) for the analysis and visualization of the data.

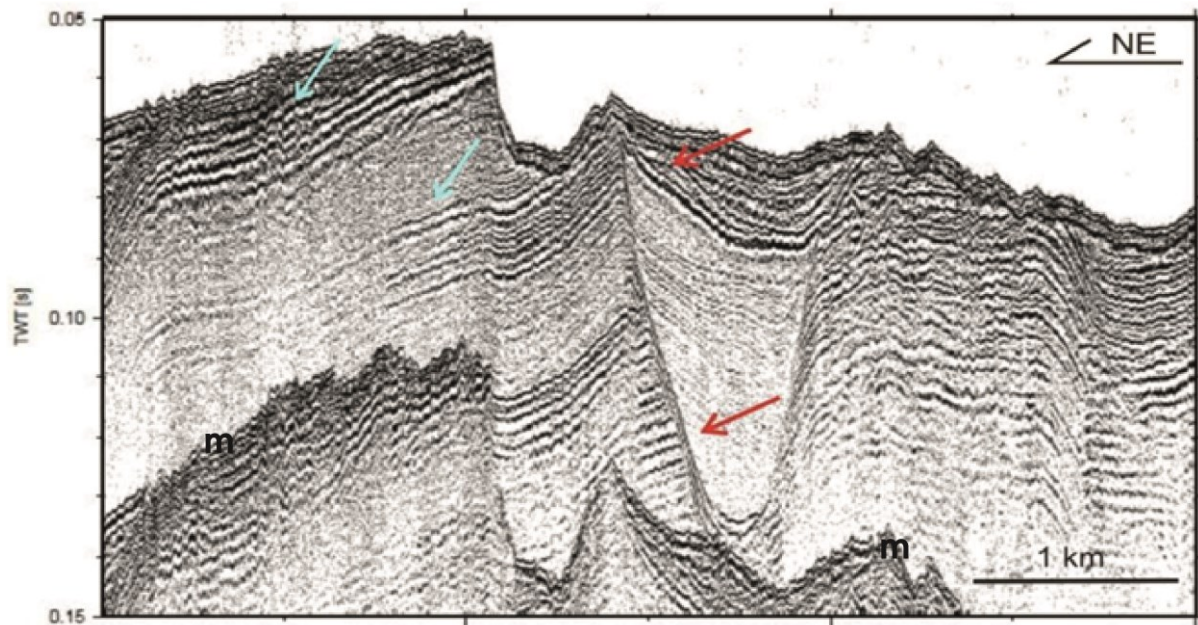


Figure 3.4: Line dan03 of the Dangeard I 2002 Seismic, illustrating the Fosse Dangeard, indicated by the red arrows. The blue arrows show strong bedrock reflectors; m indicates the seabed multiple. VE is 3 (with bedrock velocity of 2000 m/sec and incision sediment velocity of 1750 m/s).

Groups of reflections (Figs. 3.3 and 3.4) were interpreted, based on configuration, continuity, amplitude and frequency, in order to define different seismic facies (Mitchum and Vail, 1977) related to the main geological layers and basal erosive surfaces. Several seismic facies were identified and different erosional surfaces were correlated and interpreted in the area, including the erosional surface that bounds the Fosse Dangeard depressions. The bedrock stratigraphy was calibrated through the TML (TransManche Limited) wells drilled in 1986 with geotechnical purposes, providing a very detailed shallow stratigraphy of the area. The diffraction hyperbolae of the Tunnel were also used as a seismic marker of the Lower Chalk Marl (see Chapter 4) for the correlation of the data. As the high resolution data shiptrack lines are parallel to the Tunnel lines in the area where the wells are located (Fig. 3.5), the wells were hence overlapped on the new data, in order to define the position of the main horizons and correlate them to the seismic markers that were then interpreted through the whole database (Figs. 3.5 and 3.6).

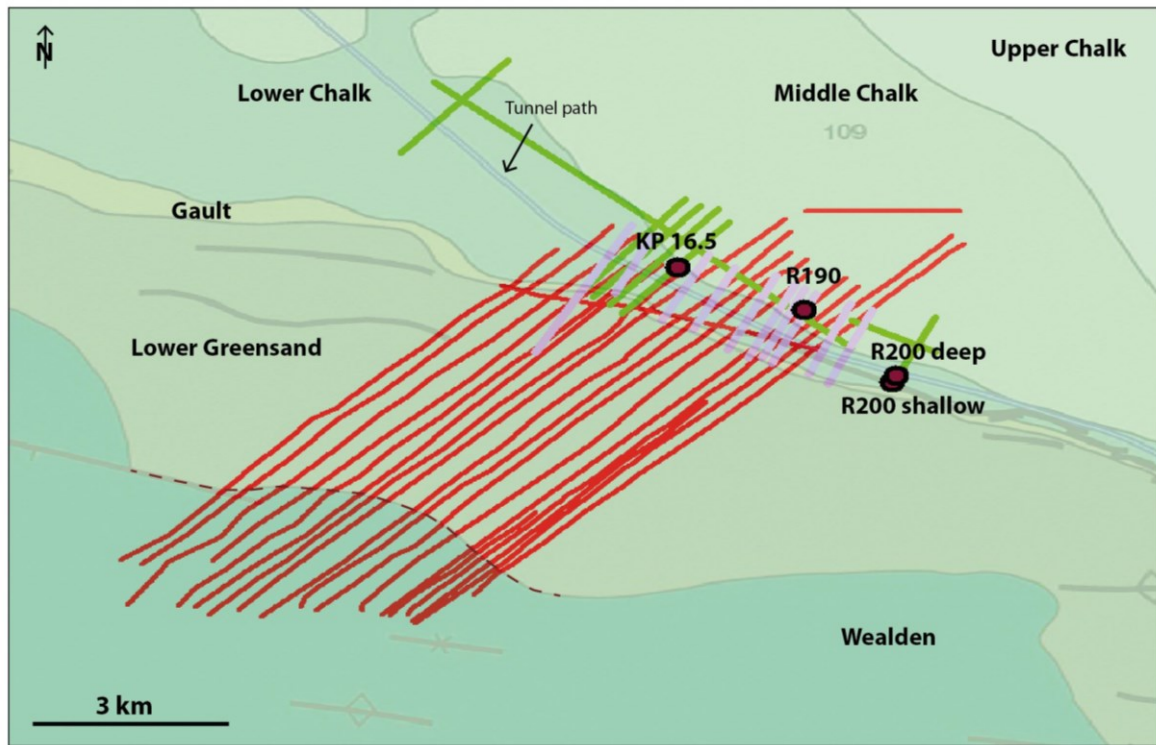


Figure 3.5: Location of the four available wells used for the seismic depth conversion and calibration, as shown in Fig. 3.6. The geology map (James et al., 2002) is displayed in the background. It was used to locate the Lower Greensand/Wealden contact (dashed line). The wells were all located in the NE area of the dataset and do not penetrate the Lower Greensand/Wealden contact.

In addition, in those areas where no wells were available, the position of the contact between the units of the bedrock stratigraphy was calibrated using the available geological maps, as in the case of the Lower Greensand/Wealden contact (Fig. 3.5). The correlation between depth and TWT was based on a velocity chart which was supplied in an acquisition report (J. Arthur, pers. comm.) together with the analogue data, listing the sediment and bedrock velocities for each line.

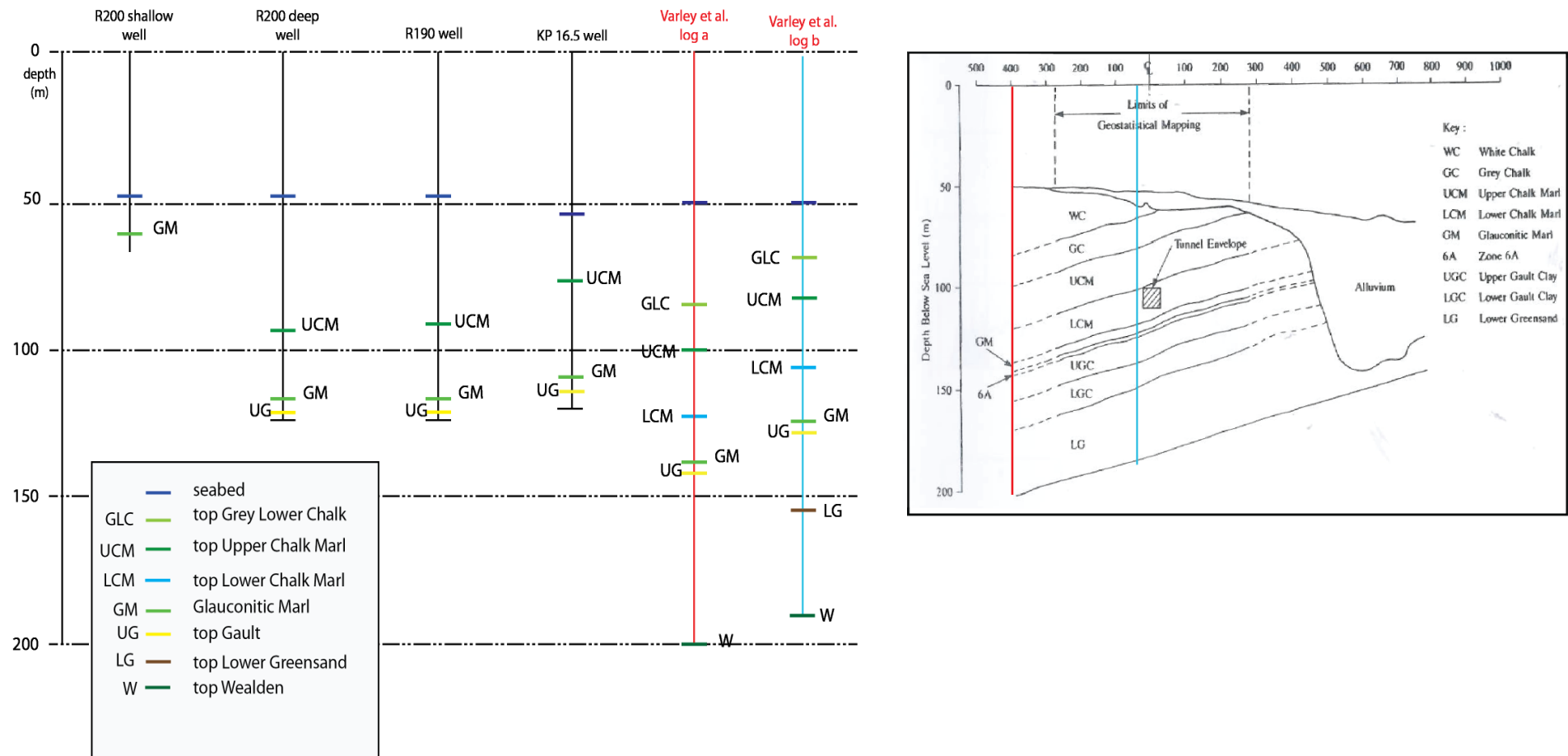


Figure 3.6: Summary of the depth of lithological contacts obtained from wells (black vertical lines) and sections (red and light blue) that were digitised from an interpreted line drawing of the Tunnel by Varley et al., 1996. The Middle Chalk and Wealden Beds were not penetrated by any wells.

Once all the main layers were interpreted, a contour map of the drawn horizons was made using TWT. This structural map of the interpreted depressions and the isopach map of the depression infill were also calculated from the difference between the TWT of the base of bedrock, the top of the infill and the TWT of the seabed. The erosive base of the Fosse Dangeard depressions was interpreted as the top of the bedrock; where this surface was covered by the multiple, or beyond record length, its depth was interpolated. The time to depth conversion was performed assuming a typical non-consolidated sediment velocity of 1750 m/s as proposed by Arthur et al. (1996), after log tests were performed for English Channel Tunnel geo-engineering investigations. The contours were generated using the 7.5 Kingdom Suite algorithm with grid cells of 1.

3.2.3 2005 MALSF Seismic

Data acquisition

The dataset was collected under the Geophysical and Multibeam Survey 2005/3_MEPPF: Eastern English Channel Marine Habitat Map Study. The survey took place in May and June 2005 in the Eastern English Channel aboard the M/V Tridens. This survey was undertaken for the Eastern English Channel Marine Habitat Map study (EECMHM) as part of a series of four surveys. The EECMHM study was funded by the Marine Environment Protection Fund (MEPF) a marine component of the Aggregate Levy Sustainability Fund (ALSF). The study area covers an extensive sea bed area of approximately 5,090 km² between Selsey Bill and Dungeness, out to the UK/France median line, centred on ten current aggregate mining licence application areas. Data were collected using a tow boomer and digitally recorded (<http://www.marinealsf.org.uk/>).

Data analysis

The xyz positioning of SEG Y files were processed in GMT in order to correct some navigation issues present in the seismic trace headers, deleting and correcting pikes and wrong positioning of data. The SEG Y files were then loaded into seismic Kingdom Suite project together with the Dangeard I 2002 Seismic reflection dataset. The Central Channel seismic reflection dataset is the largest dataset available in this work and allowed for a regional analysis of the flood features visible from bathymetry.

This dataset was used for three purposes:

1) Performing seismo-stratigraphic analyses of the bedrock succession and comparing the interpreted units to those interpreted in the Dover Strait area.

2) Validating the lithological boundaries, folds and faults that were previously mapped using the acoustic texture of the seabed from bathymetry. Lithological boundaries and tectonic features were mapped on this dataset through the seismo-stratigraphic analyses and the result was compared to the geologic map of the seabed that had been previously drawn using bathymetry.

3) Seismic analysis of flood bedforms present in the central channel area such as channels and bedrock remnants. In order to do so, the isopach map of the seabed sediments and of the palaeovalley infill were calculated on the entire dataset using Kingdom Suite, with the aim of distinguishing bedrock features from seabed and channel infill sediments. Three main reflectors were therefore targeted for this analysis: the seabed reflector, the top of the sand waves and the base of the channels (Fig. 3.7).

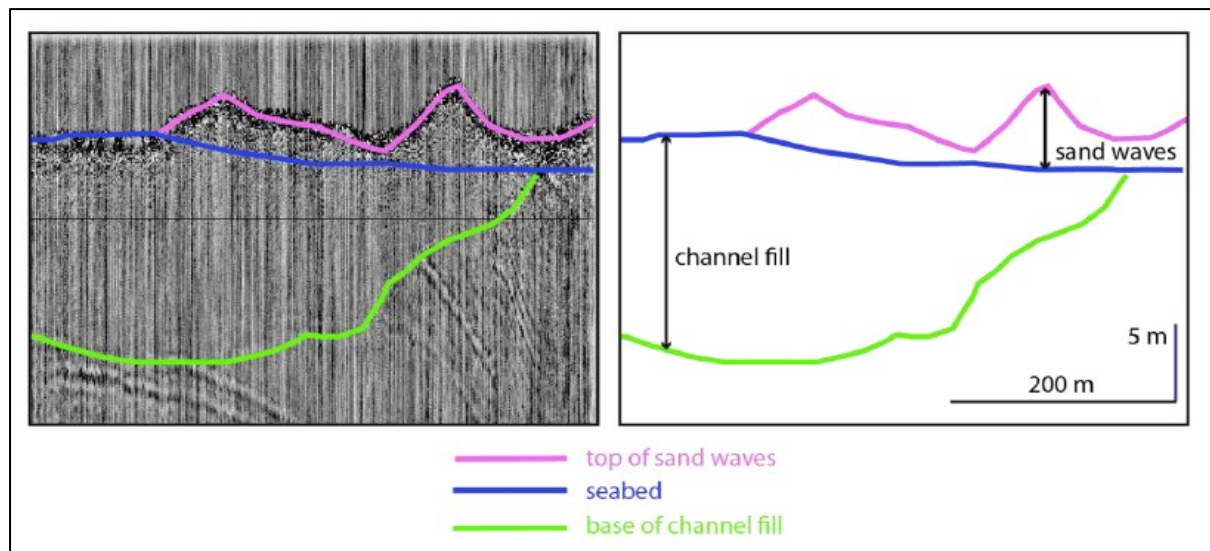


Figure 3.7: This figure illustrates the three main reflectors targeted for the analysis of the channel infill isopach map (channel base and seabed) and for the sand waves isopach map (seabed and top of sand waves). The blue reflector was considered as seabed as reference depth as the sand waves are mobile sediments whose position continuously varies with time.

3.3 Bathymetry data

Five main bathymetric datasets (Fig. 3.8) were used in this work to perform geomorphic analyses and create a final bathymetry merge, using all five datasets combined at the highest possible resolution. Table 3.3 summarizes the name and resolution of the available datasets.

Name of the survey	Type of data	Gridding bin size	Extent
UK Hydrographic Office UKHO English Channel	single beam + multibeam + seismic	30 m	Whole study area
General Bathymetric Chart of the Oceans GEBCO	single beam	80 m	Whole study area
Service Hydrographique et Océanographique de la Marine SHOM	single beam	80 and 40 m	Dover Strait – French sector
Royal Observatory of Belgium-Renard Centre of Marine Geology ROB-RCMG MBES	multibeam	5 m	Central Dover Strait – French Sector
Maritime and Coastguard Agency MCA MBES	multibeam	7 and 1.5 m	Dover Strait – British sector

Table 3.3: Summary of the available bathymetric datasets, indicating the extent and the resolution of the DTMs. For distinction between data provided as .xyz and as grid see text.

3.3.1 UKHO English Channel grid

This dataset is composed of merged bathymetric data, gridded at 30 m bin size (Fig. 3.9). It is the result of the integration of single beam, multibeam and digitised bathymetry from seismic data (on the French side, as no bathymetric data were available in this sector) created using the software IVS 3D Fledermaus by G. Potter of the UK Hydrographic Office (UKHO) in November 2009. The bathymetry in the British sector of the dataset was collected by the Maritime and Coastguard Agency and was archived at the United Kingdom Hydrographic Office. since 2004 (Fig. 3.9). The dataset had been previously edited, tide corrected and interpolated with the available depth points by G. Potter in order to complete a first merged DTM of the study area.

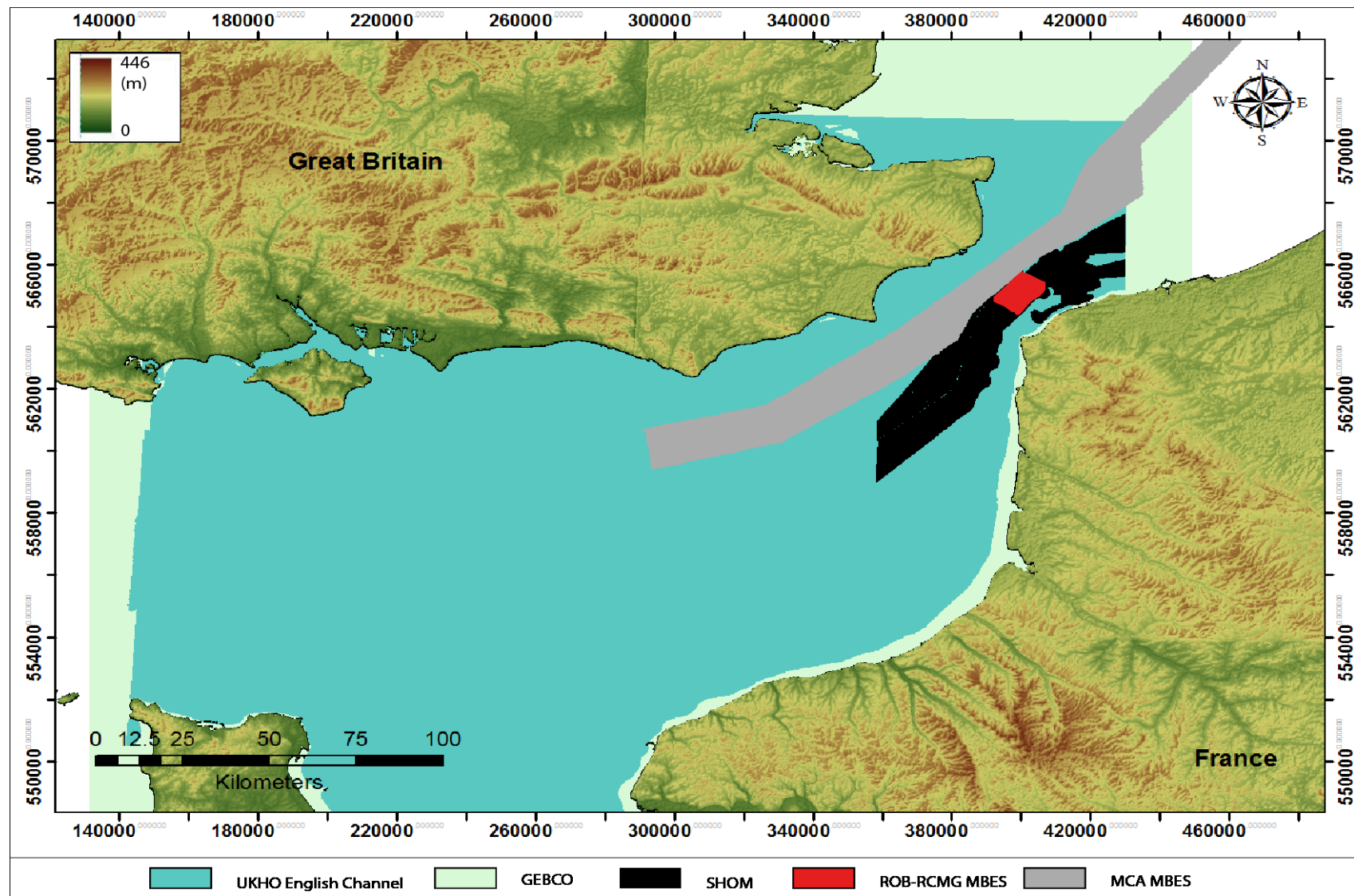


Figure 3.8: Map showing the extent of the bathymetric datasets available in this project, represented with different colours. Onshore topography from SRTM data (USGS, 2006), see Fig. 3.1 for colour-scale.

The grid includes both single beam data, collected since the 1980's, and modern MBES data collected. As the French sector was gridded from interpolated seismic data only (Fig. 3.9), the resolution of the data in this area is much lower and the interpretation of the seabed features was not possible from this dataset. For the gridding of the final merge DTM, in order to overcome the low resolution issue, the data was gridded together with the higher resolution SHOM dataset, located in the French Sector of the Dover Strait (see section 3.2.6 for details).

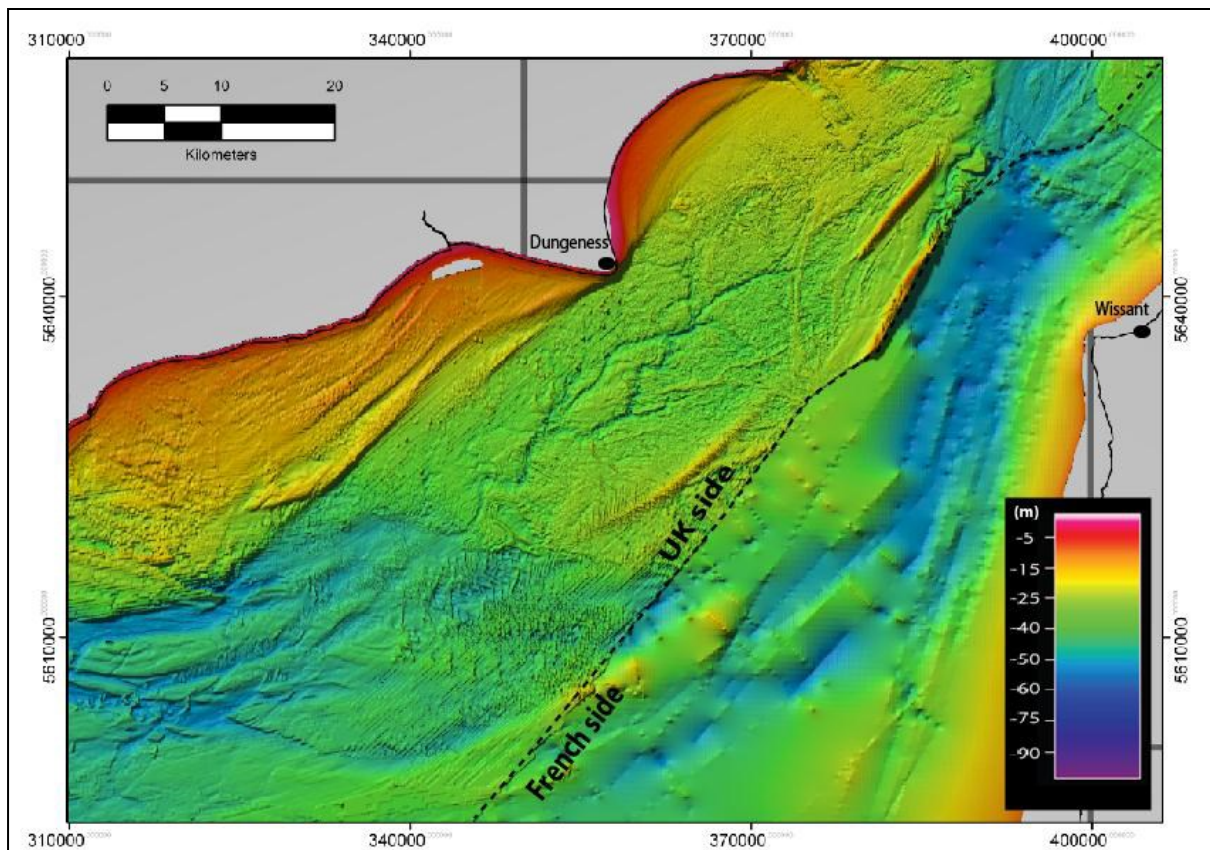


Figure 3.9: Extracted merged 30 m bin size DTM of the English Channel area from UKHO dataset, showing the UK and French border (dashed line). The border is visible on the DTM as it corresponds to a strong decrease of the resolution in the French sector, due to the reliance on interpolated depth points obtained from seismic data only.

3.3.2 GEBCO grid

The GEBCO (General Bathymetry Charts of the Ocean) grid used in this work consists of the GEBCO_08 Grid version 20100927, released in September 2010 (<http://www.gebco.net>). The dataset was gridded at 80 m bin size, corresponding to 30 arc-secs. As the grid covers the entire world surface, only the portion of the grid that corresponds to the area of study was extracted for this work, using GMT. The data points used to construct the GEBCO grid

consists in a compilation of ship-track depth soundings (although not valid for the case of the English Channel area, in some other areas the data are interpolated with soundings performed by satellite derived-gravity data using mean sea level datum). Because of the low resolution of this dataset, no geomorphic analyses were performed on this dataset that was only used as a base for the creation of the final merge, being the widest available dataset.

3.3.3 SHOM data

This dataset consists of a collection of single beam datasets collected by SHOM (Société Hydrographique et Oceanographique de la Marine) during the last forty years, covering the French area of the Dover Strait (Fig. 3.10). Data were gridded at the University of Lille, France, using IVS3D DMagic and Fledermaus v. 7, from two .xyz files provided by SHOM to the University of Lille (contract N° 205/2011).

As the z field values of the provided .xyz files were positive, the .xyz files were manually modified in Microsoft Excel in order to obtain a negative z value, consistent with depth. The density of the soundings in the dataset was highly variable, ranging from areas with high or complete coverage to zero or very poor sounding density. The data were plotted in order to check the sounding density per bin before gridding (Fig. 3.11): considering an 80 m bin, the number of soundings per bin varies from 1 to 10 decreasing to 1 to 4 for a 40 m bin. This is due to the dataset being the result of multiple single-beam datasets collected by SHOM at different times in a wide range of time and with different acquisition parameters. The two .xyz files were gridded at 80 and 40 m bin size in UTM Zone 31 Coordinate System, obtaining two DTMs that cover an area that is 140 km long and up to 45 km wide in its south-western portion, with a measurements that range from -86 to + 11 m. The positive value is due to the variable level of the tide during high tide levels as the reference level corresponds to low tide.

The DTMs show a variable coverage (3.12) due to the sounding density issue mentioned above. Two main types of artifacts also affect the dataset: i) lineations parallel to the ship tracklines and ii) artifacts that are due to the high mobility of seabed sediments. The artifacts parallel to the ship tracklines are visible in the dataset in different areas (Fig. 3.12): some areas show wide gaps and are only covered by data in those area where the ship tracklines are present, while some other areas have a complete data coverage but show strong tracklines lineations visible on the seabed. The artifacts follow multiple directions. The second type of artefact is caused by mobile bedforms, located in sand bank or dune areas of

the Strait. Dunes frequently display a multiple crest-line due to repeated surveying. This is due to bedform mobility between the surveys or even during a single survey, as the area is affected by strong tide and wind-driven processes (Le Bot and Trentesaux, 2004).

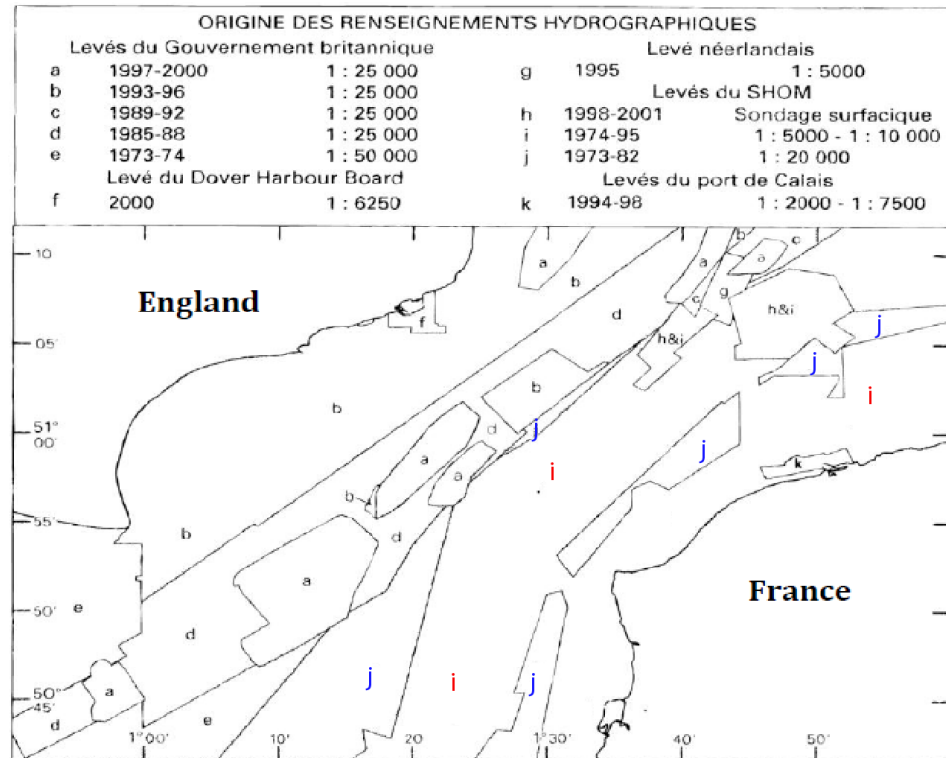


Figure 3.10: Map of the collected hydrographical data by SHOM indicating the survey dates: the data provided for this project is represented by letters *i*(red) and *j*(blue) (source of the map is SHOM database).

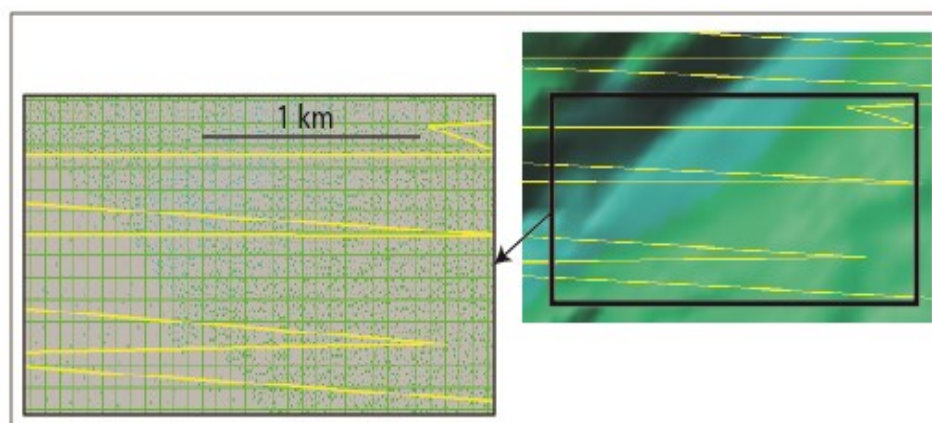


Figure 3.11: Variation in the number of soundings per bin: the left box shows the number of soundings (represented by the area on the right side, occupied by dots representing soundings) for each 80 m bin (represented by the green lines). The strong variation in the sounding number is not visible on the DTM (right box) as the data is gridded at a low resolution (80 m). The yellow lines represent the ship tracklines.

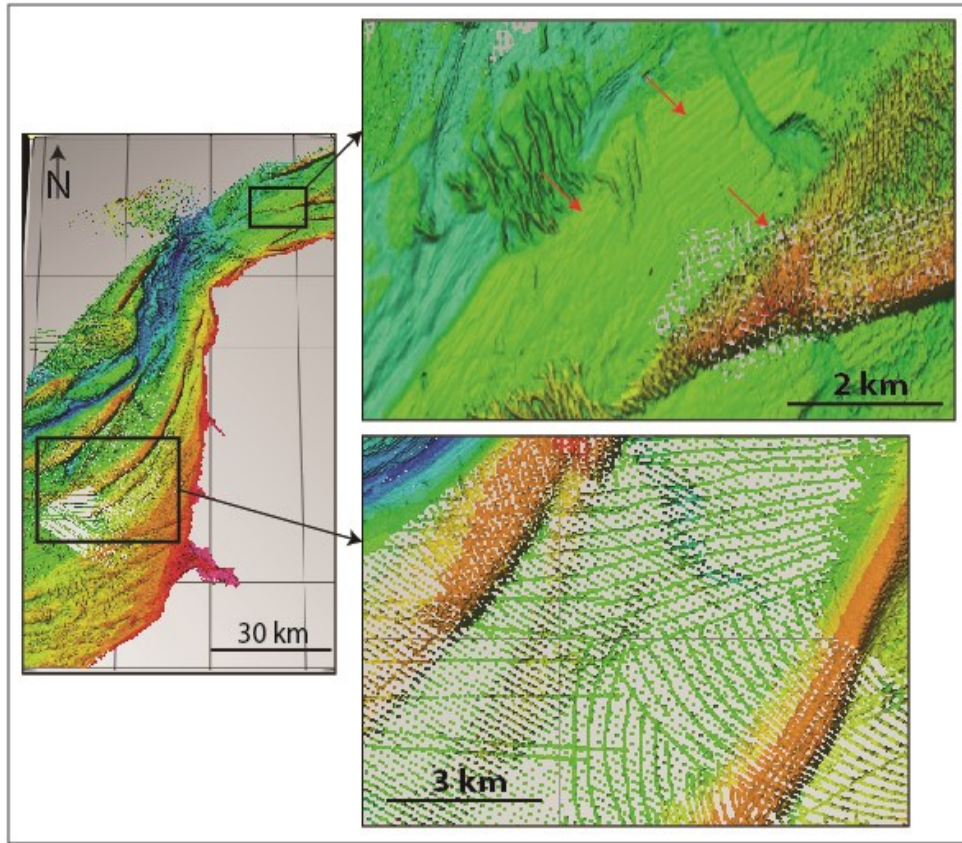


Figure 3.12: Artifacts parallel to the ship tracklines, visible throughout the 40 m dataset that displays areas of 100% coverage (upper box, lineations pointed by the red arrows) and area of very poor coverage where the data is present only along sets of parallel tracklines.

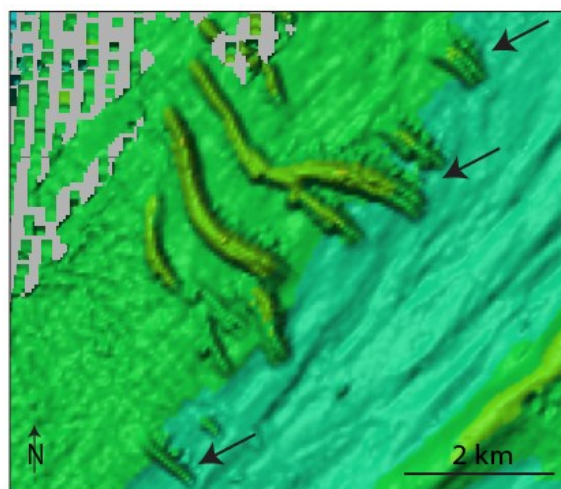


Figure 3.13: The black arrows point the irregular shape of the sand waves (~5 m of amplitude) that look truncated or show a double crest because of the high mobility of the seabed sediments and the time intervals between the acquisition of the survey lines.

3.3.4 ROB-RCMGBES

Data were provided by Thierry Camelbeeck of Royal Observatory of Belgium and Marc De Batist of Renard Centre of Marine Geology of University of Ghent. The Royal Observatory of Belgium-Renard Center of Marine Geology (Ghent University) (ROB-RCMG) dataset was collected in April 2010 on board of the R.V. Belgica, by the Belgian Management Unit of the North Sea Mathematical Models (MUMM). This data was collected using the Kongsberg EM3002 multi-beam echo-sounder from the Belgian Ministry of Economic Affairs, which is installed permanently on the vessel. Sound absorption coefficient was calculated in the survey area from the temperature, salinity and pH of the water, given by the ODAS-II system installed on the vessel. Based on the estimated sound-absorption coefficient and a sound profile that was taken with the SV-plus sound velocity probe on 15 April 2010 at 51.025°N, 1.51°E, an average sound velocity of 1483 m/s was selected for the water layer, calculated using the following parameters: sound absorption coefficient in water, calculated from temperature, salinity and pH of the surface water, given by the ODAS-II system. Data were collected with a 10-20% overlap between adjacent swaths. This resulted in an average line spacing of 200 m for an average depth of 40m. Due to bad weather conditions, all multibeam lines were shot in the direction of the waves (NE-SW) to minimize the influence of vessel roll. Data were successively processed by David García Moreno of Royal Observatory of Belgium and Renard Centre of Marine Geology using Sonar Scope software and provided as .xyz data to Imperial College. xyz data were then gridded at 5 m bin size and used for geomorphological analyses at the Dover Strait.

3.3.5 MCA MBES

The following paragraphs describe the acquisition and gridding phases of the high-resolution swath bathymetry dataset. This is the primary bathymetric dataset used in this work for the geomorphic analyses and interpretation of the flood-related features.

Data acquisition and filtering

All the details reported on the acquisition and data screening phase in the following section were extracted from the navigation report (provided by G. Potter, UKHO). The data were collected under contract to the Maritime and Coastguard Agency as part of their “Civil Hydrographic Programme”. The surveyed area is located in the HMOI 1159 Dover Strait Traffic Separation Scheme and was collected to meet the IHO order 1 standards (IHO s-4

publication, 2012). The survey area was subdivided in three main blocks: Block 1, Block 2 and Block 3 (Fig. 3.14) that are composed of smaller cells numbered from 001 (SW) to 175 (NE).

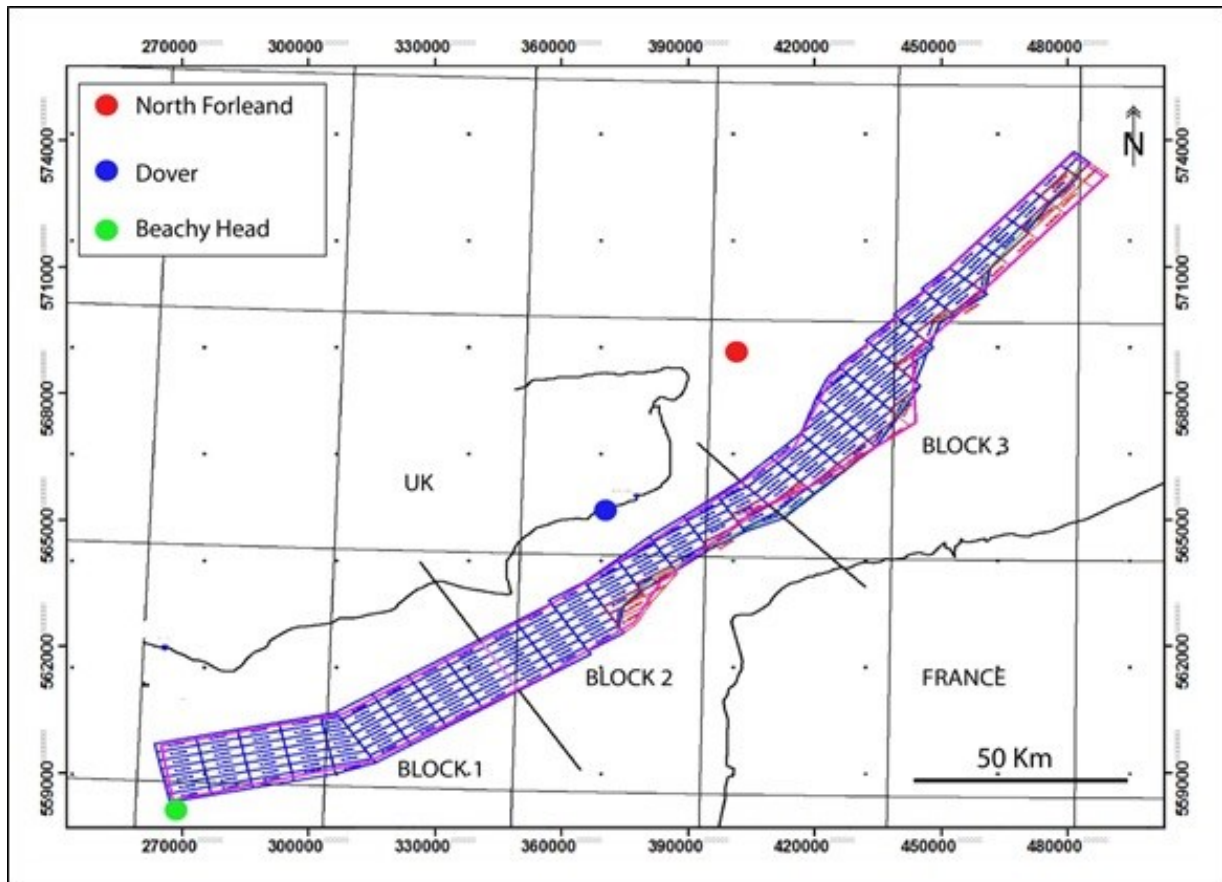


Figure 3.14: Overview of the MBES survey that is divided in three blocks, each subdivided in smaller cells. The three coloured dots represent the tide gauges employed for the survey (from navigation report), for details on Tidal corrections see Appendix A of this work.

Vessels and equipment

The swath bathymetry data were collected from the UK Hydrographical Office in 2006/2007 onboard of different vessels with different survey systems (3.15). In particular, four vessels were employed (Table 3.4 and Fig. 3.15): M/V Victor Hensen, M/V Geniusbank, M/V Meridian and M/V Jetstream with, respectively, the following MBESs operating onboard:

M/V Victor Hensen: The Kongsberg EM 710 MBES can be operated in shallow, medium, deep and very deep mode down to 2000 m water depth. The system operating frequency is a band of 60 kHz to 100k Hz. The system has a maximum ping rate of 25 Hz and 400

equiangular beams. The swath angle can be of up to 150° but was generally set to 130°. The swath width is up to 7 x the water depth.. The EM710 performed generally well.

M/V Meridien: Reson SeaBat 8101ER and SeaBat 7125 were mounted on this vessel. The ResonSeaBat 8101ER (extended range) is operating at frequency of 240 kHz. It uses 101 beams with a beam spacing of 1.5° resulting in a maximum swath of 150°. The beam width is 1.5° across and along track. The ResonSeaBat 7125 is configured as a dual head system. As this is a new system, only the 400 kHz frequency was available at the time of the survey. The system uses up to 512 beams per head. The beamwidth is 0.5° horizontally and 1° vertically. The estimated range with the 400 kHz is up to 300m (manufacturer specification).

Block	Vessels	Date
Block 1: Cell 001 – 063	M/V Victor Hensen	22.10.2006 – 16.04.2007
	M/V Meridian	
Block 2: Cell 064 – 111	M/V Geniusbank	29.06.2006 – 18.04.2007
	M/V Victor Hensen	
	M/V Meridian	
Block 3: Cell 113 – 175 (cell 112	M/V Geniusbank	09.08.2006 – 19.04.2007
does not exist)	M/V Jetstream	
	M/V Victor Hensen	
	M/V Meridian	

Table 3.4: Summary table of the MBES surveys indicating date and vessels used for each block of the dataset.

M/V Geniusbank: The Kongsberg EM 3002D is a dual-head sonar system, which can be operated in shallow and medium water depths. The system operating frequency is 300 kHz. The system has a maximum ping rate of 40 Hz and 508 beams equiangular spaced in target detection mode. The swath angle can be of up to 200° and the swath width is up to 10 x the water depth. The EM3002D performed generally well. A certain amount of noise is evident within the data set.

M/V Jetstream: The M/V Jetstream was equipped with a Kongsberg EM 3002D multibeam echosounder that was previously employed on M/V Geniusbank, described above.

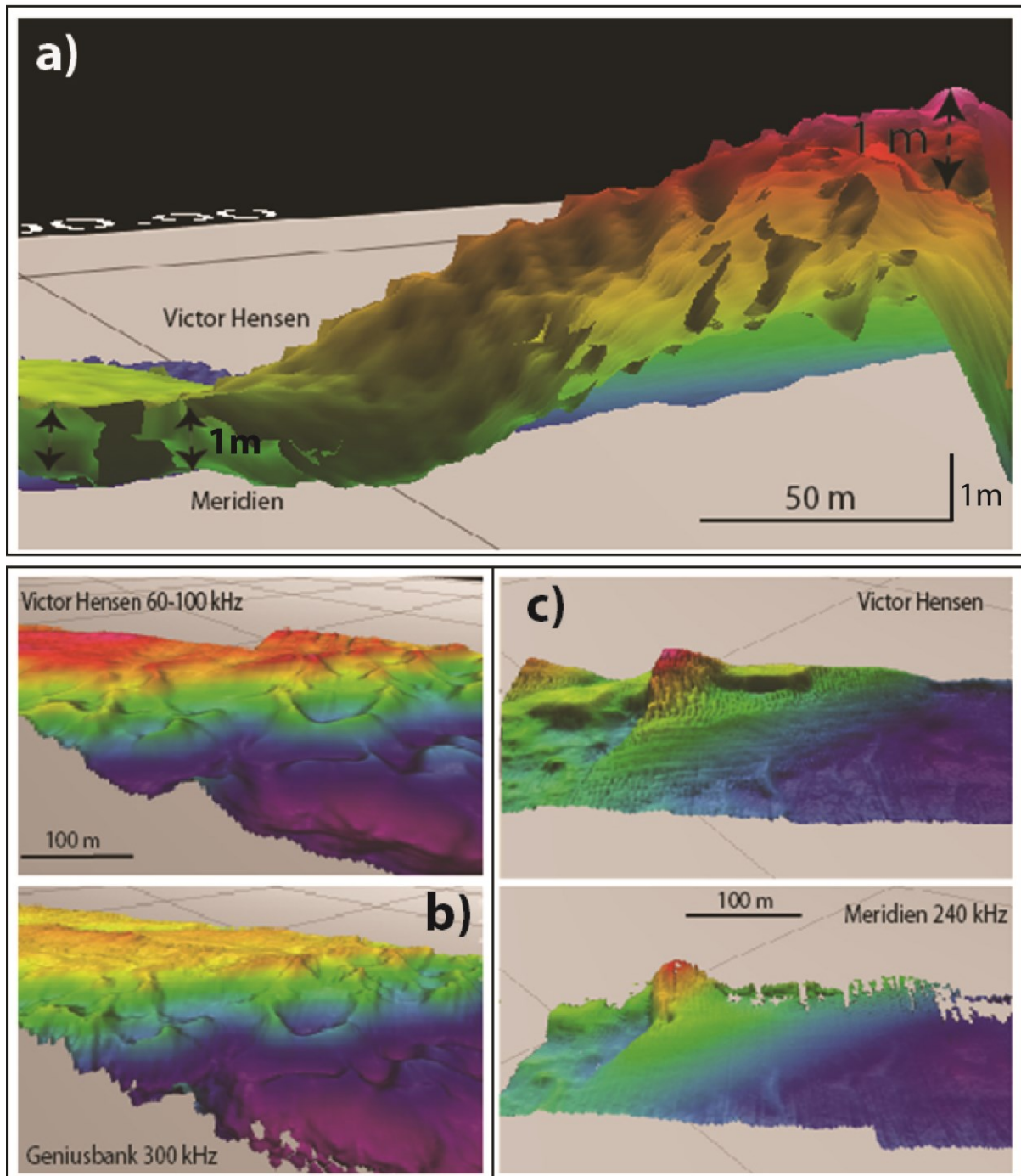


Figure 3.15: Comparison of the data collected by the three available multibeam systems. Figure a) displays the 1 m height difference between the surface collected by the M/V Victor Hensen and the M/V Meridien. Figures b) and c) show the difference in resolution over the same portion of seabed due to the different operating parameters of the systems: colour of depth scale and bin size (1.5 m with 9 soundings per bin) are identical in all figures. M/V Victor Hensen EM710 operates at lower frequency and therefore has a lower resolution than M/V Geniusbank multibeam system (left side) and M/V Meridien sonar (right side). Note that the instrument operating on M/V Geniusbank was also employed on M/V Jetstream.

Acquisition geometry

The main transects of the whole dataset were collected with by M/V Victor Hensen, as its onboard equipment is suitable to acquisition at a wide range of depth. On the other hand, M/V Geniusbank and M/V Jetstream were employed at different times for acquisition in areas where ship wrecks were present, collecting several shorter and close spaced lines, in order to gain a higher resolution in those area. M/V Meridien was also employed for the acquisition of wreck surveys and to cover some portions that had poor or zero coverage from the M/V Victor Hensen survey.

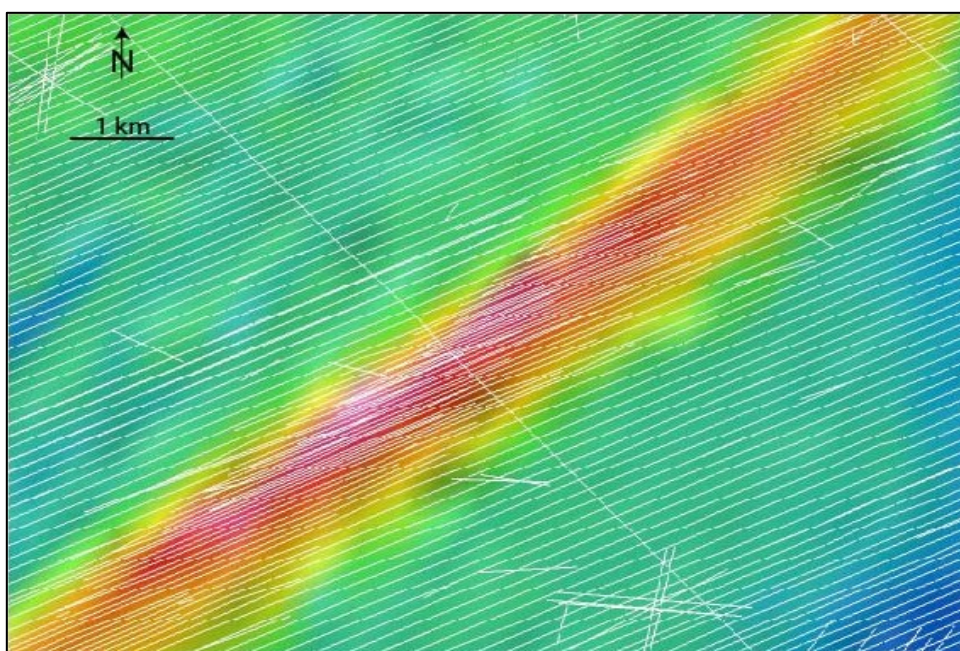


Figure 3.16: Ship tracklines displayed in white on the DTM, showing higher density of lines in correspondence of the sand bank (depth varies from -30 m, blue, to -18 m, red).

During the processing, the data were evaluated according to the requirements of IHO order 1 (IHO s-4 publication, 2012). This order requires a standard of nine valid measured depths in a cell of 2m x 2m in water depths of up to 40 m. After a QC of the data, a gap filling survey was performed but, because the sediments in some parts of the survey area were found to be highly mobile, a survey line collected to fill a gap often does not fit to the previously surveyed data. In order to avoid mismatching in the data, a filling line was only collected for those areas where more than one cell was not covered as required. Nevertheless, in areas of

larger bathymetric variations where data of every cell was necessary to model the seafloor correctly, such as sand wave areas, all gaps were filled (Fig. 3.16).

Data standards and accuracy

The required IHO order 1 standards were met for all three blocks (as indicated by the navigation report provided with the dataset), despite the tidal model not performing well. This is visible from some area of the dataset where gridding lineations are visible on the seabed. The water sound velocity corrections made in real-time during acquisition were considered sufficient and no significant post-cruise velocity corrections were required. A crossline was surveyed for each cell in order to perform a crosscheck. Each crossline was collected at the centre of the cell, and perpendicular to the main transects, giving crosslines every ~8 km. These were used during the processing phase as a crosscheck to verify the quality of the data, comparing it to the field sheet that was obtained after the processing: the following paragraph describes the main problems and accuracy level of the data for each block.

Block 1: the eastern and western parts of the block do not display any major issues affecting the required data standard, while the central part shows more variations that are caused by sand waves and variable depth of the seabed.

The block field sheet was compared to the crossline, showing a height difference of up to 60 cm in some areas, due to the lack of precision of the tidal model (see Appendix 1 for details). An example of the difference of precision obtained with the tidal model and GPS derived height is reported from the navigation report in Fig. 3.17, giving evidence of the linear and regularly spaced artifacts present in the data where the tidal model was applied.

Block 2: cell 064 to 090 (western part) show neighbouring swaths not matching with the field sheet because of height problems caused by the tidal model (these cells are far from Dover) while 091 to 111 (eastern part) which are directly situated near the tide gauge of Dover were processed using the tide model of Dover that fitted well in this area. The average mean value for the height difference is 10 to 15 cm respect to the cross profiles.

The high mobility of the sediment caused problems in the area of cells 091 to 105 as the survey was carried out in more than one phase at different times and the data do not match because the sediments had moved and the seabed morphology changed. In the Varne Bank, difference between two tide periods is visible as the sediment moved (infill survey 17.05.2007, see Fig.3.18).

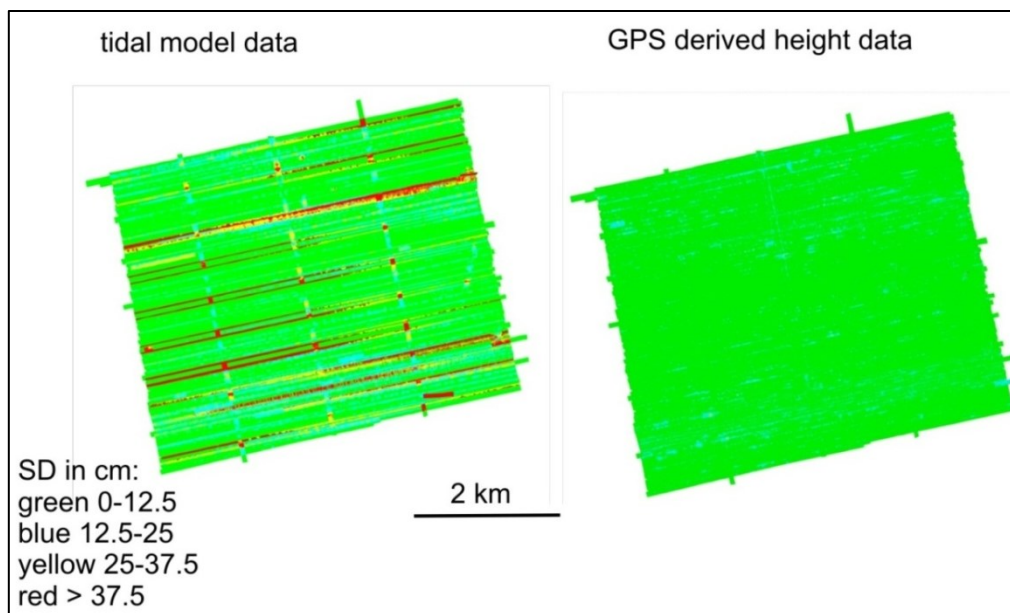


Figure 3.17: Comparison between the standard deviation calculated using different models: mean depth and standard deviation were derived from the resulting grid for each cell. The tidal model data (left) and the GPS-height derived data (right) are compared. The cell is located at great distance from tide gauge Dover (from navigation report, modified). Note that the orientation of the large SD “stripes” parallels the ship tracks (azimuth ~N45E).

Block 3: cells 113 to 124, located near the tide gauge Dover, were processed using the model Dover that fitted well. On the other hand, cells 125 to 175 shows some height difference with the cross profile and own tide gauge data were applied to them, resulting in a 15 to 20 cm standard deviation between crosslines. Finally, the south-western part of the block is affected by moving sediment that also caused a variation between the main set and the cross line that was successively surveyed.

Table 3.5 summarises the Total Propagated Error (TPE) obtained for each block after the processing.

	Deep TPE (30-50 m)		Shallow TPE (0-40 m)	
	horizontal accuracy inner/outer beam (m)	vertical accuracy inner/outer beam (m)	horizontal accuracy inner/outer beam (m)	vertical accuracy inner/outer beam value (m)
Block 1	1.0/1.1	0.20-0.25/0.60	1.0/1.0	0.20/0.45
	depth > 50 m	depth > 50 m	depth < 30 m	depth < 30 m
Block 2	1.0/1.0	0.27/0.50	1.0/1.0	0.17/0.20
	depth > 30 m	depth > 30 m	depth < 10 m	depth < 10 m
Block 3	1.0/1.0	0.25/0.50	0.9/0.9	0.17/0.17
	at 40 m depth	40 m depth	depth < 40 m	depth < 40 m

Table 3.5: Summary of the Total Propagated Error for the multibeam data as reported in the navigation report.

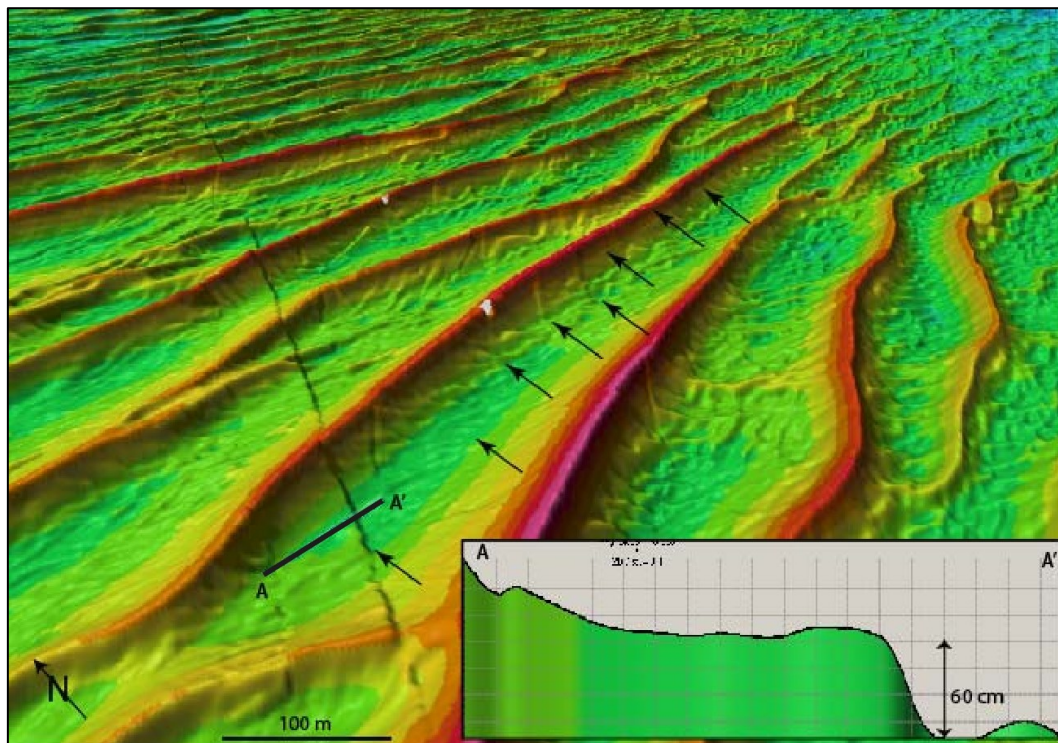


Figure 3.18: Height difference in the data up to 60 cm displayed in the sand wave area, cell 091_105. Some gaps are also present in the 1.5 m cell size data despite the very close spacing of the tracklines, as shown in Fig. 3.16. The black arrows indicate parallel lineations (azimuth ~N45E) due to the tidal problem and to the high mobility of the sediments. Depth varies from - 28 m (green) to -18 m (pink).

With regard to xyz positioning corrections, acquisition reports indicate that corrections that were applied to xyz positioning were extracted from Vessel Configuration Files (VCF). VCF contains sensor definitions with information regarding offsets, correction values and system configurations (beam numbers of echosounder) as follow:

- offsets relative to the centre of gravity (COG),
- sound velocity information
- dynamic draft changes
- dynamic vessel motion (gyro heading, roll, heave and pitch).
- static corrections for gyro heading and error for roll, heave and yaw heading alignment of the multibeam system

Position information included in the raw data was based on geographical coordinates, so no UTM definition had to be specified during the conversion. Each sensor was edited separately, creating a separate file with correlation through time, to which VCF were applied. Data was edited for possible spikes. This was done either manually or by using filters. If spikes were found, they were marked and flagged as not to be used for further calculation within the software. The resulting positioning gaps were interpolated over time by calculating new position values from neighbour measurements.

Data gridding

The tidal corrected bathymetry data were provided to Imperial College London in a GSF files format on 3 TB harddrives. Owing to the tidal correction issues it was decided not to include the data collected for the wreck investigation. Data from M/V Victor Hensen is the primary set of lines, together with the infill collected with the three other vessels.

Two main sets of DTMs were gridded using DMagic Fledermaus v.7, following different procedures: the first set, gridded at 7 m, and the second set, gridded at 1.5 m.

For the first set, all the parallel lines were gridded systematically with a 0.0001° (corresponding to ~ 7 m) bin size, creating a DTM for each block with the default gridding parameters (weighted moving average with weight diameter of 3 units). The cells were then merged into a single DTM for each block and then into a final 7 m bin size DTM of the entire

survey. The cross-lines (Fig. 3.19) were not included in the DTM as lines collected with a perpendicular direction to the main set can introduce artifacts (i.e. because of the tide correction), especially if collected at different times. To merge the DTMs the “surfacemerge” routine of Fledermaus FMCommand was used.

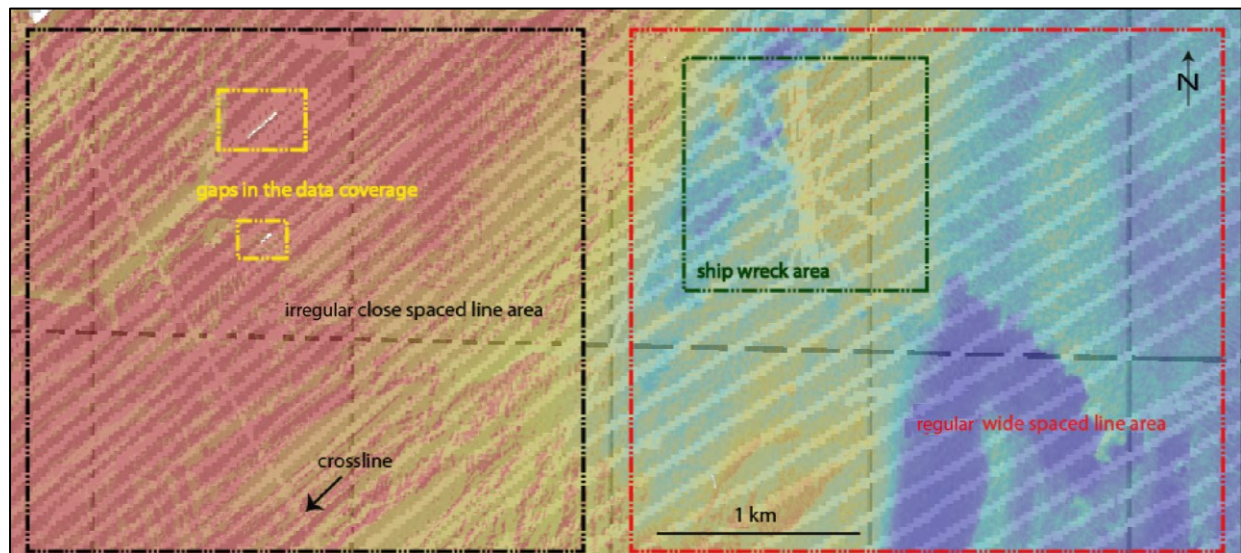


Figure 3.19: Ship tracklines of a small area of the MBES survey showing the main acquisition transects with NE-SW direction, a cross-line, a ship-wreck survey and some gaps in the data coverage.

For the second 1.5 m set, because of the high-resolution, the files were gridded selecting only part of the data, in order to avoid overlapping of multiple lines in the same area and introduction of artifacts due to the tide model. This problem was particularly important in some areas of the dataset where the spacing of the lines was closer. In order to gain a complete coverage of the DTM, more lines were gridded, introducing linear artifacts (Fig. 3.20) that are obvious in some areas. Although these features look similar to real seabed lineations, it is possible to identify the artefacts as they are parallel to the ship tracklines and they show a lower relief and different pattern if compared to the seabed lineations identified in other area of the dataset (see profiles in Fig. 3.20).

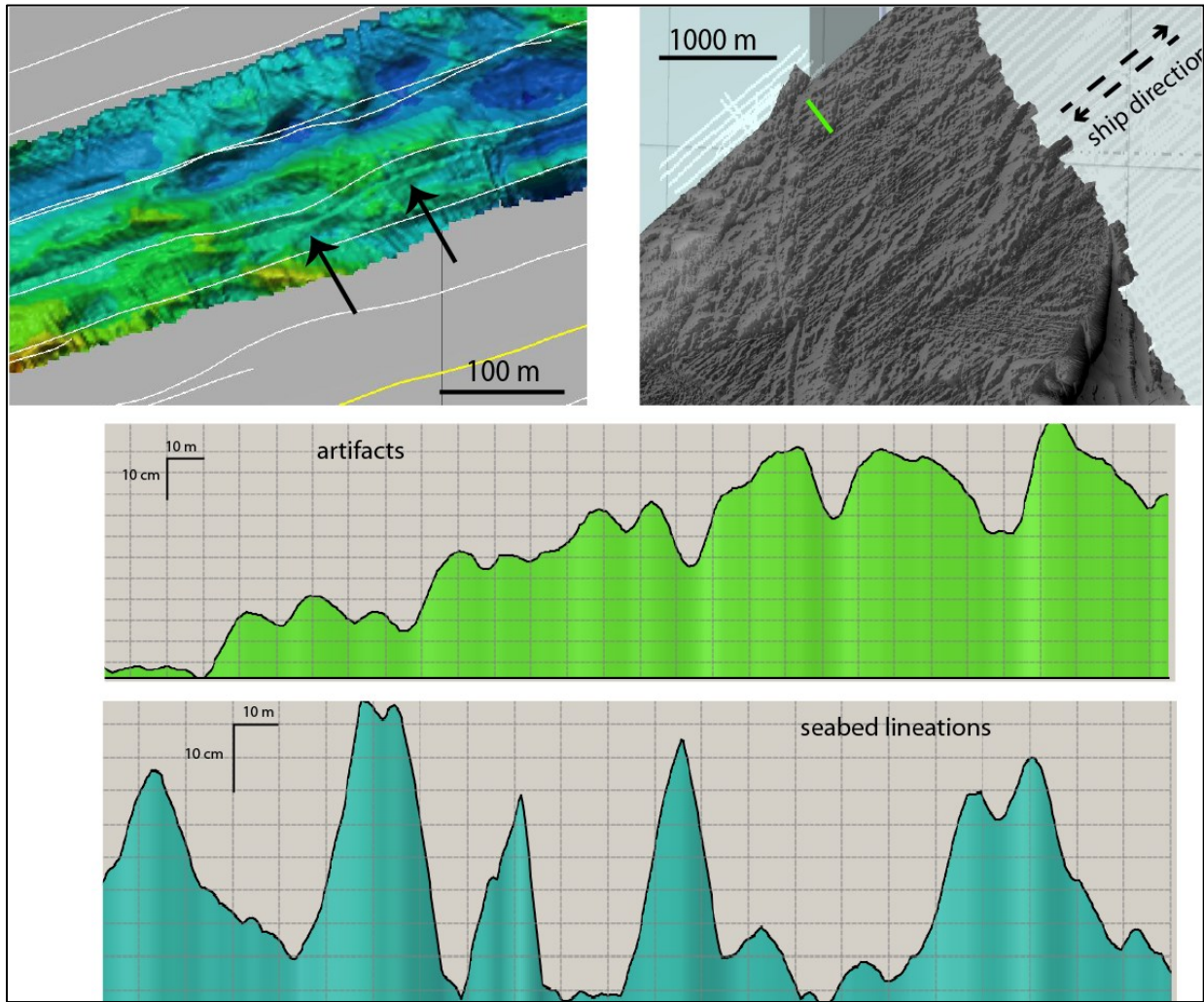


Figure 3.20: Artifacts parallel to the ship tracklines (oriented ~N45E and displayed in white). The upper left image displays the DTM obtained from four parallel lines: the black arrows indicate the area where the artifacts have an undulating geometry that follows the ship tracklines. The upper right image shows the final cell DTM displayed with grey colour-scale, showing the effect of the lineation artifacts on the final DTM. Green line is the location of the profile (green, of trackline artifacts (bright green)). Blue profile displays real seabed lineations, showing two very distinct geometry and higher amplitude than data artifacts.

As the quality of the data and the swath width was variable throughout the dataset, different gridding parameters were used in different areas where the default parameters did not perform well. Other than the default moving average, the shallow, deep and Root Mean Square algorithms are available for the gridding, giving different results as indicated by the tests shown in Fig. 3.21. Not to modify the morphology of the seabed, the average algorithm was used for all the dataset, increasing the dimension of the weight diameter when necessary and therefore allowing a wider coverage at the same cell-size but with a lower resolution, as shown in Fig. 3.21.

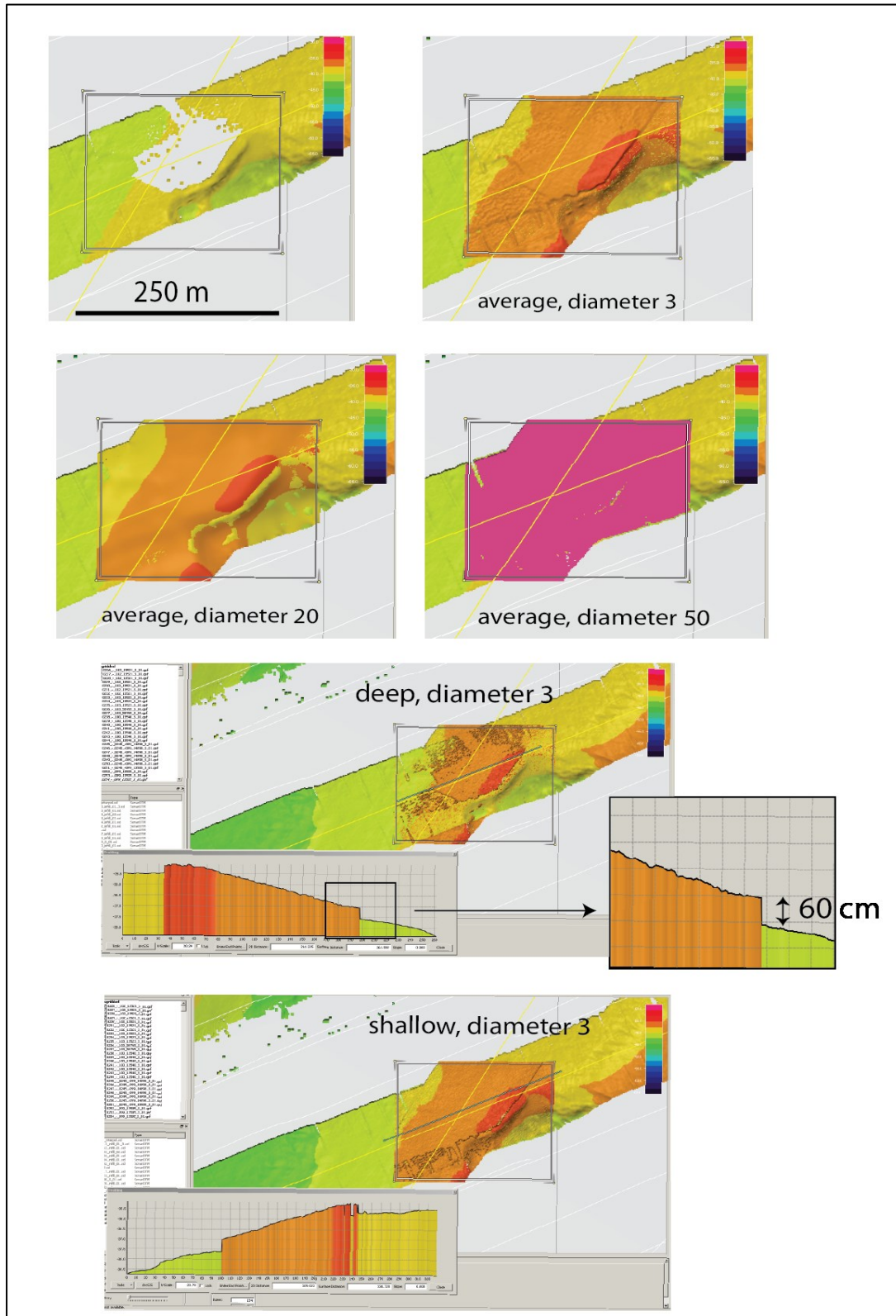


Figure 3.21: Examples of the gridding tests that show how the selected diameter and algorithm influence the DTM. The plots show the same piece of seabed, gridded using two tracks (yellow lines). As a rule, increasing the weight diameter results in a smoother DTM. Variations in the interpolation method (selecting deeper or shallower value) produce a strong variation in depth with respect to the average algorithm (selecting an average value), significantly changing the seabed morphology.

Where it was not possible to cover all the gaps with data, small areas were interpolated. The interpolation algorithm allows the selection of the value of the surrounding cells. This can be chosen from average, maximum and minimum value. The minimum number of neighbours determines the minimum number of filled cells. Because of the very small cell size, the interpolation did not performed well and only very small gaps were filled using this method. Fig. 3.22 shows an area where it was not possible to cover the entire gap without changing the cell size. In order to fill all gaps with the interpolation algorithm, bigger cell size, and lower resolution, would have been required.

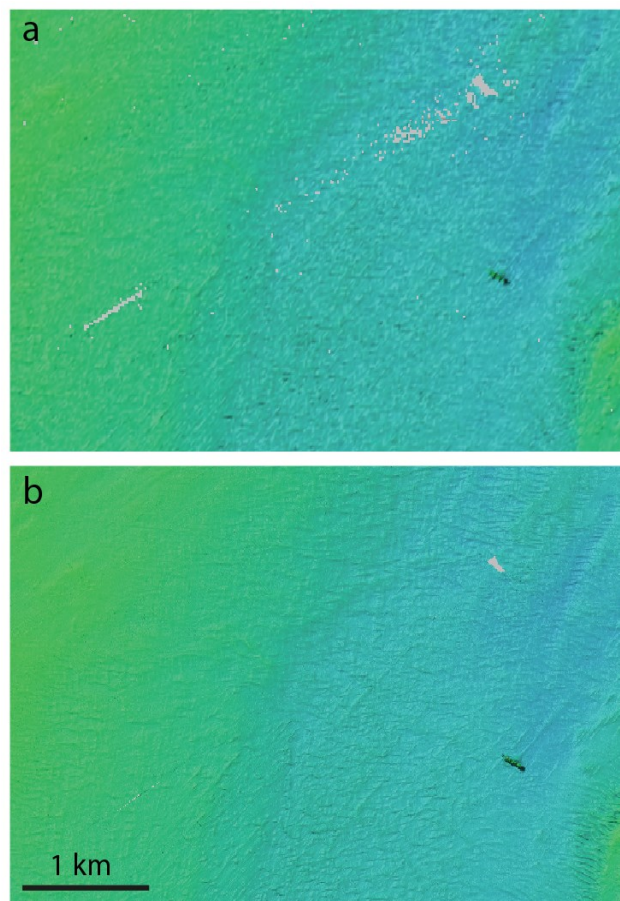


Figure 3.22: Non interpolated (a) and interpolated (b) DTM. The biggest blank area (shaded grey) was not completely filled, due to a low number of neighbour cells selected to perform the interpolation.

The single DTMs were then merged in a single cell DTM through the “surfacemerge” routine, which produces the final DTM with the lowest available resolution. Because of this reason, it was necessary to keep a constant cell size leaving some gaps in the dataset. In

particular, the sand wave area has a high number of gaps and low resolution areas as there were some gaps in the data and the quality of the data was generally poorer than the rest of the dataset. This was due to the sediment transport and tidal model problem as described in the previous section (see Fig. 3.18).

Despite the availability of different sets of DTM at different resolutions (1.5 to 80 m) allows for a regional mapping (on the 30 and 80 m DTMs) of the seafloor features and geology, the detailed analysis of small scale features is only possible on the high-resolution dataset, as shown in Figs. 3.23 and 3.24 where the level of details visible at different resolutions is compared. In order to integrate a regional dataset and a high level of detail, a final merge was produced (see following section 3.2.6), allowing for the analysis of the entire study area and including in the same regional-scale dataset those areas where high-resolution dataset is available.

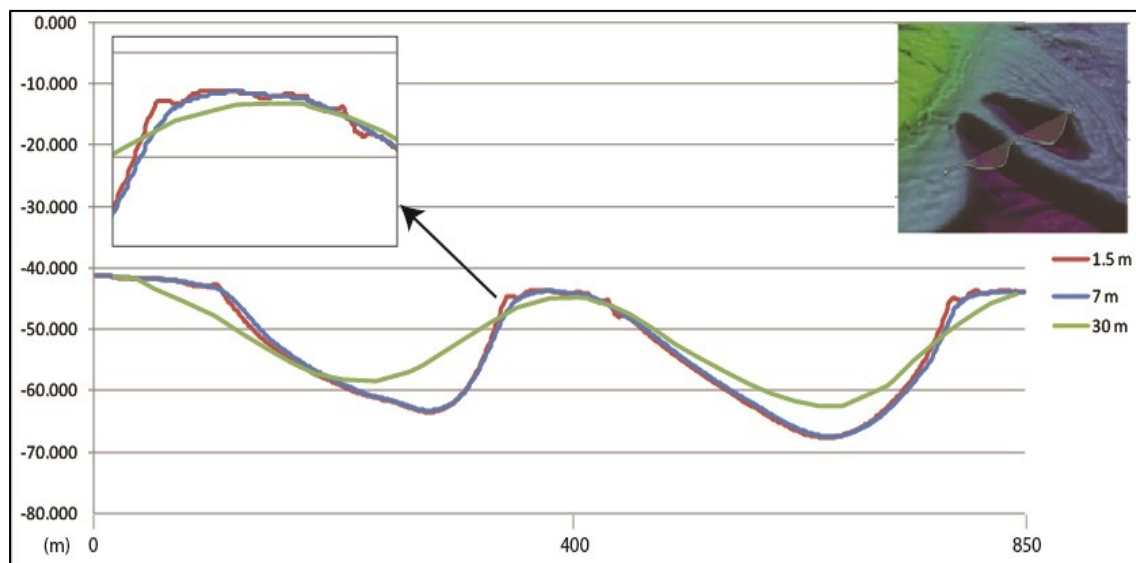


Figure 3.23: Comparison of the same depth profile extracted from the three different DTMs; the profile is located on the feature shown in Fig. 3.24. A strong difference is displayed between the two higher resolution DTMs and the 30 m DTM.

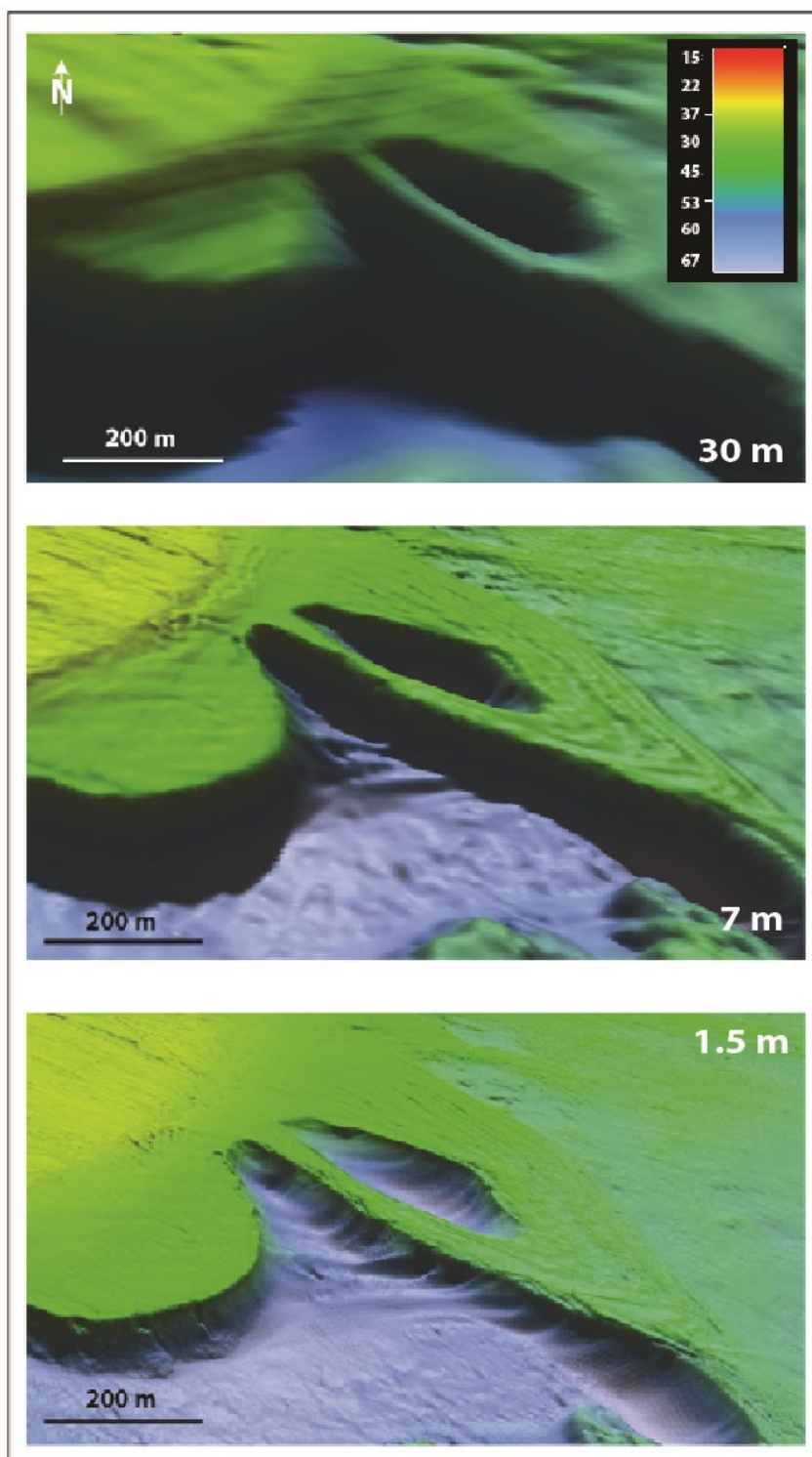


Figure 3.24: Comparison of the resolution of the three DTMs displayed using the same colour-scale: upper box is the 30 m bin size; the central figure is the 7 m DTM and lower is 1.5 m bin size DTM where the resolution was strongly improved. Depth is in m.

3.3.6 Bathymetry Data merge

Two final merged grids of the study area were produced from the five available bathymetry datasets to obtain the highest possible resolution for each area of the dataset, integrated in a single grid. The first merge was named “merge v.1”: it consists of a Fledermaus scene file containing different DTMs at different resolution. These were cookie-cut and loaded into the same workspace. It was not possible to merge the DTMs into a single file as the obtained DTM would have been gridded at the lowest resolution (80 m). The second merge was named “final merge”: it consists of a single DTM file containing all the available dataset merged into a .asc file and containing different datasets, each at its resolution using Global Mapper.

A detailed description of the procedure adopted for the gridding of the two final merges is given in this section.

Data gridding

Merge v.1

This merge was gridded using four different datasets at different resolutions, as shown in Table 3.6:

Dataset	Bin size (m)
GEBCO	80
UKHO English Channel	30
SHOM	40 and 80
ROB-RCMG MBES	5

Table 3.6: Summary of the employed datasets for the gridding of merge v.1

The gridding of merge v.1 was performed using DMagic Fledermaus v7. The .xyz files of the four datasets were exported from the available DTMs in order to re-grid all three dataset at the same bin size. Two bin sizes (80 and 30 m) were selected following two criteria: the sounding density of the three datasets and the final .sd file size. Once the DTMs were re-gridded, for each area the dataset with the better quality was selected, performing a detailed small scale cookie-cut of the three surfaces avoiding areas of gap or overlap. The .xyz files of the cookie-cut surfaces were then exported and re-gridded in a single DTM in order to avoid

the border effects that are present when merging three surfaces with different resolution. Despite this, a strong difference in depth is visible in both 80 and 30 m bin size DTMs between the GEBCO grid (which has mean sea level datum) and the merged bathymetry and SHOM bathymetry, which have lower tide datum. The gap between the surfaces varies from 1 up to 5 m in some areas.

Final merge

The second version of the final merge was produced using DMagic Fledermaus 7 and Global Mapper 14 in order to produce a single 30 m bin size DTM, called “final merge”, where different surfaces gridded at different resolutions were merged. The final DTM was composed of the same DTMs that were used to produce the “scene file merge” with the addition of the MCA MBES data.

The used DTMs and their bin size are summarized in Table 3.7:

Dataset	Bin size (m)
GEBCO	80
UKHO English Channel	30
SHOM	40 and 80
ROB-RCMG MBES	5
MCA MBES	7

Table 3.7: Summary of the employed datasets for the gridding of Final merge

Fledermaus 7 was used to extract the .asc Arc Grid files from the previously gridded .sd files that were used for merge v.1. The .asc files were then loaded into Global Mapper into a single workspace and overlapped in order to display the dataset at the highest resolution in each area. For example, in some area where the SHOM data was gridded at 80 m, it was chosen to display the UKHO data, also available in the same area, as it has higher resolution (Fig. 3.25). Importantly, differently from merge v.1, the grid obtained in Global Mapper allows for more than one dataset to be used in the same areas, in particular in those areas where one of

the dataset displays small gaps. Using more than one dataset at the same time masks the gaps as values from a second dataset are used in the gap areas, as for the case of the south-eastern portion of the SHOM bathymetry that could not be used in merge v.1 because of areas with big gap in the data. Once gaps were filled and boundaries between dataset were adequately smoothed, the obtained DTM was exported as a single .xyz file gridded at 30 m in Fledermaus.

If compared to merge v.1, “final merge” shows a strong improvement in the quality of some areas of the merge and in the smoothing of the boundaries between different datasets (Fig. 3.26). The areas that were improved the most are the French Sector of the Channel and the Dover Strait area that are better imaged as the higher quality of the MBES dataset was preserved during the merge. On the other hand, the gap due to different tide datum between UKHO and GEBCO bathymetry present in merge v.1 is still visible, ranging between 1 to 5 m amplitude.

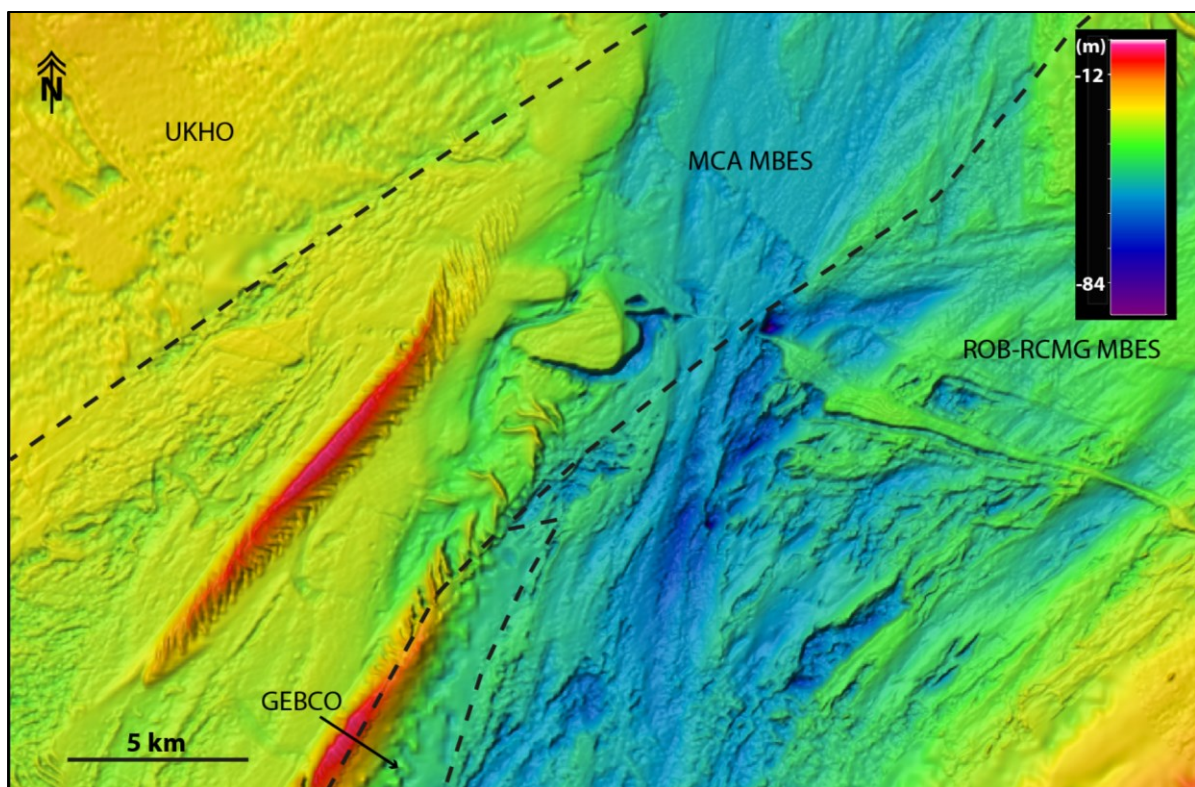


Figure 3.25: Extracted image from the “final merge” DTM, showing the boundaries (black dashed lines) between different datasets that were used. Note that the boundary between MCA and ROB-RCMG MBES is the UK-France border.

A smaller final merge of the Dover Strait area was produced using the available high-resolution datasets, obtaining a final DTM gridded at 0.00002 arc deg (≈ 1.5 m) bin size. As the resolution of the data is very high, the data was loaded, processed and gridded in lat/long WGS84 system as the projected UTM grids show lower resolution and linear artifacts due to the projection of the data. The same procedure described for the bigger “final merge” was used, loading the dataset in a Global Mapper workspace and overlapping the files in order to achieve the possible highest resolution for each area. The following datasets were used are indicated in Tab. 3.8:

Dataset	Bin size (arc deg)
GEBCO	0.001
UKHO English Channel	0.0004
SHOM	0.0005
ROB-RCMG MBES	0.00007
MCA MBES	0.0001
Cell 91_95 of MCA MBES	0.00002
Cell 96_105 of MCA MBES	0.00002
Obtained Dover Strait	0.00002
Merge	

Table 3.8: Summary of the used datasets and their bin size expresses in arc degrees used for the lat/long WGS84 system used to grid the data.

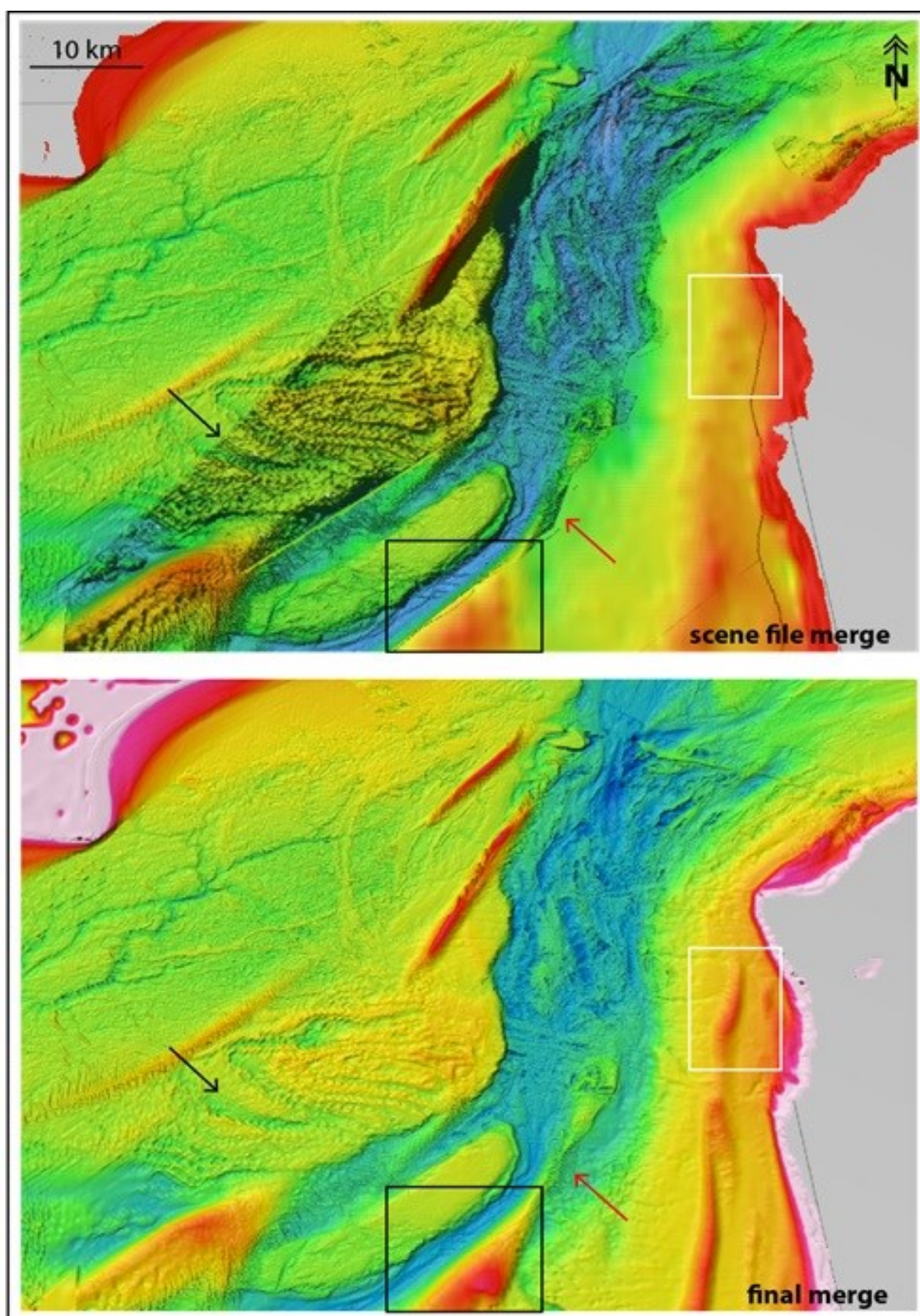


Figure 3.26: Comparison between the “scene file” and the “final merge”, showing the improvement in the data quality. The arrows show the difference in the smoothness of the boundaries between the different datasets in two different areas (black and red arrows), looking visibly smoother in the “final merge”. The two boxes show how the resolution of the French sector of the dataset was improved thanks to the integration of more than one dataset in the same area.

Data analysis

The final merge obtained from the integration of the five available datasets was used for the regional scale interpretation of the seabed geology (described in Chapter 4) and for the morphological analysis of the flood-related bedforms (described in Chapter 6) present in the palaeovalley channel network. As the main targets of the analyses were the bedrock and its geology, a map of the areas of the seabed that are covered by sediments was produced, in order to evaluate the portion of dataset where the bedrock is visible at the seafloor. The seabed sediments can be treated as “noise” when performing the mapping and morphometric analyses of the bedrock and of the flood-related bedforms that are masked by the sediments deposited on the seafloor. By measuring the portion of seabed that is covered by sediments it is possible to evaluate which areas allow for a reliable interpretation of bedrock geology, flood features and tectonic structures from bathymetric data.

Seabed sediment mapping

The sediment cover map reveals the extent and distribution of the seabed sediments as they mask the bedrock outcropping at the seafloor, making the mapping and analysis of the seabed geology more difficult or not possible in some areas. The sediments were also mapped in order to assess the relationship between their distribution and the contemporary seabed shear stress across the English Channel. The sediment cover map was drawn using both high-resolution MBES and “final merged” bathymetry, ranging from 1.5 to 80 m bin size resolution. Due to a large variation in both the scale of the structures that compose the Quaternary sediment cover and in the resolution of the available data, it is important to define the scale of the features that are considered as sediment cover. Only those features that were visible at the resolution of the “finalmerge” DTM were therefore included in the map of the seabed sediments.

Geological mapping

The geological mapping of the study area was based on the interpretation of the acoustic response and texture of the seabed, with the aim of mapping the structural elements and the boundaries between the bedrock formations. To do so, the regional-scale “final merge” and the 1.5 m MBES datasets were used, together with the SRTM onshore topography (USGS, 2006, <http://srtm.csi.cgiar.org>) at 3 arc-second (~ 90 m) resolution for the onshore area. In

particular, .geotiff images of SRTM sheets 36_02, 36_02, 37_03 and 37_03) with resolution of 6000 x 6000 pixels (Jarvis et al., 2008) were imported in the ArcMap project. These were integrated with the BGS 1:250000 solid geology maps (sheets Thames Estuary, Dungeness-Boulogne, Wight) (BGS, 1988, 1989, 1995) that were used as reference, together with BGS (digitised from available BGS maps cited above) and commercial cores (J. Collier, pers. comm.) available in the area. The onshore geological mapping was based on the available BGS 1:250000 onshore maps (DiGMapGB-250, Bedrock Geology) for the British side and on the BRGM Carte Geologique de la France, scale 1:250000, sheet Rouen (BRGM, 1979) and Carte Geologique de la France scale 1:1000000 (BRGM, 2003) for the French side. The data were then loaded into an ArcGIS project where the geological map was drawn.

The mapping was performed using the seabed acoustic response (which produces different “textures” and “patterns” depending on the lithology) and its variations visible on the seabed, in order to identify the boundaries between the different lithologies and to map the tectonic structures outcropping at the seabed. The geological mapping of the seabed was also based on the similarity between the offshore acoustic texture of the different formations and their onshore outcrops. The analysis and comparison of the bedrock outcrops and seabed texture was ground-truthed by field observations made at Lulworth Cove and Kimmeridge Bay, Dorset. The scale of the outcrops was selected in order to show the similarity of the seabed texture and surficial expression at different scales, ranging from centimetre to hundreds meter scale. The acoustic texture of the seabed was also analysed using slope and rugosity values for each outcropping lithology. Slope and rugosity analyses were performed using DMagic Fledermaus v.7. Slope was calculated using “fitted plane” algorithm, obtaining a surface that represents the maximum slope at each grid centre, expressed in decimal degrees. Rugosity was calculated as the ratio of surface area and planar area of a surface. Both attributes were calculated on both the merged and the high-resolution bathymetry, although only completely sediment-free areas could be used for the bedrock interpretation as quaternary sediments display higher slope and rugosity values than bedrock. In order to overcome this problem, both value ranges were restricted to low values, strictly related to bedrock, while high values associated to the sediment cover were not considered in the analyses. This filtering permitted to obtain an imaging of the seabed heterogeneities exclusively due to lithological variations or to structural features present at the seabed.

3.4 Summary

The availability of both seismic reflection and bathymetry data at different scale and resolution in the study area allows for an accurate and detailed stratigraphic and geomorphologic analysis of the bedrock geology and of the flood-related features. If, on the one hand, the seismic reflection data can be used for the analysis of the bedrock shallow stratigraphy and of the sediment filled flood features carved into the bedrock, bathymetry can be used to understand the relationship between bedrock geology and flood-features geometry and distribution. While the availability of different sets of DTM at different resolutions (1.5 to 80 m) allows for a regional mapping, a final merged dataset permits for the integration of multiple resolution into a single DTM, covering the entire study area. The gridding of a merged dataset with multiple-resolution data also improved the quality of some areas that were poorly imaged in the single-resolution DTMs.

4 . Bedrock and surficial geology

4.1 Introduction

A detailed description of the bedrock geology of the English Channel seafloor and of the upper part of the bedrock succession will be given in this chapter, together with the classification and mapping of the distribution of seabed sediments covering the bedrock. The geology of the area was mapped using seismic and bathymetry data in order to have both an in-depth and plan view of the bedrock geology, with the aim of understanding the impact of the bedrock geology on the flood-related features present on the English Channel seafloor.

The obtained interpretation of the bedrock geology and tectonics will be used in the following chapters to understand how the geology of the area influenced the formation, morphology and distribution of the flood-related features present on the seabed of the English Channel that represent the target of this work. Particular focus was put on key areas such as the Dover Strait and the main lithological boundaries between adjacent basins, where sharp changes in the seafloor geology are marked by the presence of faults and folds. In order to characterise the bedrock stratigraphy and geology of the seabed, the following analyses were performed:

1) Mapping of the seabed sediment cover from the regional bathymetry dataset, to identify the areas of the seabed where no sediment cover is present and geological analyses of the seabed can be performed from bathymetric data.

2) Seismic reflection data were used for the analysis of the stratigraphy of the bedrock succession, through the calculation of the unit thickness and comparison of the seismic stratigraphic facies of the different formations throughout the study area. To do so, seismic-stratigraphic analysis was performed on the two available seismic datasets (Fig. 4.1), with particular focus on the high-resolution sparker data collected at the Dover Strait. The seismic facies of the different formations at the Dover Strait were then compared to the acoustic facies of the bedrock interpreted in the Central Channel area, using boomer dataset with lower resolution. In this area, no boreholes were available and it was therefore not possible to

calculate the thickness of the interpreted formations composing the succession as the seismic data was not converted to depth. Acoustic facies and their lithological interpretation was made using the seabed geology map, correlating interpreted reflector on data to lithological boundaries on the bedrock geology map.

3) Plan view mapping of the bedrock was performed through a detailed description and mapping of the acoustic character of the different formations using bathymetry data, together with a comparison with land outcrops of the same formations in the Central Channel area. The acoustic texture of the seabed was used to draw a geological map of the study area, identifying the main lithological boundaries and structural elements visible at the seafloor using bathymetry. The geological boundaries obtained from the seismic-stratigraphic analysis were successively included in the final geological map.

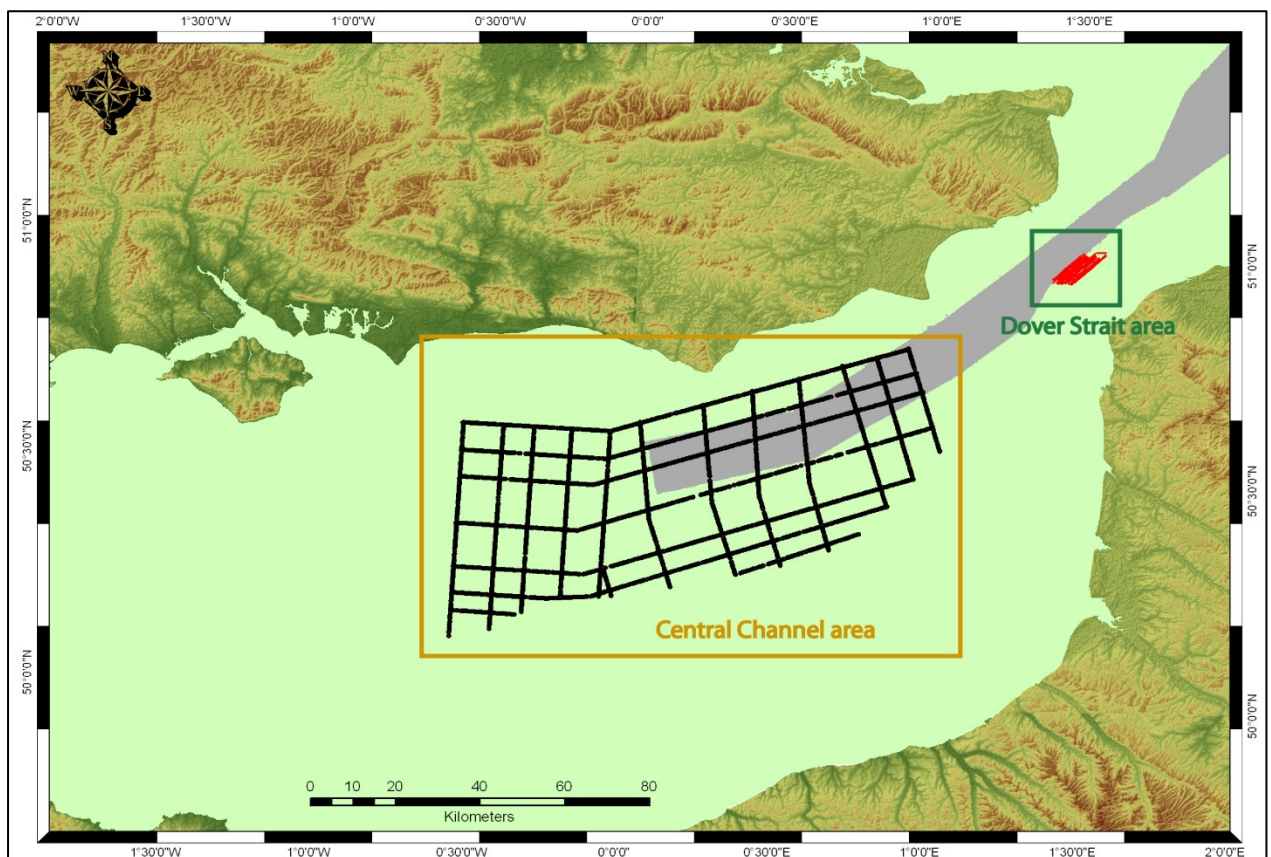


Figure 4.1: Location of the two seismic reflection datasets used for the seismo-stratigraphic analysis of the bedrock, displayed with onshore SRTM topography (USGS, 2006); see Fig. 4.9 for colour-scale.

4.2 Surficial seabed sediments

As the main target of this work is the analysis of the bedrock erosional features carved into the seafloor, all the areas where modern sediments mask bedforms carved into the seabed (with variable scale that ranges from metres to kilometres), were considered as sediment-dominated areas (see Section 3.3.6 for further details). The distribution of the seabed sediments in the study area was mapped for two main reasons: 1) estimate the percentage of the area of the seabed covered by sediments masking the bedrock geology and bedforms present on the seafloor 2) briefly characterise the sedimentary structures that compose the sediment cover and estimate their scale and distribution.

The obtained sediment cover map (Figs. 4.2 and 4.3) indicates 3800 km² of seabed covered by sediments, corresponding to ~18% of the study area. The distribution of the seabed sediments in the area is highly variable, depending on the effect of the currents that dominate the seabed in different areas. The mapped distribution of the sediments was compared to the seabed shear stress values calculated for the English Channel (Mitchell et al., 2010), displaying a good match between the areas with calculated high values of seabed shear stress and sediment-free areas (Fig. 4.2). The map shows the presence of wide sediment-free areas, particularly in the by-pass areas located at the Dover Strait and in the Central Channel. Conversely, a thicker sediment cover is present in the convergence areas located south-west of the Strait and in the Southern North Sea, where divergent currents drag and deposit sediments (Grochowski et al., 1993). The thickness of the sediment cover varies between 0.5 up to 5 m in areas such as the shelf on the east of the Isle of Wight (Hamblin et al., 1992).

The sediments that mask the bedrock are mainly composed of coarse-grained sand deposits locally overlapped by finer sandy and muddy sediments (Hamblin et al., 1992; James et al., 2007). The sediments present on the seafloor form sedimentary structures that show dimensions of highly variable scale, ranging from ripples of 0.15 m amplitude to 7 m relief sand waves, up to 30 m-high sand banks. In particular, the high-resolution MBES data show a variety in the shape and geometry of the sand waves that range from isolated sharp crested to 3D rounded crested or barchan shaped sand waves (Fig. 4.4).

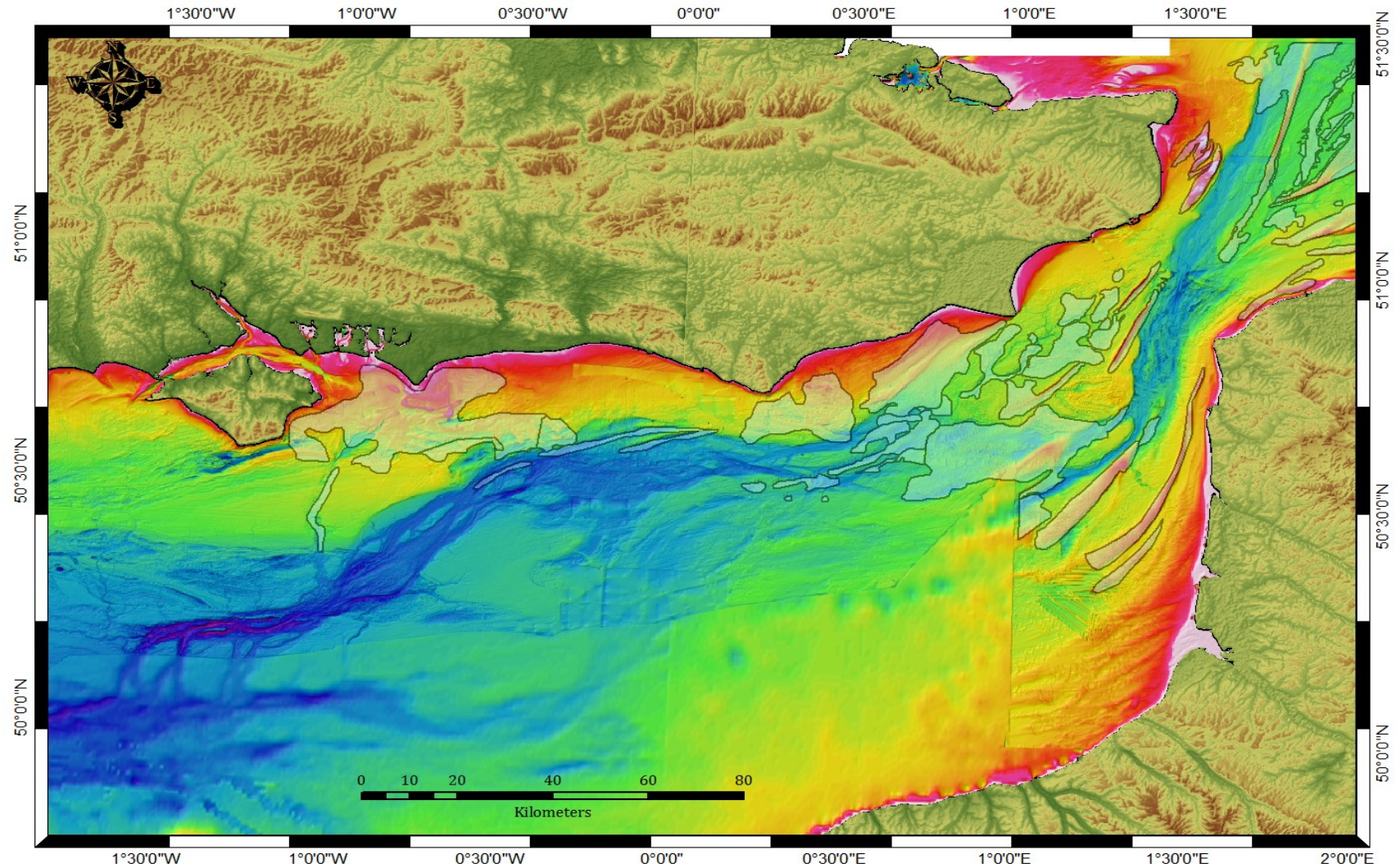


Figure 4.2: Overlay of the areas of the English Channel seabed covered by sediments (polygons) and bathymetry and onshore SRTM topography (USGS, 2006). The total surface of the seafloor covered by sediment corresponds to 18%. ;See Fig. 4.9 for colour-scale.

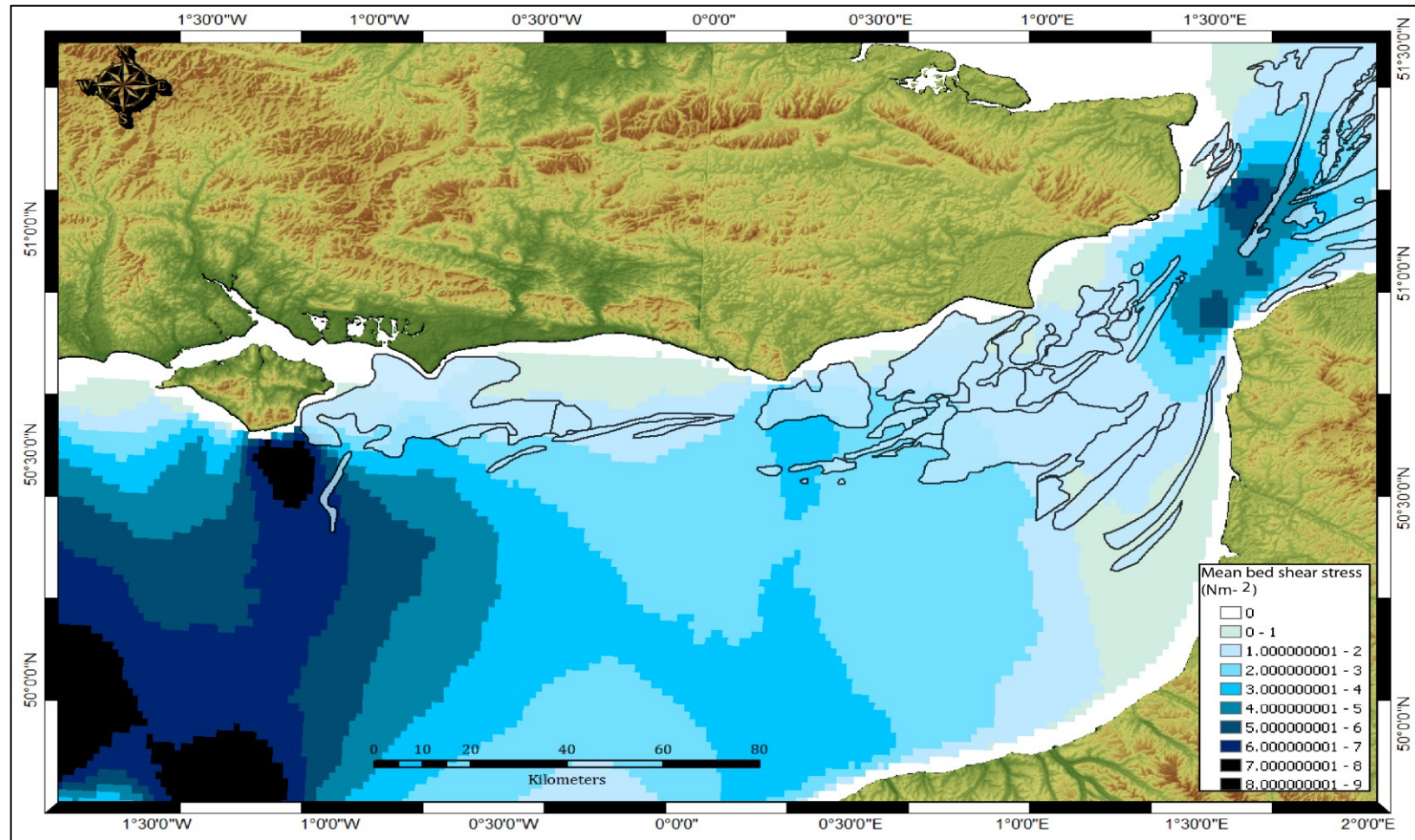


Figure 4.3: Overlap of the mapped sediment cover and of the shear stress values calculated by Mitchell et al. (2010) that developed a mesh finite-element model which simulates astronomical forcing and co-oscillating boundary condition. The first parameter is controlled by gravitational and rotational effects within the Earth, Moon and Sun orbital system, the second is calculated from tidal waves propagating into the domain from adjacent bodies of water. The two maps show very good correspondence as the areas with higher shear stress values ($> 2 \text{ Nm}^{-2}$, in blue) are almost completely sediment-free, while areas with low shear stress value ($< 2 \text{ Nm}^{-2}$, in light blue) display the presence of sediments in variable quantity. Onshore topography from SRTM (USGS, 2006), see Fig. 4.9 for colour-scale.

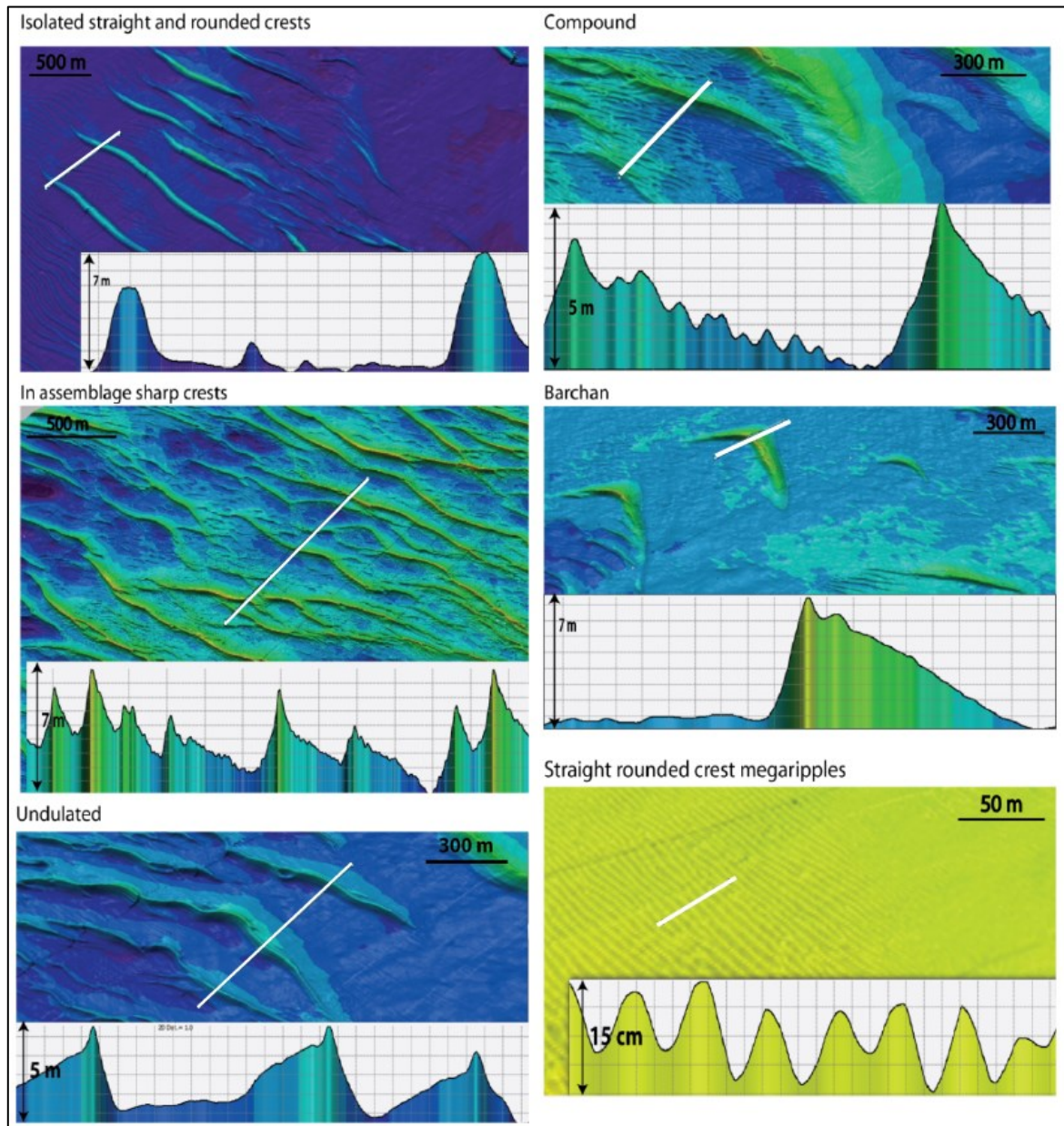


Figure 4.4: Examples of the different types of sedimentary structures that compose the seabed cover of the English Channel, ranging from 0.15 m high ripples to > 7 m sand waves. The DTMs have a bin size of 1.5 m. Location of profiles is indicated by white lines.

In conclusion, this analysis indicates that more than 80% of the seabed is sediment free, allowing for the geological mapping of the bedrock from bathymetry other than from reflection seismic.

4.3 Bedrock seismic stratigraphy

As shown by Fig. 4.1, the seismo-stratigraphic analysis of the study area was performed on two different datasets: a high-resolution dataset located at the Dover Strait (Dangeard I 2002 seismic) and regional dataset (Central Channel Seismic), characterised by lower resolution, wider line spacing and located in the Central Channel area (see Section 3.2.3 for details). The Dangeard dataset, given its high-resolution, allowed for the estimation of the thickness of the interpreted formations that was then compared to the stratigraphic column available on the BGS geology maps, scale 1:250000 (BGS, 1988, 1989, 1995). On the other hand, the lack of boreholes in the Central Channel area did not allow for a correlation of the units interpreted in the Dover Strait area with the same acoustic facies present further westwards.

4.3.1 Dover Strait area

The interpretation of the bedrock stratigraphy from the high-resolution dataset located at the Dover Strait resulted in the identification of 10 seismic units (Figs. 4.5 and 4.6) that correspond to the Cretaceous geological formations (see Fig. 3.6) outcropping in the Dover Strait area, on the north-eastern limb of the Weald-Artois anticline fold.

Middle Chalk: the base of this formation is represented by the reflector *WLC-pink*. This formation outcrops in the north-east area and thins towards the south-west, with a maximum value of more than 20 m in the north-east. The Middle Chalk formation displays a seismic facies characterised by internal high amplitude reflections and small-scale discontinuities and diffractions due to the presence of several flint horizons intercalated in the carbonate layers. The strata display a flat geometry with a maximum dip of 3° (see reflector *WLC* in Fig. 4.5).

Lower Chalk: this formation, given its up to 80 m thickness, is subdivided in four units in this area: White Chalk (its top is represented by the reflector *WLC-pink*), Grey Chalk (whose top is represented by *GLC-apple green*), Upper Chalk Marl (its top *UCM-dark green*) and Lower Chalk Marl (top is *LCM-light blue*). All four units display constant thickness and similar seismic character with internal sub-parallel layers. The units are monoclinaly folded where their dip sharply increases from 3° to 11°. The Lower Chalk Marl are characterised by the diffraction signal of the Channel Tunnel, composed of two distinct diffraction hyperbolae (Fig. 4.5).

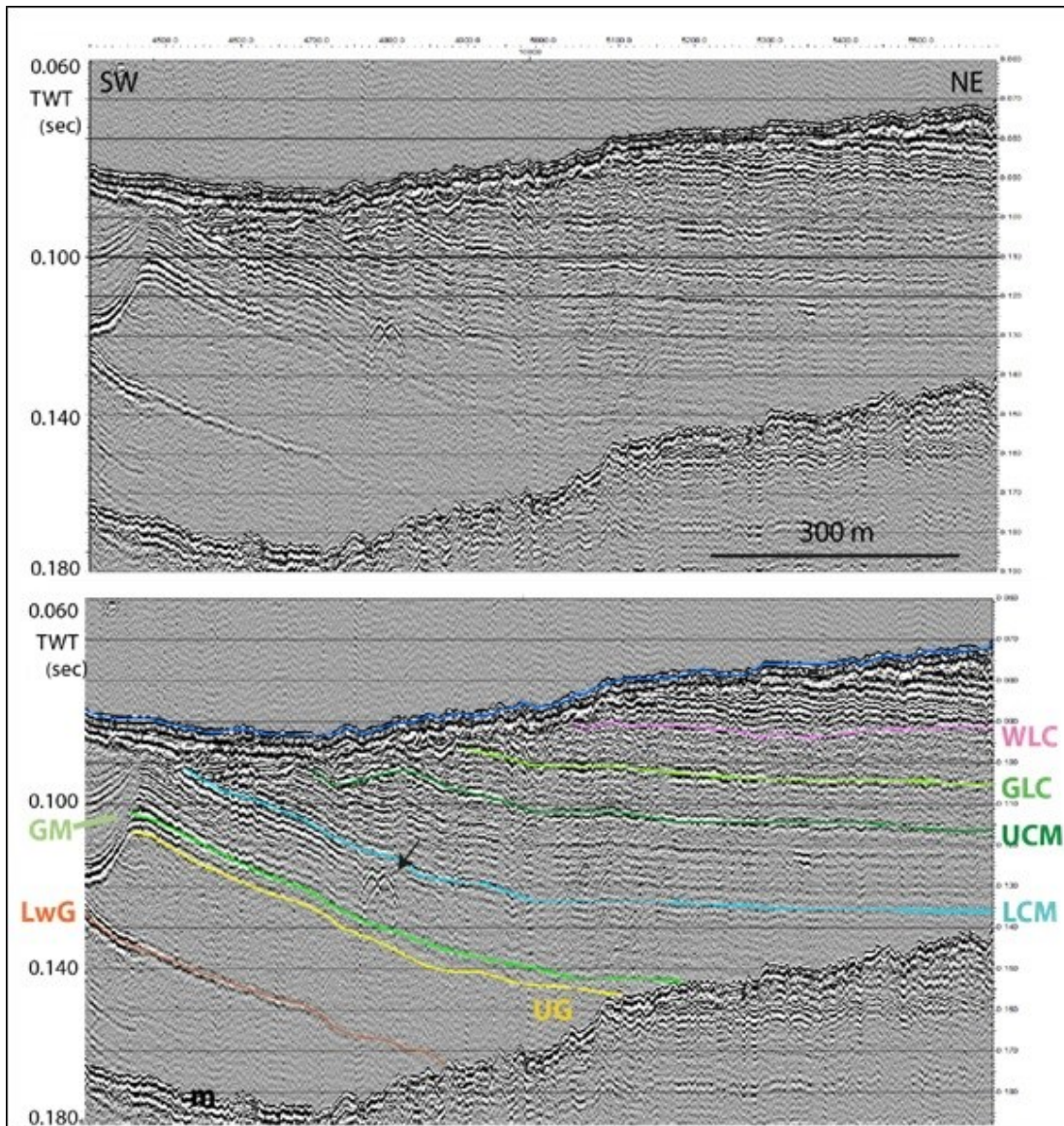


Figure 4.5: Seismic stratigraphy of the Dover Strait, showing the upper part of the Cretaceous succession on the north-eastern flank of the monocline fold. The black arrow indicates the Tunnel diffractions, the blue reflector corresponds to the seafloor. The reflectors are labelled with names as in the text, m represents the seafloor multiple. VE is 4 (all the VE in this chapter were calculated using velocities indicated in the acquisition report (J. Arthur, pers. comm.), see Section 3.2 for details). Location of the seismic line is shown in Fig. 4.6b.

Glaucconitic Marl: this formation is distinguished by a couple of very high amplitude reflections (*GM-green*) that mark a strong change in the acoustic character of the succession. This is due to a strong contrast in the acoustic impedance between the overlying Chalk and by the presence of condensed glauconitic layers in the Glaucconitic Marl. As the thickness of this unit is 5 m, on the seismic data the formation is only represented by the couple of high

amplitude reflectors. The average vertical resolution of the dataset is > 0.5 m and does not allow for the characterisation of the internal geometry of this thin unit. The resolution was calculated using frequency of 1 kHz and velocity of 2000 m/s as indicated in the seismic acquisition report (J. Arthur, pers. comm.).

Gault Clay: this formation is subdivided in two distinct units called Upper Gault (whose top is *UG-yellow*) and Lower Gault (whose top is *LwG-orange*), divided by a high amplitude reflector visible at ~60 m below the seabed when it is not masked by the seabed multiple. As this formation is mainly composed of clays, it displays a transparent facies that marks a strong acoustic and lithological contrast with the upper (Glaucconitic Marl) and lower (Lower Greensand) formations as indicated by the high amplitude of the top and basal reflectors of the Gault. Both the upper and lower units are strongly folded in the monocline area, showing a very steep north-east dip (up to 11°) and a maximum thickness of 70 m.

Lower Greensand: this formation is 40 m thick in this area. It is bounded at the top by the reflector *LG-brown* and displays a heterogeneous seismic facies, revealing internal stratified high amplitude reflections alternated to almost transparent layers: this is due to the presence of chert levels intercalated to glauconitic sand layers. Small scale channels (< 10 m) are carved into the upper part of the unit. Diffraction mark the erosional surface. The strata have a dip that varies from 2° on the south-west to 10° degrees in the monocline fold.

Wealden: two distinct seismic units were identified in the Wealden Group, bounded by the reflector labelled as *W-green*. This reflector represents the top of the Wealden which is divided into two seismic units, the lower masked by the seabed multiple at its base. Despite the multiple prevents from measuring the Group total thickness, a thickness greater than 100 m was measured in this area.

Both upper and lower units display heterogeneous seismic character due to the mixed lithology that composes the Wealden, formed by an alternation of fine deposits and amalgamated sandy channel bodies. The upper unit displays a variable dip of the internal layers (that ranges from 0° - 2° to 10°) that show sub-parallel undulating geometry, deformed by sub-vertical small scale faults and folds. Reflector displacement produces diffraction

hyperbolae. Sediment-filled channels are incised into the upper part of the Wealden Group where it outcrops at the seabed, as described for the Lower Greensand. The undulated reflectors present in the lower part of the formation display high amplitude and good lateral continuity despite being deeper than the overlying seismic units.

Two distinct seismic units are observed in the Wealden (Fig. 4.6). The Group is composed of two formations (the Weald Clay and the Hastings Beds, respectively composed by silt and sandstone), the internal reflector *W-purple* that separates the two units is likely to represent the boundary between the two formations. This is consistent with the observation on the seismic data as *W2-purple* corresponds to a change in the acoustic character, which would be due to a variation in the lithological composition, and is therefore interpreted as the base of the Weald Clay and the top of the Hastings Beds. The available BGS stratigraphic column from Dungeness-Boulogne sheet of solid geology map (BGS, 1988) indicates a total thickness of the Wealden Beds of up to 150 m: as the total thickness interpreted here is greater than 100 m, *W-green* could then represent the top of the Hastings Beds. As no wells crossing the Wealden Beds are available and the thickness of the formations is highly variable in this area (Kennedy, 1969; Hamblin et al., 1992), it is only possible to estimate the thickness of this formation where the reflector interpreted as the possible top of the Hastings Beds formation is not covered by the seabed multiple reflection. The base of the seismic unit interpreted as Hasting Beds is not visible in the data.

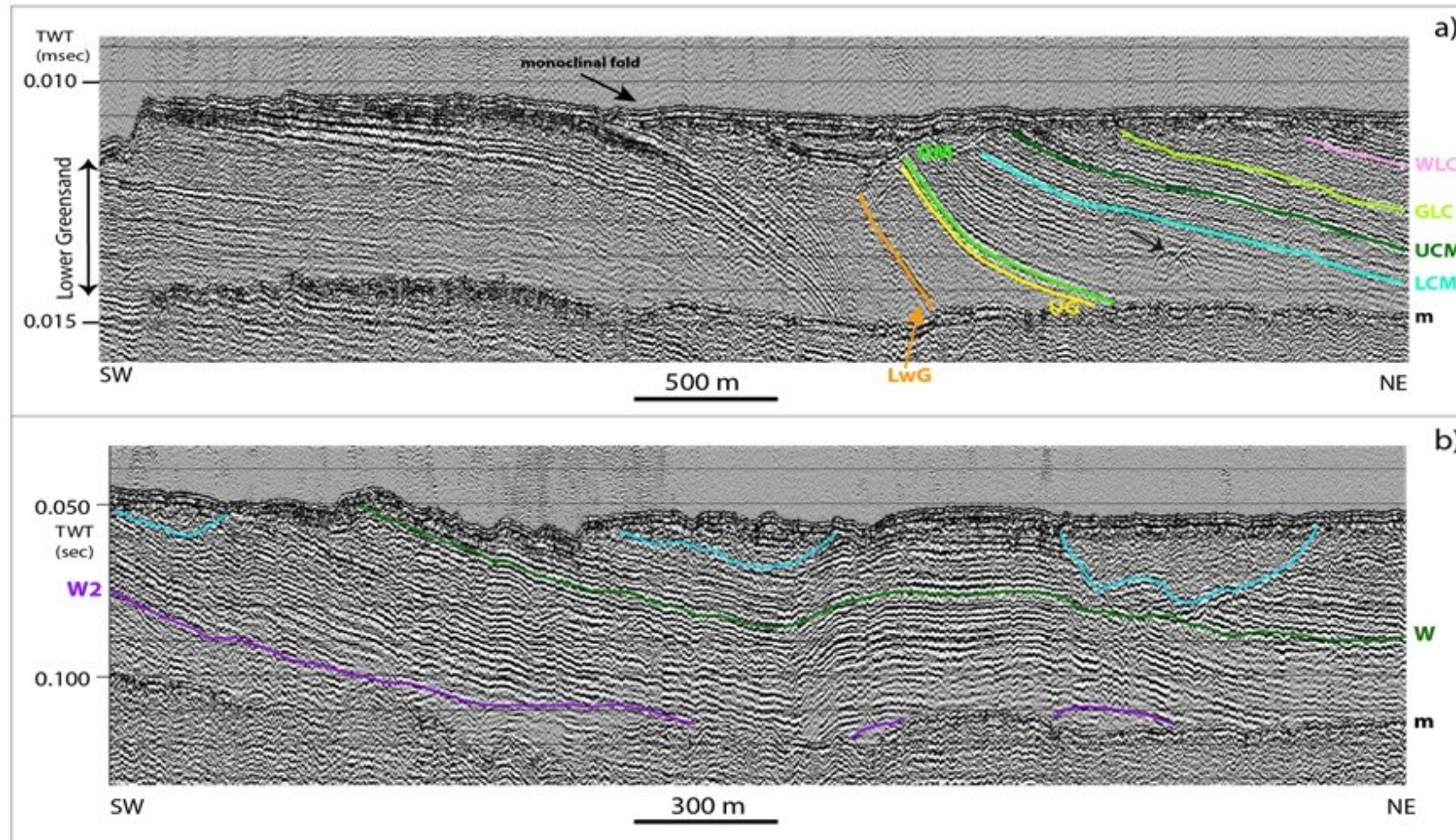


Figure 4.6: a) Seismic line showing the bedrock strata deformed by a monoclinical fold. The reflector on the NE of the fold are the same as in Fig. 4.5, including the black arrow pointing at the Tunnel hyperbolae. b) Geometry of W and W2 representing the top of Wealden Formation and an internal reflector, probably associated to the boundary between Weald Clay and Hastings Beds. Small scale sub-vertical faults and an anticline fold are visible in this section that displays shallow reflectors strongly deformed by the small scale tectonics of the area. W is truncated by sediment-filled incisions of the seabed whose base is marked by the blue reflector. Vertical exaggeration (VE) is 12(a) and 10 (b), m indicates the seabed multiple. The names of the labelled reflectors are the same as in the text. Location of the local seismic lines in Fig. 4.6b.

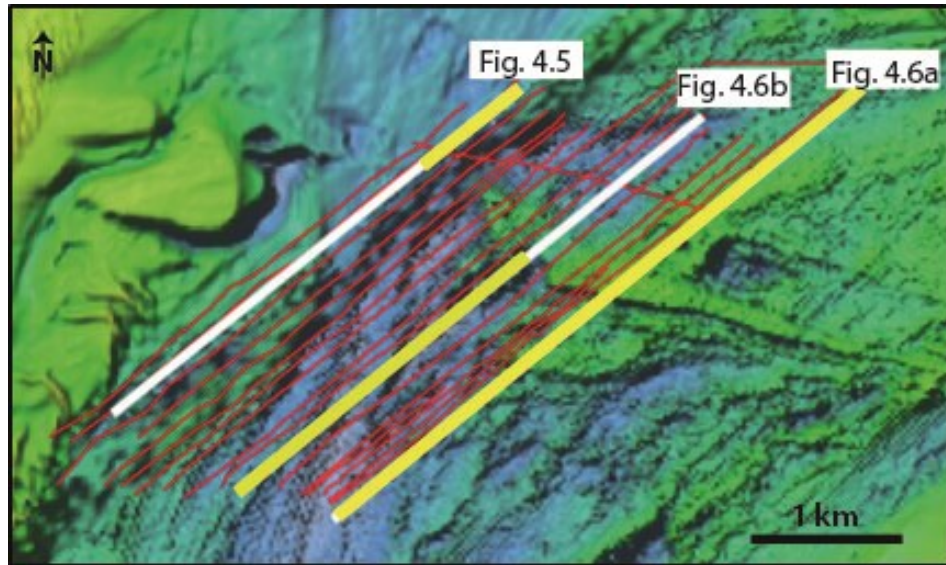


Fig. 4.6b: Location of seismic lines shown in Figs. 4.5, 4.6a and 4.6b. White lines indicate the extent of the reflection seismic line on which the figure is located. The yellow lines indicate the portion of the line shown in the indicated figures.

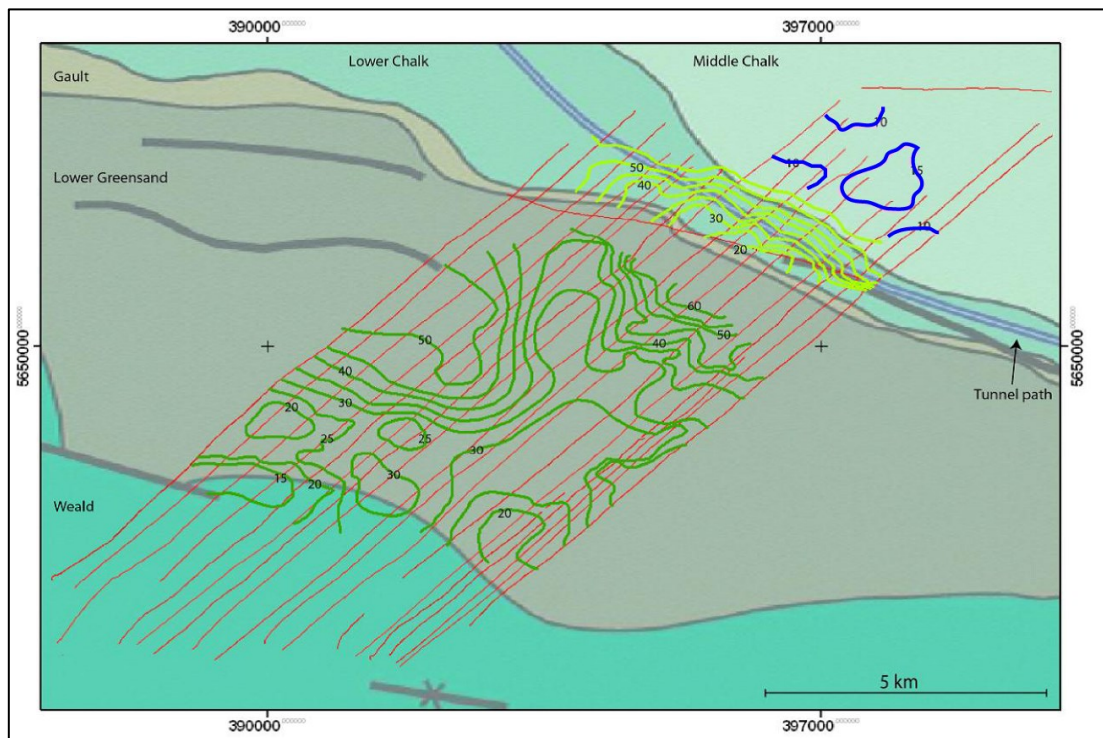


Figure 4.7: Contours of the formation thickness plotted on solid geology map (James et al., 2002). The contour interval is 5 msec TWT for all the formations. The Middle Chalk is represented with blue contours, Lower + Middle Chalk in apple green and Lower Greensand in dark green. The area between the Lower Chalk and the Lower Greensand contours is empty as there is no bedrock present in the upper part of the succession as the area is dominated by the presence of a monocline fold and by a sediment-filled erosional feature elongated parallel to the fold axis (see Chapter 5 for the mapping of the erosional features).

Because of the lack of well data in big part of the reflection seismic dataset, a comparison between the interpreted lithological boundaries and available geology maps (BGS, 1988; James et al., 2002) was made in order to assess the accuracy of the stratigraphic interpretation. In order to do so, the contours of the thickness (in TWT) of the different formations were plotted on the solid geology map edited by James et al. (2002) (Fig. 4.7), with the aim of comparing the location of the interpreted lithological boundaries. The figure shows that the boundary between the mapped formations is not always well detected as erosional features truncate the bedrock reflectors. It was not possible to pick the exact point where the reflectors were outcropping on the seabed. Nevertheless, the figure shows a good correspondence between the two interpretations.

4.3.2 Central Channel Area

The seismic units that characterise the seismic stratigraphy of the bedrock at the Dover Strait are also present on the seismic dataset located in the Central English Channel, making it possible to analyse the seismic facies of the same units on the two different datasets. Nevertheless, due to a much lower quality of the Central English Channel dataset and to the absence of data to calibrate the seismic, the interpretation is only based on the comparison of the seismic acoustic facies. No thickness of the interpreted units were measured.. Despite this, as the dataset is located in the Hampshire and Channel basins, it has been possible to analyse the seismic facies of a wider number of bedrock formations, including those formations that are not present in the Dover Strait area, such as the Palaeogene Beds and the Kimmeridge Clays that were therefore not visible on the seismic in the upstream dataset.

Fig. 4.8 shows the BGS generalized stratigraphic column of the Dover Strait area (BGS, 1988) (centre-right) compared to the stratigraphy interpreted from the Dangeard seismic dataset (right), together with the BGS stratigraphic column of the Hampshire Basin (BGS, 1995) (centre-left) compared to the acoustic facies of the formations interpreted in the Central Channel area (left). With regard to the thickness of the formations, the stratigraphic column of the Dover Strait shows a thickness of the Chalk Group up to 200 m below the seabed in the north-eastern Dover Strait. Moving westwards to the Hampshire Basin and to the younger part of the stratigraphic succession, according to the BGS chart, the Palaeogene rocks are more than 500 m thick in the area where Barton Beds outcrop at the seabed.

Fig. 4.8 also highlights the strong difference in the acoustic facies of the Cretaceous rocks compared to the Palaeogene rocks. The example seismic sections show a strong contrast between the acoustically transparent facies of the Chalk formation and the layered, heterogeneous facies of softer formations such as Palaeogene Beds (in particular the layered sands and clays that compose the Bracklesham Beds) as well as the layered Kimmeridge beds. This strong contrast in texture between the smooth and homogeneous acoustic character of Chalk and the layered Palaeogene rocks, deposited during transgression times over the Chalk eroded surface, allows for mapping of lithological boundaries using both seismic and bathymetry, given the clear seabed expression of lithological boundaries. For this reason, and thanks to the regional extent of the boomer seismic dataset, other than interpreting the bedrock seismic stratigraphy, it has been possible to map the lithological boundaries and structural elements outcropping at the seafloor using bathymetry, integrating the seismic interpretation with the seafloor geological mapping.

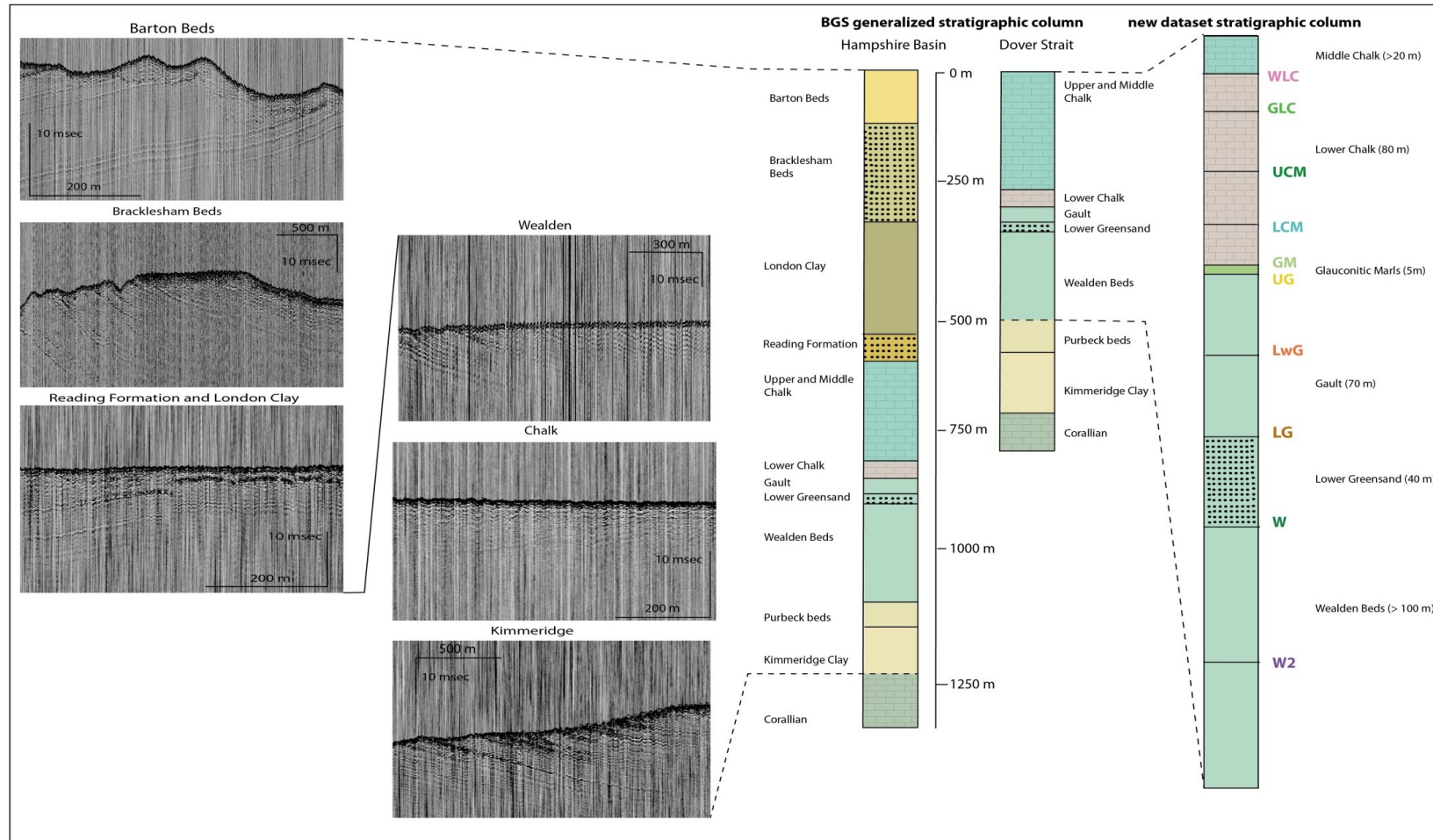


Figure 4.8: Seismic-stratigraphic interpretation of the upper part of the bedrock succession across the study area. The Dover Strait area is represented by the BGS stratigraphic column (center) and by the detailed interpretation performed on the Dangeard seismic dataset (right column). On the left hand side is represented the BGS stratigraphic column of the Hampshire Basin area is shown together with the seismic image of acoustic facies of the bedrock formations present in the Hampshire area. The BGS generalized stratigraphic columns are taken from BGS 1:250000 solid geology maps, sheets Dungeness-Boulogne and Isle of Wight (BGS 1988, 1995). The seismic sections in the figure are extracted from the seismic lines of the Central Channel seismic dataset (see Fig. 4.1).

4.4 Seabed geology from bathymetry

4.4.1 Seabed texture interpretation

The interpretation of the bedrock geology detailed in this section is the result of the integration of the existing geological maps (BGS, 1988, 1989, 1995) and cores and of the interpretation of the seabed texture on the high- resolution bathymetry data (Collier et al., 2006). As shown in Fig. 4.8, the bedrock geology of the seabed is represented by characteristic seismic facies and textures for each formation: the change in acoustic character associated with the change in lithology and visible from bathymetry allows for the mapping of the outcropping geology where the seabed is free or not completely masked by recent sediments. Other than the texture, different resistance to erosion characterises the bedrock formations, determining a variety of morphologies (such as morphological steps or bedrock remnants present at the seabed) that allow for the identification of boundaries between different lithologies, other than tectonic structures. A strong similarity was observed between the acoustic texture of the formations outcropping at the seabed and the outcrop expression of the same formations on onshore outcrops observed in the study area (Table 4.1). Despite the difference in scale between outcrops (with stratification and internal structures ranging from cm to m scale) and bathymetry (with resolvable features usually two orders of magnitude higher), it is clear how the different grade of bedding, stratigraphic packages, weathering characteristics, fractures and variation in slope and rugosity values of the seabed bedrock are visible in both bathymetry and the outcrops (Table 4.1). In addition, structures such as faults and folds, that subdivide the English Channel area into smaller basins, can be detected from bathymetry, often corresponding to formation boundaries and tectonic structures (Collier et al., 2006). Moving from east to west, the study area is subdivided into three basins (Fig. 4.9): the Wealden Basin, the Hampshire Basin and the Channel Basin (Curry and Smith, 1975). The interpreted seabed geology of the three basins is described in this section.

Wealden Basin

The Wealden Basin is distinguished by the presence of the Weald-Artois anticline, clearly visible from bathymetry across the Dover Strait area (Fig. 4.9), as the bedrock formations outcrop at the seafloor on both sides of the fold with boundaries oriented in NW-SE direction parallel to the fold axis.

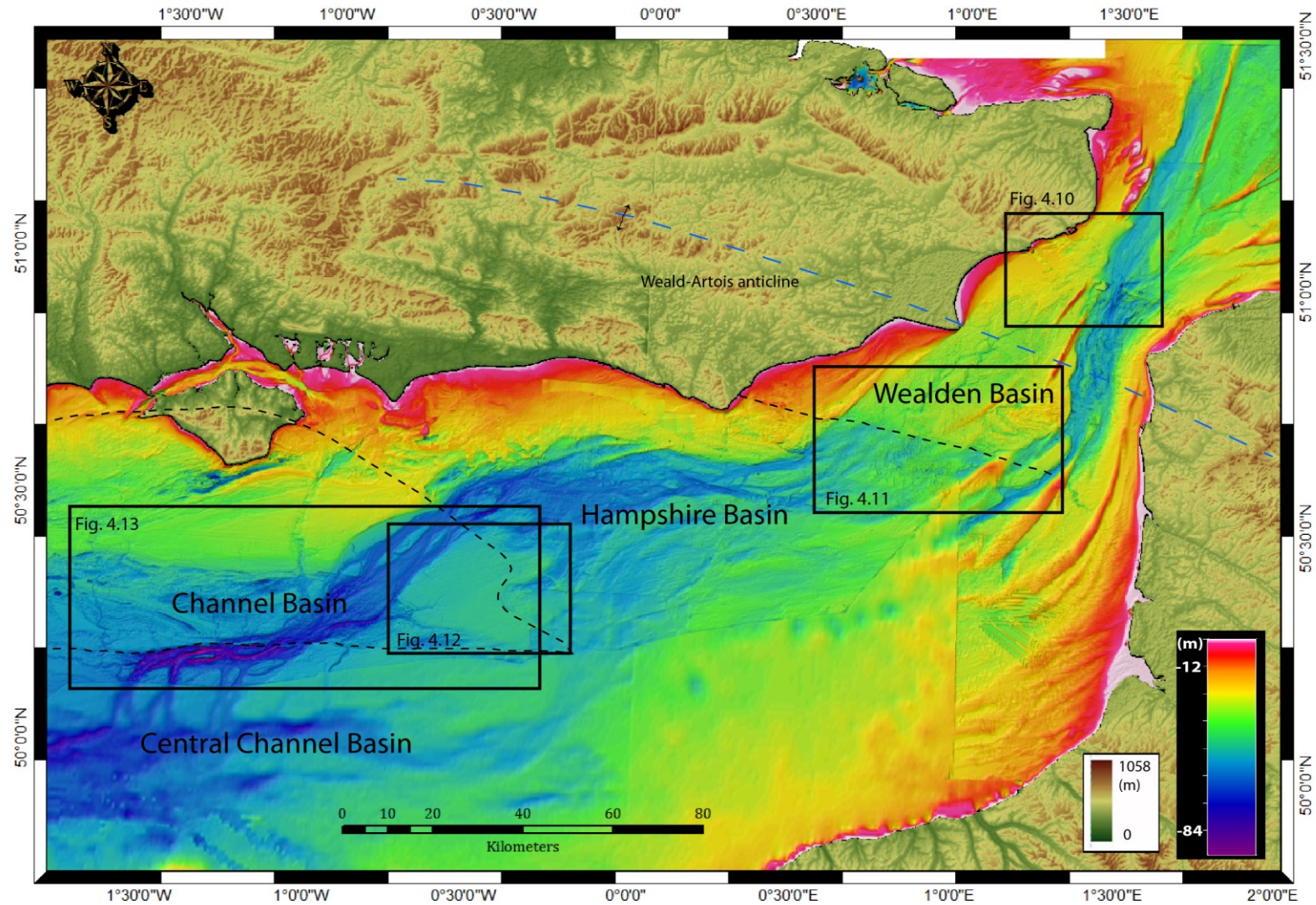


Figure 4.9: Merged bathymetry and onshore SRTM topography (USGS, 2006) displaying the location of the bathymetric images shown in the text. The basin boundaries are indicated by black dashed lines, the axis of the Weald-Artois fold is represented by the blue dashed line.

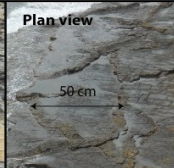
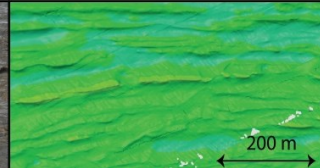
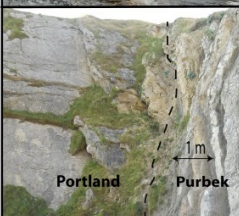
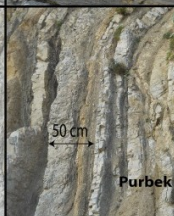
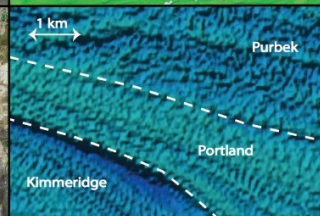
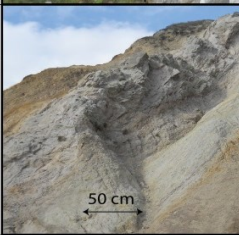
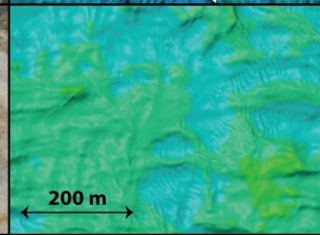
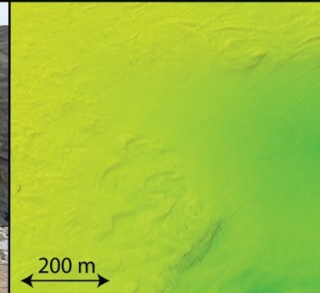
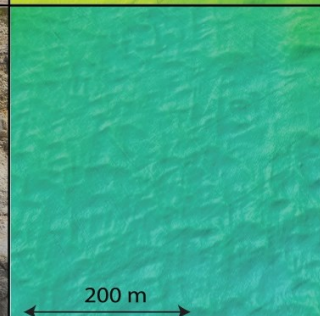
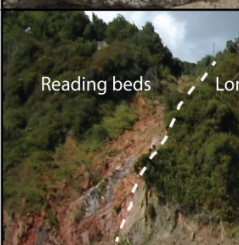
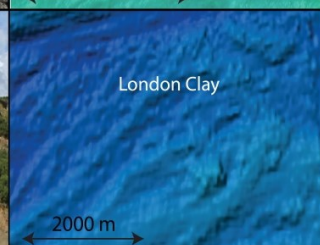
Time	Formation	Outcrop	Seabed texture	Maximum local Slope/Rugosity	
UPPER JURASSIC	Kimmeridge Clay: 10 to 30 cm layers of laminated clays with organic content alterned to dolomitic stone bands				3° 1.007
	Portland: massive limestone Purbek beds: 30-100 cm layers of alternated mudstone/calcite				2° 1.007
LOWER CRETACEOUS	Wealden: fine material deposits with amalgamated sand channel body and pebbles.			1.5° 1.005	
	Greensand: glauconitic sand with chert nodules Gault: clay marine deposits			0.5° 1.003	
UPPER CRETACEOUS	Chalk: shallow marine carbonates with flint layers			0.5° 1.001	
PALAEOGENE	Reading beds: reddish silty clays London Clay: brown grey clays with aragonitic shells. Bracklesham: grey glauconitic clayey sands Barton beds: clays and bioturbated argillaceous sand.			1.75° 1.005	

Table 4.1: Comparison between the outcrop expression of the bedrock formations and their acoustic texture at the seabed. The Jurassic outcrops are located at Kimmeridge Bay, the Cretaceous outcrops are located at Durdle Door and the Palaeogene rocks displayed in the table are located at Whitecliff Bay.

On both sides of the core of the fold outcrop the Chalk Group (Upper, Middle and Lower Chalk), Gault, Lower Greensand and Wealden formations, displaying different dips on the two flanks: steeper dip on the north-east flank and gentler on the south-western flank, as indicated by the bathymetry from which the boundary between the formations can be traced (Fig. 4.10). The contrast in lithology is clearly detectable as the differential resistance to erosion marks the boundaries between the formations that outcrop in the area, in particular on the north-eastern flank where the bedrock Cretaceous formations outcrop in an area of less than 5 km. The contrast between Middle and Lower Chalk, already pointed out in the seismic analyses of the bedrock, is also clearly visible from bathymetric data as the higher erodibility of the Lower Chalk respect to the Upper and Middle Chalk formations determines an up to 10 m high morphological step that is persistent along the entire boundary between Middle and Lower Chalk, both onshore and offshore (Fig. 4.10).

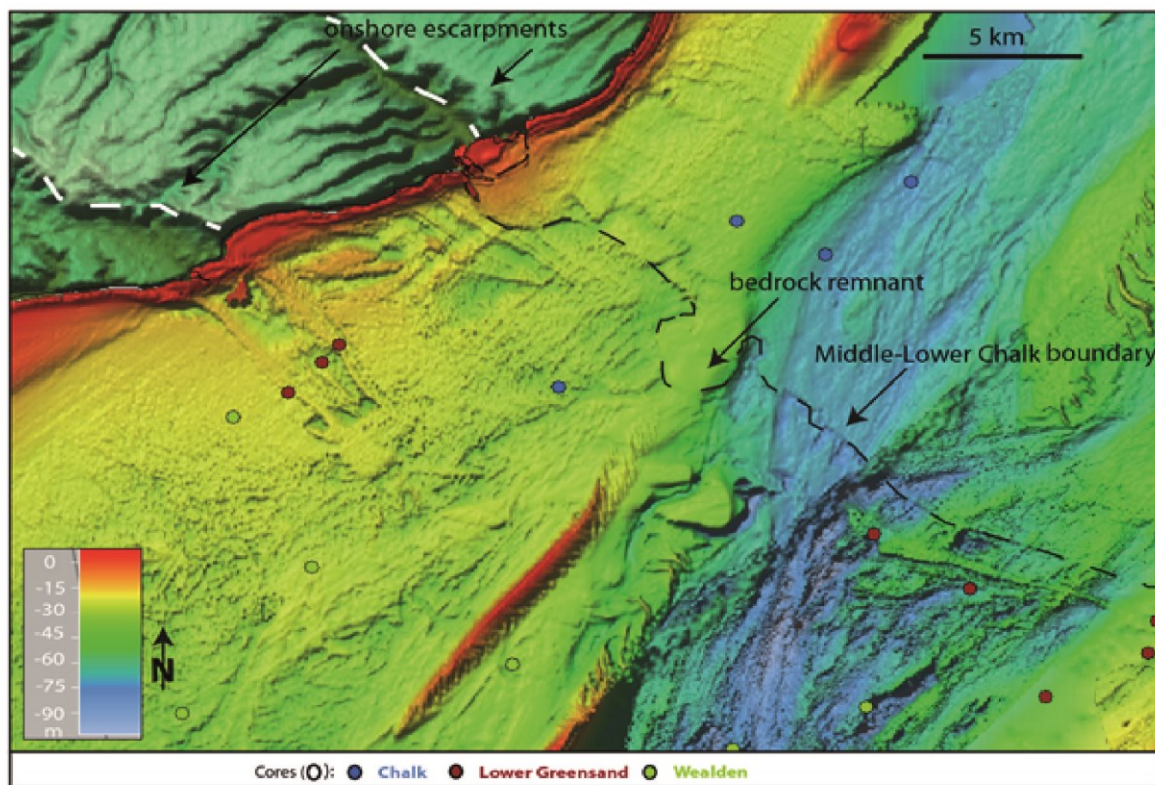


Figure 4.10: Bathymetric image of the north-eastern flank of the Weald-Artois fold, whose axis is located on the south west. The image shows the Middle-Lower Chalk boundary in its onshore (white dashed line) and offshore (black dashed line) expression. The second scarp present onshore (white dashed line) represents the Chalk Escarpment, marking the boundary between the Chalk Group and Gault-Lower Greensand formations. The displayed cores are from the BGS Solid Geology maps (BGS, 1988, 1989, 1995).

This step is a characteristic morphological feature present in the Chalk Group in south-eastern England and known as Chalk escarpment (e.g. Catt and Hodgson, 1976), representing morphological steps formed by changes in composition and/or resistance to erosion of the rocks composing the Chalk. In particular, two marked escarpment are visible in the study area (Figs. 4.9 and 4.10): the first is the Middle-Lower Chalk boundary. The second is located at the boundary between Chalk and Gault Clay/Lower Greensand, locally showing up to 100 m relief onshore, as visible in Fig. 4.9. The boundary between Middle and Lower Chalk formations is also marked by a round-shaped Middle Chalk bedrock remnant that is 3 km long and 2.5 km wide, displaying 15 m relief at the seabed (Fig. 4.10), formed as a result of the higher resistance to erosion of the Middle Chalk Formation respect to the eroded Lower Chalk rocks that surround the remnant. A strong contrast in the texture of the Gault Clay/Lower Greensand formations and Wealden Beds is also visible from the seabed texture: several incisions are carved into the seafloor, up to 7 m deep, in the area where the Gault and Lower Greensand outcrop, parallel to the boundary between Lower Greensand and Wealden.

Moving westwards into the core of the fold where the Wealden Beds outcrop, the acoustic character of the seafloor becomes very heterogeneous due to the mixed composition of the Wealden Beds. This formation is in fact composed of fluvial to deltaic deposits with sand channels amalgamated in the highly erodible mudstone (e.g. Lake and Karner, 1987), resulting in a rough and irregular texture of the seabed. The wide area where the Wealden Beds outcrop extends for 45 km in core of the anticline: the southern portion of this area is characterised by the presence of variably thick layers of surficial sediments locally masking the bedrock (Figs. 4.2 and 4.11). Where the seafloor is sediment-free it is possible to observe another change in the bedrock geology where the Kimmeridge Formation outcrops in correspondence of a small-scale anticline on the south-eastern flank of the Weald-Artois anticline core. The axis of the small-scale fold shows a NW-SE direction, a length of 5 and 10 km (Fig. 4.11) and is clearly visible from the bathymetry as, where high-resolution data is available, a clear change in the texture marks the passage between Wealden and the Kimmeridge.

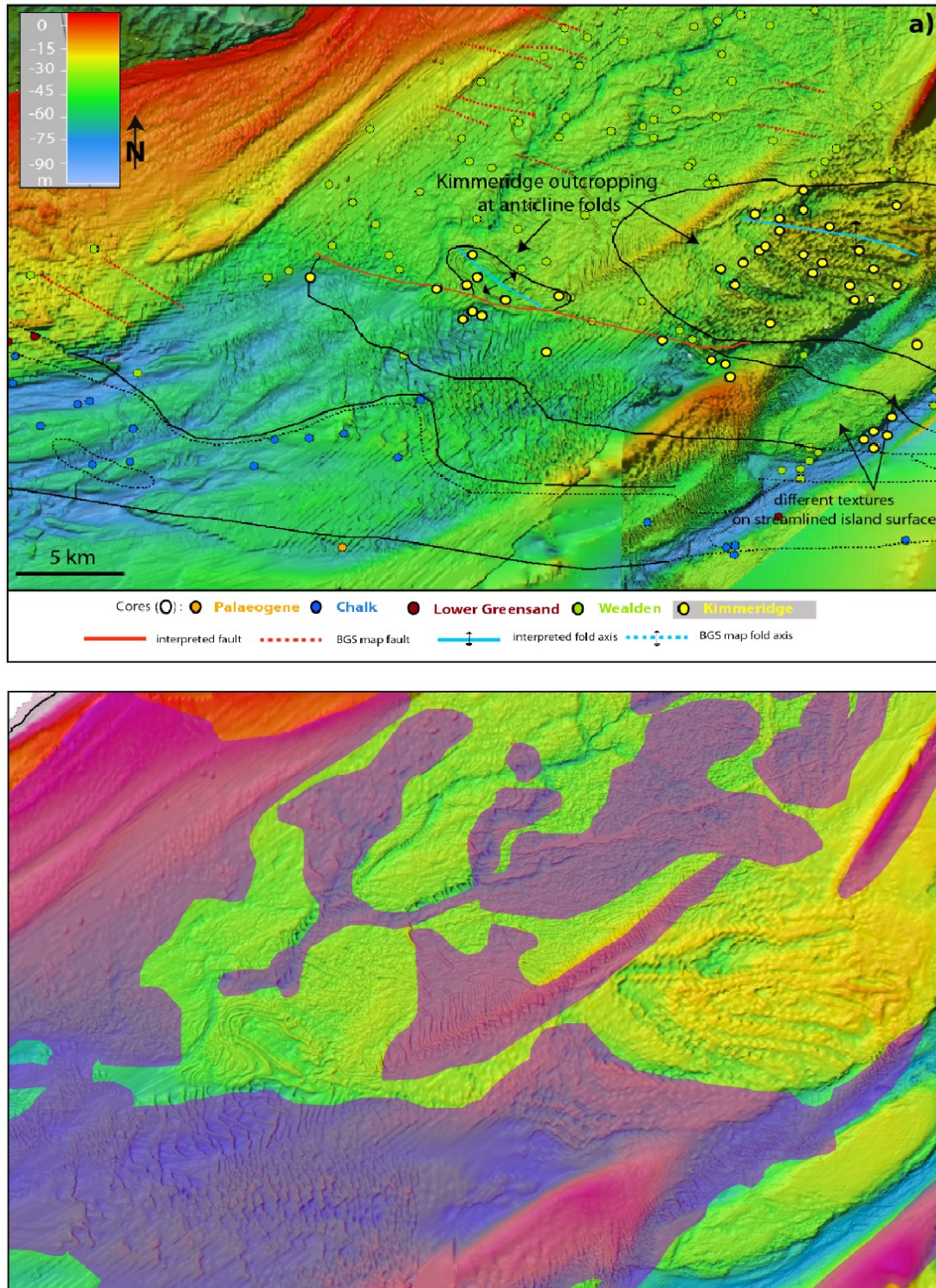


Figure 4.11: a) South-western portion of the Weald Basin showing a series of WNW-ESE oriented faults along the bedrock platform, together with two anticlins with approximately the same orientation. b) Quaternary sediments present as sand banks, sand waves and mega-ripples locally mask the seabed in this sector. The sediment cover is indicated by the purple transparent areas, as shown in Fig. 4.2.

This formation is characterised by a series of folded layers composing a typical trough-and-ridge pattern with ridges of 1.5 up to 3 m amplitude at the seabed and regular spacing (100 to 400 m) between the crests.

The Kimmeridge formation also outcrops on the south of a NW-SE 30 km long fault that is present at the termination of the bedrock platform on which the Wealden Beds outcrop (Fig. 4.11). The fault is marked by a 5 m step between the platform and the adjacent Kimmeridge rocks..

On the south-western portion of the Wealden Basin the seabed geology is represented by the Cretaceous strata of the Lower Greensand, Gault and Chalk outcropping in correspondence of the south-western limb of the Weald-Artois anticline. Here, the Chalk Group displays its typical smooth acoustic texture, characterised by the presence of high amplitude erosional bedforms carved into the seafloor (see Chapter 6), and marking a strong contrast with the adjacent Wealden Beds outcropping on the bedrock platform. Despite this, the boundary between the two formations is not clearly visible on the bathymetry (Fig. 4.11) as the sediment cover becomes thicker in this area that represents a current convergence point (Hamblin et al., 1992). In particular, the seabed is covered by three NE-SW elongated sand banks that, from north to south, respectively measure 13, 15 and 20 km long and are 2 km wide. Groups of 1 to 5 m amplitude sand waves are also present on the seafloor, superimposed on the Wealden Beds and Chalk formation.

Hampshire Basin

The Hampshire Basin is the only area of the English Channel where Palaeogene rocks outcrop at the seabed, representing a structurally elevated area that separates the two adjacent Wealden and Channel basins. The Palaeogene rocks visible at the seabed on the bathymetry are characterised by a typical texture that allows for their identification throughout the whole dataset. As the Palaeogene rocks are characterised by a highly heterogeneous fluvial to marine fine deposits (e.g. Curry, 1962; Keen, 1977; Bone, 1986; Young and Lake, 1988) the alternation of rough and smoother layers is reflected in an irregular acoustic texture, showing highly erodible rocks alternating to harder layers. An example is visible in Fig. 4.12 where a harder layer present in the Bracklesham Formation outcrops at the seabed, forming a 2 to 3 m ridge, also visible on the seismic reflection data. The bedforms carved into the seabed in this area also show a lesser relief at the seabed and a less defined and highly eroded geometry, if

compared to the features present in the Wealden Basin, being more easily eroded and incised by the flows.

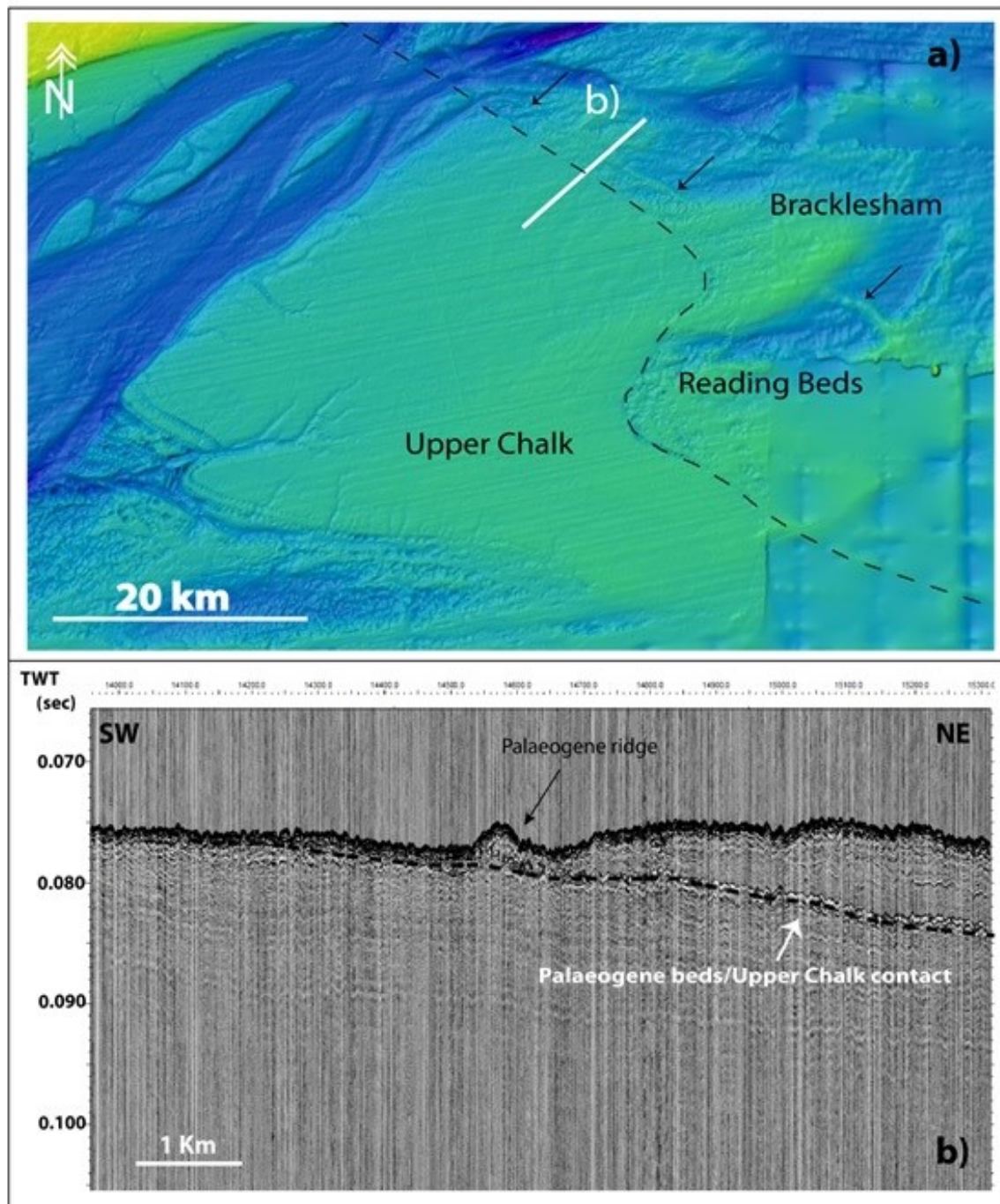


Figure 4.12: Bathymetric (a) and seismic (b) expression of the transition from Hampshire to Channel Basin and from Palaeogene to Chalk rocks outcropping in the two basins. The black arrows in figure (a) indicate a more resistant layer present in the Palaeogene rocks, forming a ridge and dividing the Bracklesham and Reading formations. The noise present in the figure is due to “ringing” phenomena, caused by reverberation of seismic energy in the water column. As very mild de-noise processing was applied to seismic data, a high amount of noise is still present.

A sharp contrast in the seabed texture, marked by an up to 5 m step, is visible at western boundary of the basin, which corresponds to the Wight Fault (Fig. 4.12) (Curry and Smith, 1975; Larssonneur et al., 1975; Lagarde et al., 2003). The fault marks the transition from Palaeogene rocks to the more competent rocks of the Upper Chalk Formation.

Channel Basin

The Channel Basin is characterised by the presence of structural elements (Curry and Smith, 1975) that strongly influence the geometry of the basin and mark its boundaries (Figs. 4.12 and 4.13). Two faults bound the Channel Basin on its eastern and southern margins: these are respectively the Wight Fault and the Mid Channel Fault (Curry and Smith, 1975; Hamblin et al., 1992; Chadwick, 1993; Chantraine et al., 1996). These two structures are visible on the bathymetry as they determine a sharp change in the lithology and consequently in the texture and acoustic character of the seabed. The Wight Fault has a NW-SE trend and is composed of a rectilinear offshore segment extending for 65 km, showing an undulating geometry for 15 km, at the end of which the fault joins the Mid Channel Fault at the south-eastern boundary of the basin. As described for the Hampshire Basin, a strong contrast in the acoustic texture of the seafloor is visible between the smooth texture of the Chalk and the heterogeneous acoustic facies of the Palaeogene rocks outcropping in the Hampshire Basin on the east of the fault. The Mid Channel Fault has a W-E direction and a more regular and rectilinear geometry, extending for 90 km and separating the Channel Basin from the southern Central Channel Basin.

In addition to the faults, two smaller-scale folds also present in this area (Fig. 4.13), marking a number of sharp changes in the lithology in a very restricted area: the bathymetry displays very clearly the change in the texture of the seabed associated to the presence of the folds.

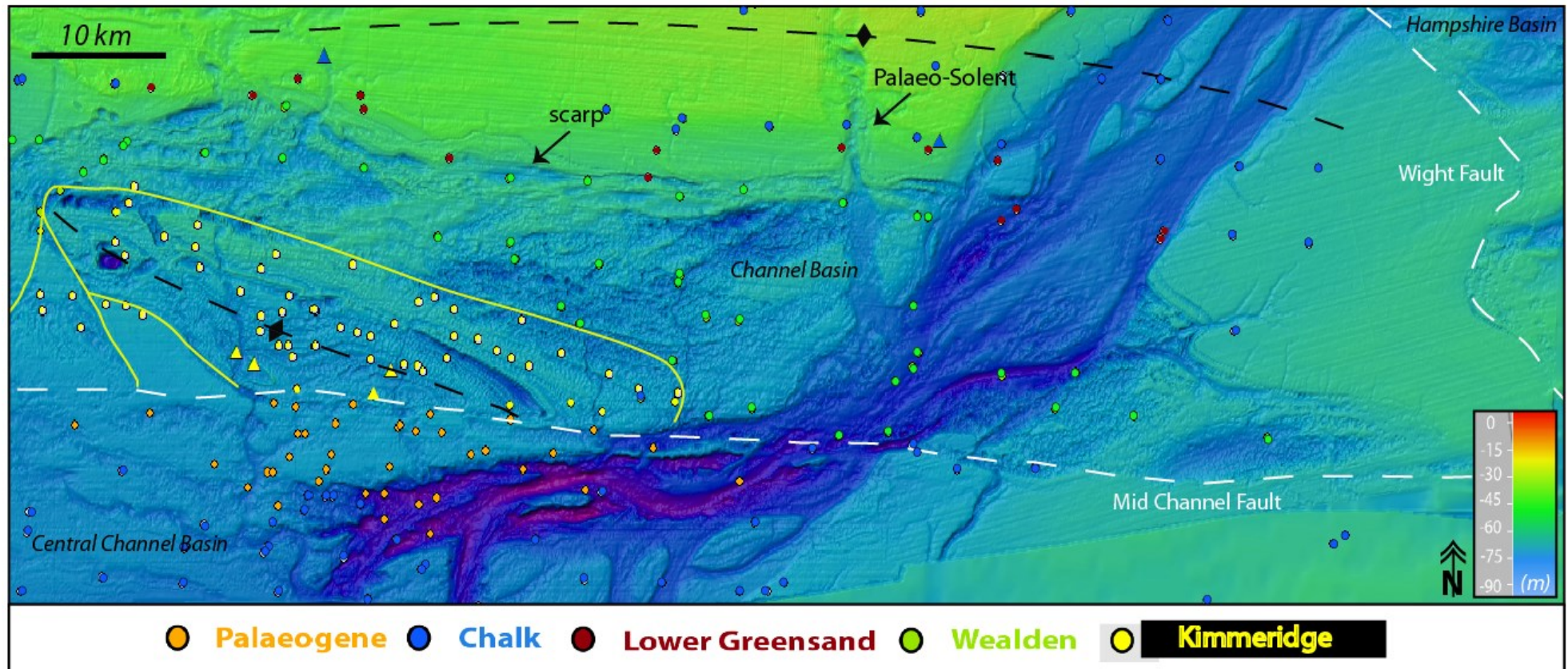


Figure 4.13: Bathymetry of the Channel Basin, showing the eastern and southern boundaries of the Basin bounded by the two faults (white dashed lines). The change in lithology on the seabed is reflected in the different resistance to erosion of the bedforms present on the channel floor. Two folds are imaged in this figure (axes represented by black dashed lines): the southern-most fold composed of the folded Kimmeridge beds, the second fold is located on the north of the basin with the Chalk outcropping at the core. The yellow line represents the boundaries of the areas where the Kimmeridge Formation outcrops.

The southernmost fold displays a NW-SE, 25 km long axis (Figs. 4.13), joining the Mid Channel Fault and determining the contact of the Cretaceous and Jurassic layers, separated by the Palaeogene rocks outcropping on the South of the Mid Channel Fault. The fold axis is clearly visible from the bathymetry as the Jurassic strata along the axis are strongly folded, showing a variable strike that ranges from 110° to 170° (Collier et al., 2006) and a length of ~35 m. The Jurassic rocks, composed of Kimmeridge, Portland and Purbeck Beds display a heterogeneous texture due to the alternation of clays (Kimmeridge, e.g. Cox and Gallois, 1981), massive limestone (Portland, e.g. Hamblin et al., 1992) and laminated mudstone alternated to carbonate beds (Purbeck, e.g. Arkell, 1947; Whittaker, 1985), resulting in a sequence of sharp, very well defined layers with a ridge and trough pattern. The seafloor shows ridges of 0.2-0.5, up to 5 m amplitude alternated to more erodible layers with a spacing that varies from 10 to 200 m. The erosional bedforms present on the channel floor in the southern area where the Chalk Group outcrops display higher relief, more defined shapes and smooth texture if compared to the geometry and texture of the bedforms carved into the Wealden and Palaeogene rocks (Fig. 4.12).

South of the Isle of Wight, another fold juxtaposes the Upper Cretaceous Chalk and Wealden Beds to the Lower Greensand: their contact is marked by a 55 km long scarp (Fig. 4.13) with up to 15 m relief visible on the bathymetry. The fold axis has a curved geometry and a WNW-ESE direction, running almost perpendicularly to the direction of the Palaeo-Solent River present in this area (Hamblin et al., 1992).

Rugosity and slope maps were calculated from the bathymetry to aid the interpretation and mapping of boundaries between Palaeogene, Cretaceous and Jurassic rocks, as well as the structural lineations mapped in the Channel and Hampshire basins (Fig. 4.14). Rugosity and slope attributes are strictly related and typically correspond to low values for smooth and homogenous surfaces and higher values for irregular and rough surfaces (e.g. Fonseca and Mayer, 2007; Wilson et al., 2007). Slope, or the measure of steepness, was derived using Fledermaus Slope tool. Output slope values are derived for each cell as the maximum rate of change from the cell to its neighbor. This is calculated by taking the steepest slope between each cell in the DTM and its 8 nearest neighbors. Rugosity describes topographic roughness with a surface area to planar area ratio. Rugosity was derived with the Fledermaus Rugosity tool using a 3x3 neighborhood analysis to calculate surface area based on a 3D interpretation of cells values. Low rugosity values indicate flat, smooth locations; higher values indicate

areas of high-relief. Rugosity calculated using this technique is highly correlated with slope. The highest rugosity values show a relationship with the high slope and lower rugosity with low slope.

The maps show that areas dominated by the Chalk Group display much lower slope and rugosity values respect to the areas where more layered and heterogeneous formations, such as Wealden and Kimmeridge, outcrop (as also shown by Table 4.1). The entire area dominated by the Chalk displays slope values lower than 2° with the only exception of those areas that represent the channel flanks or bedrock remnant margins.

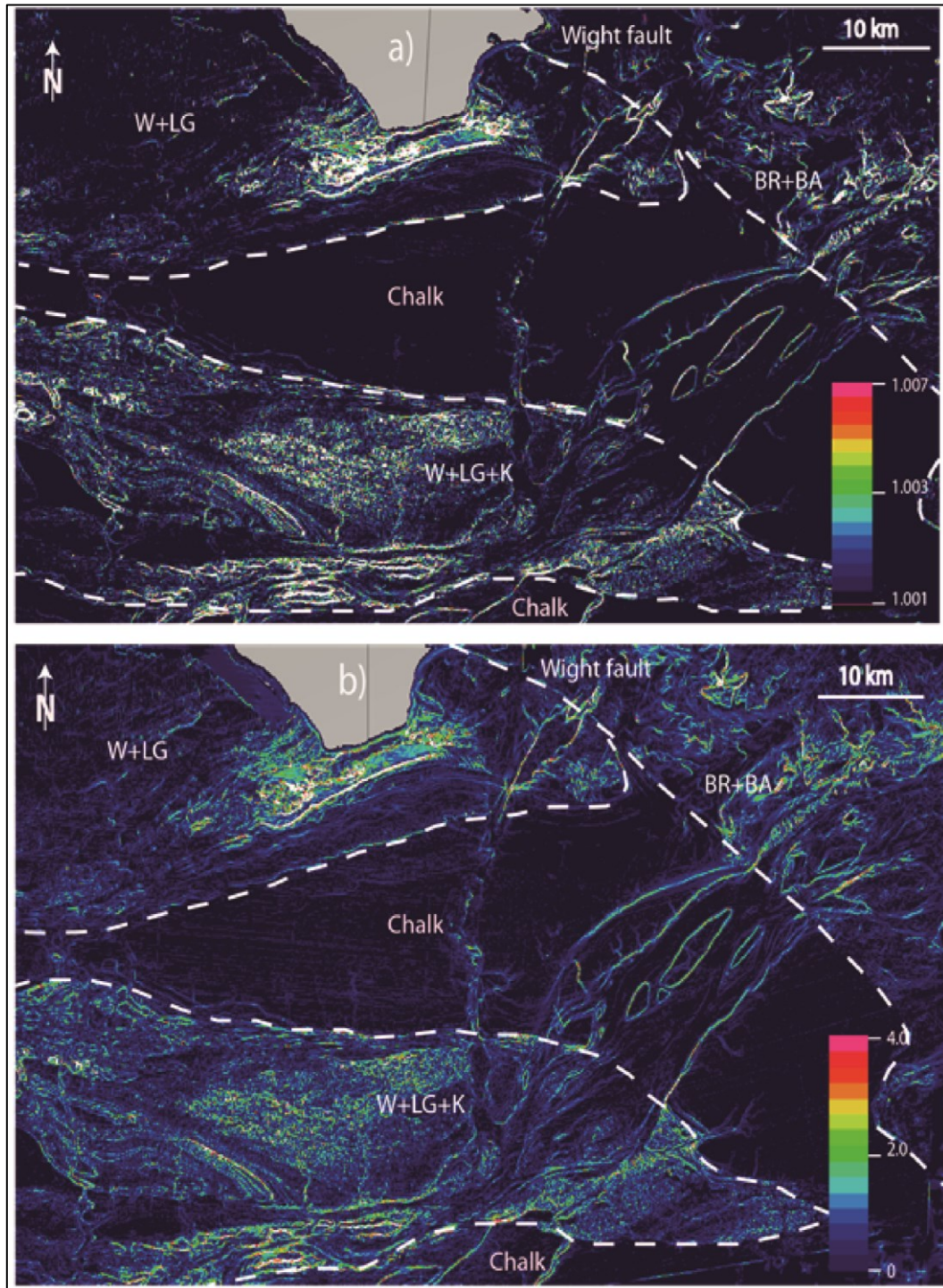


Figure 4.14: Rugosity (a) and slope expressed in degrees (b) calculated on the 30 m bin size bathymetry. The white dotted lines represent the limits of the central area where the Chalk outcrops, as indicated by the low values of both rugosity and slope. The outcropping formations are labelled as follow: BR (Bracklesham), BA (Barton), W (Wealden), LG (Lower Greensand), K (Kimmeridge).

Conversely, the areas dominated by heterogeneous lithologies show slope values up to 4° , as shown by Fig. Fig. 4.14 where high slopes values are represented by green and red colours.

Finally, the two obtained maps also image the structural elements outcropping at the seabed in this area: faults at the seabed are marked by a sharp lineation showing high slope values. Folds are also visible: an example is given by the folded Jurassic layers displaying high values along the deformed ridges in correspondence of the fold located at the south-western margin of the basin.

4.5 Bedrock geology map

Following the interpretation of the bedrock stratigraphic units, their distribution at the seabed, and the mapping of the main tectonic elements, a map of the bedrock geology was drawn (Fig. 4.15), illustrating the interpreted lithological boundaries, faults and folds. The new map was produced merging the interpretation of the new datasets to the existing BGS geology maps (BGS, 1988, 1989, 1995; James et al., 2002) and cores (that were used in those areas where it was not possible to map using the new datasets only, see Section 3.3 for details). In order to distinguish between the BGS and the new interpretation, the new interpreted boundaries and tectonic features are represented in the map with a solid line while the features previously interpreted by BGS, and not visible from the bathymetry, are represented with a dashed line. The map shows how the big scale tectonic structures and the main lithological boundaries present on the British side of the Channel were mapped using the available bathymetry, with the exception of those areas where the sediment cover did not allow for the mapping. Conversely, in the French sector of the Dover Strait, the low resolution of the available bathymetry did not allow for the identification of new lithological and tectonic features and the available BGS geologic maps were used to draw the boundaries in this area.

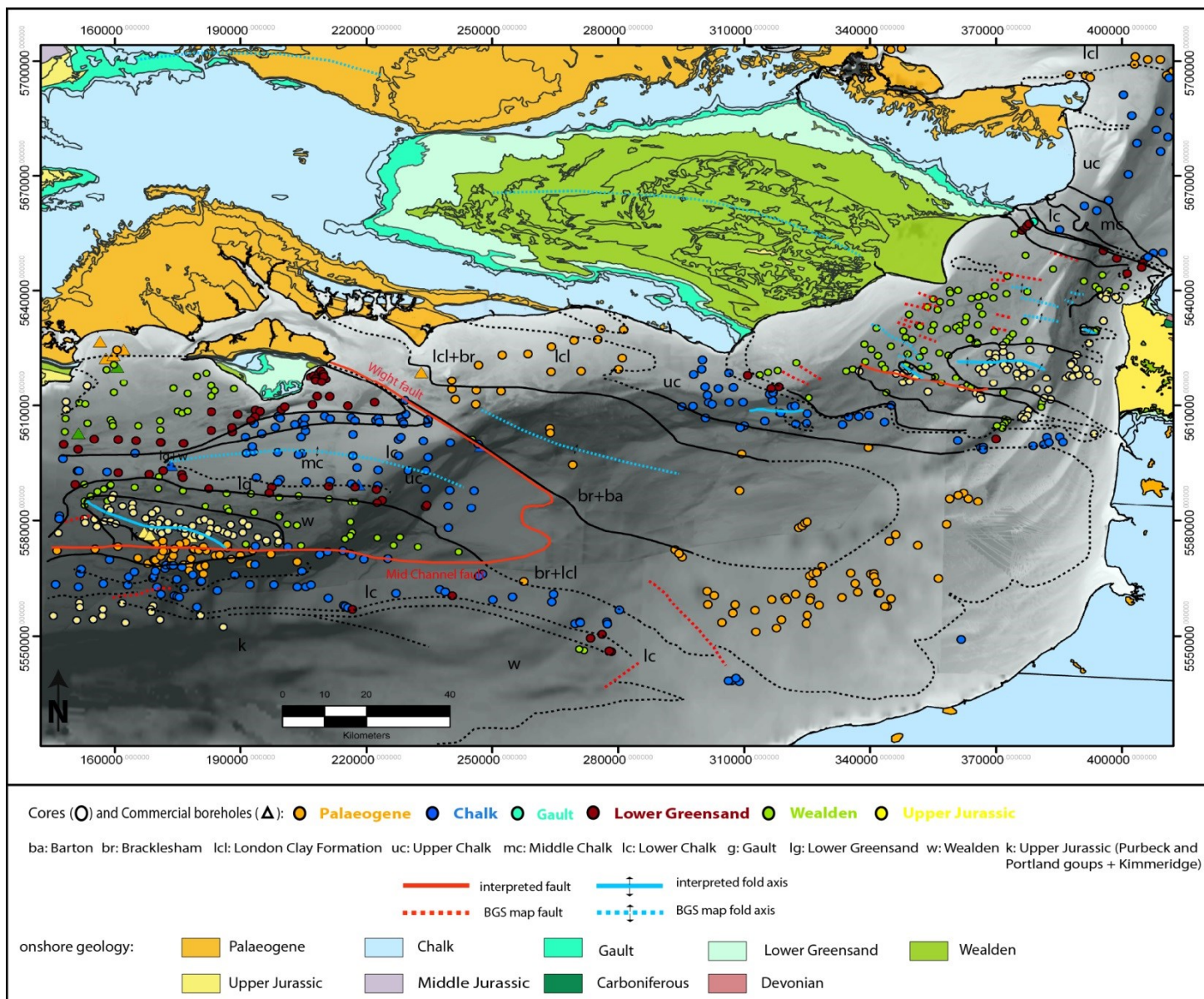


Figure 4.15: Final geological map of the study area. Dashed lines represent boundaries from BGS geology maps (BGS, 1988, 1989, 1995).

4.6 Conclusions

The seismo-stratigraphic analysis of the upper part of the bedrock succession, together with the geological mapping of the seafloor, show that regional geology defines many the features present at the seabed. It is therefore crucial to keep into account the influence of the lithology and tectonics when performing geomorphological analyses on the flood-related bedforms present in the study area. In addition, the surficial sediment map shows that more than 80% of the seabed is sediment free, allowing for the geological mapping of the bedrock from bathymetry augmented with seismic reflection data. With regard to the distribution of the surficial sediments, the mapping of the sediments from bathymetry and the comparison of their distribution to shear stress at the seabed (Mitchell et al., 2010) indicate that the surficial sediment distribution is related to tide-controlled shear stress, particularly in the Dover Strait area.

Finally, the seismic-stratigraphic and bathymetric analyses revealed a close relationship between seismic acoustic facies, seabed acoustic texture and onshore outcrop facies of the bedrock succession. The different level of layering, fractures, compaction and weathering of each formation determines a different resistance to erosion and therefore a variation in the geometry of the bedforms at the seabed. This is also reflected by the variation in slope and rugosity values calculated on the bathymetry, indicating that slope and rugosity can be used as an additional attribute for the mapping of lithological and tectonic boundaries from bathymetry.

5 . The formation of the Fosse Dangeard

5.1 Introduction and background

The Fosse Dangeard is composed of a series of high-relief incisions carved into the bedrock and scattered across the Dover Strait. These incisions are key-features in the interpretation and reconstruction of the breaching of the Dover Strait, proposed as the megaflood spill-point. The Fosse was discovered during Channel Tunnel investigation works and first interpreted by Destombes et al. (1975) who presented a map of the depressions, interpreted as interconnected bedrock incisions formed by glacial processes. Although after the mapping of the Fosse no further investigations were conducted, throughout time, Destombes interpretation was followed by different hypotheses proposed by other authors. Tidal, fluvial and tectonic mechanisms were suggested as responsible for the formation of the Fosse (Kellaway et al., 1975; Hamblin et al., 1992; Lericolais et al., 2003; Van Vliet-Lanoe et al., 2004). Despite different theories being proposed, a lack of direct evidence did not allow for a final interpretation of the process that formed the depressions.

The availability of new high-resolution seismic data, located in the French sector of the Dover Strait, where the Fosse was mapped by Destombes in 1975, allows for detailed analyses on the geometry, infill and distribution of the incisions, giving insights on the processes that led to the formation and filling of the depressions. Firstly, the morphometric characterisation of the incisions allows for the analysis of their geometrical pattern and spatial distribution, with the aim of comparing the incisions to similar features described in literature. Secondly, the seismic stratigraphy of the incision infill and the mapping of internal erosional surfaces permits interpretation of processes that carved and filled the incisions. Finally, these results will make it possible to discuss the role of the incisions in the megaflood history framework, with the aim of understanding the relative timing of the Fosse Dangeard formation in relation to the breaching of the Dover Strait and to the English Channel megaflood history.

5.2 Aims and Objectives

The aim of the work described in this chapter is the characterization of the Fosse Dangeard, located in the Dover Strait.

In order to do so, there are five main objectives:

- To map and characterise the bedrock surface, in order to constrain the geometry and morphology of the Fosse, including the dimensions and spatial distribution of the incisions.
- To determine the morphometric properties of the incisions in order to quantitatively characterise their morphology and enable comparison with similar features identified from the literature.
- To analyse and interpret the acoustic character and seismic stratigraphy of the infill and identify internal erosional surfaces, particularly with respect to the relative timing of erosion and deposition.
- To determine the influence of regional geology on the morphology and spatial distribution of the Fosse Dangeard.
- To determine the origin of the incisions with relation to the evolution of the Dover Strait, with particular interest in the role of the incisions in the megaflood event hypothesis.

These objectives will permit the following questions to be answered: (1) what are the mechanisms responsible for the formation of the incisions, (2) What factors control the morphology of the basal surface and the spatial distribution of the incisions? (3) What is the composition of the sediments in the infill and what is the relative timing of deposition? (4) Is the morphology of the Fosse Dangeard consistent with the megaflood hypothesis?

5.3 Bedrock acoustic facies and structure

5.3.1 Bedrock units

The analysis of the bedrock seismic stratigraphy described in Chapter 4 enables us to distinguish 10 seismic units that correspond to Cretaceous-age stratigraphic units present in the Dover Strait area (Fig. 5.1). The base of each unit is associated with a distinct seismic reflector (see Chapter 3 for seismic data calibration methods and Chapter 4 for detailed description of the bedrock seismo-stratigraphic units), as shown in Table 5.1.

Unit	Reflector name	Acoustic facies
Middle Chalk	<i>WLC-pink (base)</i>	Internal high amplitude reflections and small-scale discontinuities and diffractions
White Chalk	<i>WLC-pink</i>	Internal layers with sub-parallel geometry. Strongly folded in correspondence of the monocline fold where dip sharply increases
Grey Chalk	<i>GC-apple green</i>	
Upper Chalk Marl	<i>UCM-dark green</i>	
Lower Chalk Marl	<i>LCM-light blue</i>	
Glauconitic Marl	<i>GM-green</i>	Couple of very high amplitude reflections
Upper Gault	<i>UG-yellow</i>	Transparent acoustic character
Lower Gault	<i>LwG-orange</i>	
Lower Greensand	<i>LG-brown</i>	Internal stratified high amplitude reflections alternated to almost transparent layers
Weald Clay	<i>W-green</i>	Sub-parallel undulated geometry, faulted by sub-vertical small scale faults and folds
Hastings Beds	<i>W2-purple</i>	Heterogeneous seismic character due to the mixed lithology

Table 5.1: Summary of the interpreted seismo-stratigraphic units and reflectors representing their top, with the exception of the Middle Chalk whose base is WLC-pink.

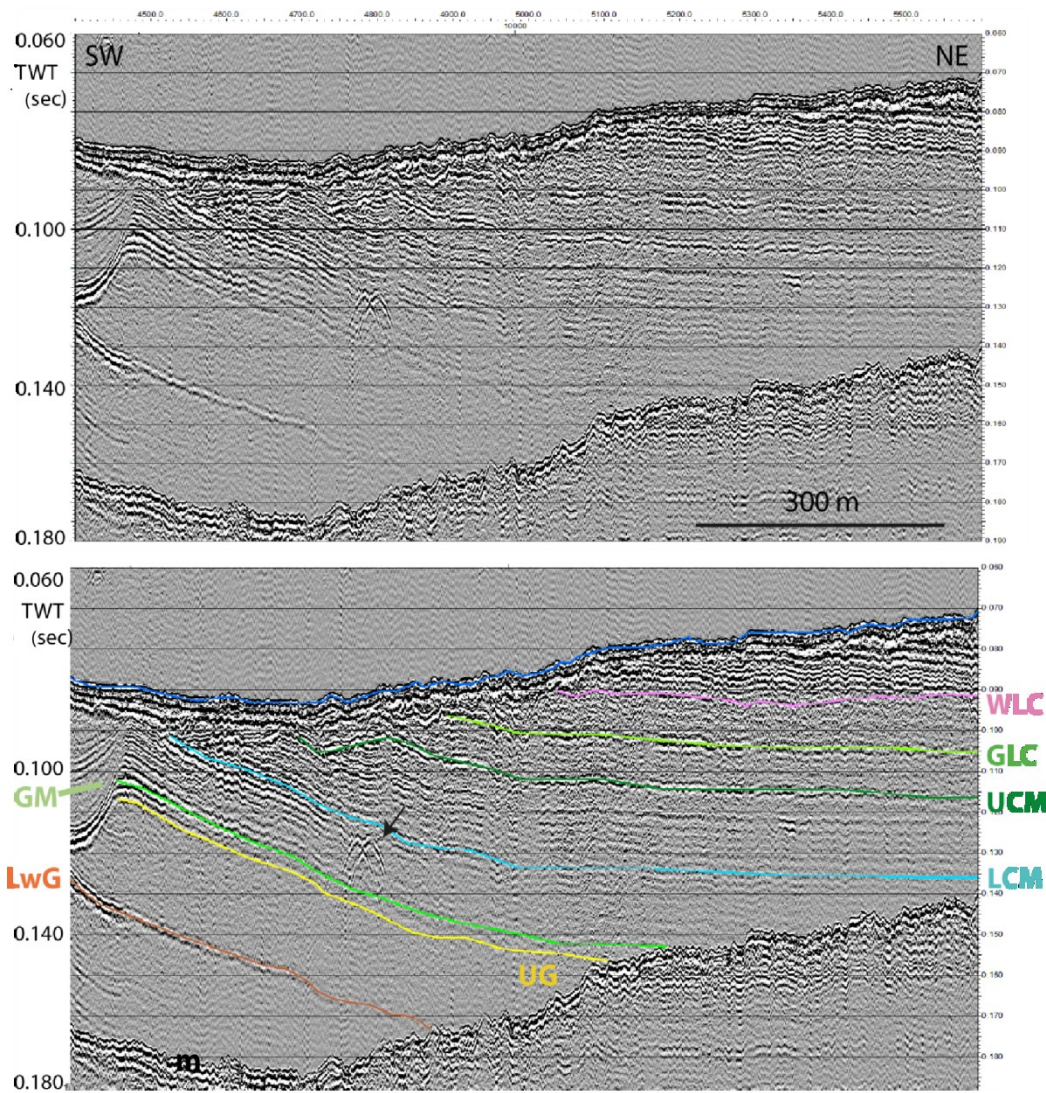


Figure 5.1: Seismic stratigraphy of the Dover Strait, showing the upper part of the Cretaceous succession with reflectors labelled as in Table 1. The black arrow indicates the Tunnel diffractions (see Chapter 4 for details). Vertical exaggeration is 4. See Fig. 5.18 for location of the seismic line.

5.3.2 Tectonic features

A line drawing of three interpreted seismic reflection lines across the dataset is shown in Fig. 5.2, summarising the geometry of the bedrock strata, the character of the tectonic features present in the area and their influence on the bedrock geology. In particular, the interpreted bedrock geology and structural elements are used in this chapter in order to understand the spatial and geometrical relationship between the Cretaceous strata, the monocline fold and fault and the incisions that cut the bedrock units.

The Cretaceous strata are characterised by the presence of small-scale faults and folds (displacement of 10 to 20 m) that determine a local variation in the dip of the strata. The only major structural feature recognised in the dataset is a monoclinical fold with NW-SE trending axis associated to a normal fault, dipping 11° towards north-east (Figs. 5.1 and 5.2); both structures are NW-SE oriented. The structure orientation follows the regional structural pattern of WNW-ESE trending faults, parallel to the orientation of the Weald-Artois anticline axis (See Section 4.5).

As shown in Figs. 5.1 and 5.2, the presence of the monoclinical fold in the study area strongly affects the local dip of the strata. In particular, the Lower Greensand and Weald formations display a dip that varies from 0° - 2° to 10° where the strata are folded and faulted in correspondence of the monoclinical fold. In the same area, the Gault Clay dip varies from 3° to 11° while the Chalk, almost flat on the north-eastern part of the dataset, dips 11° where folded. The presence of the fold also determines a lateral contact between the Gault and the Lower Greensand dipping strata as the displacement caused by the fault brings the Gault Clay, on the hanging-wall, in contact with the Lower Greensand Formation present at the footwall of the fault (Fig. 5.2).

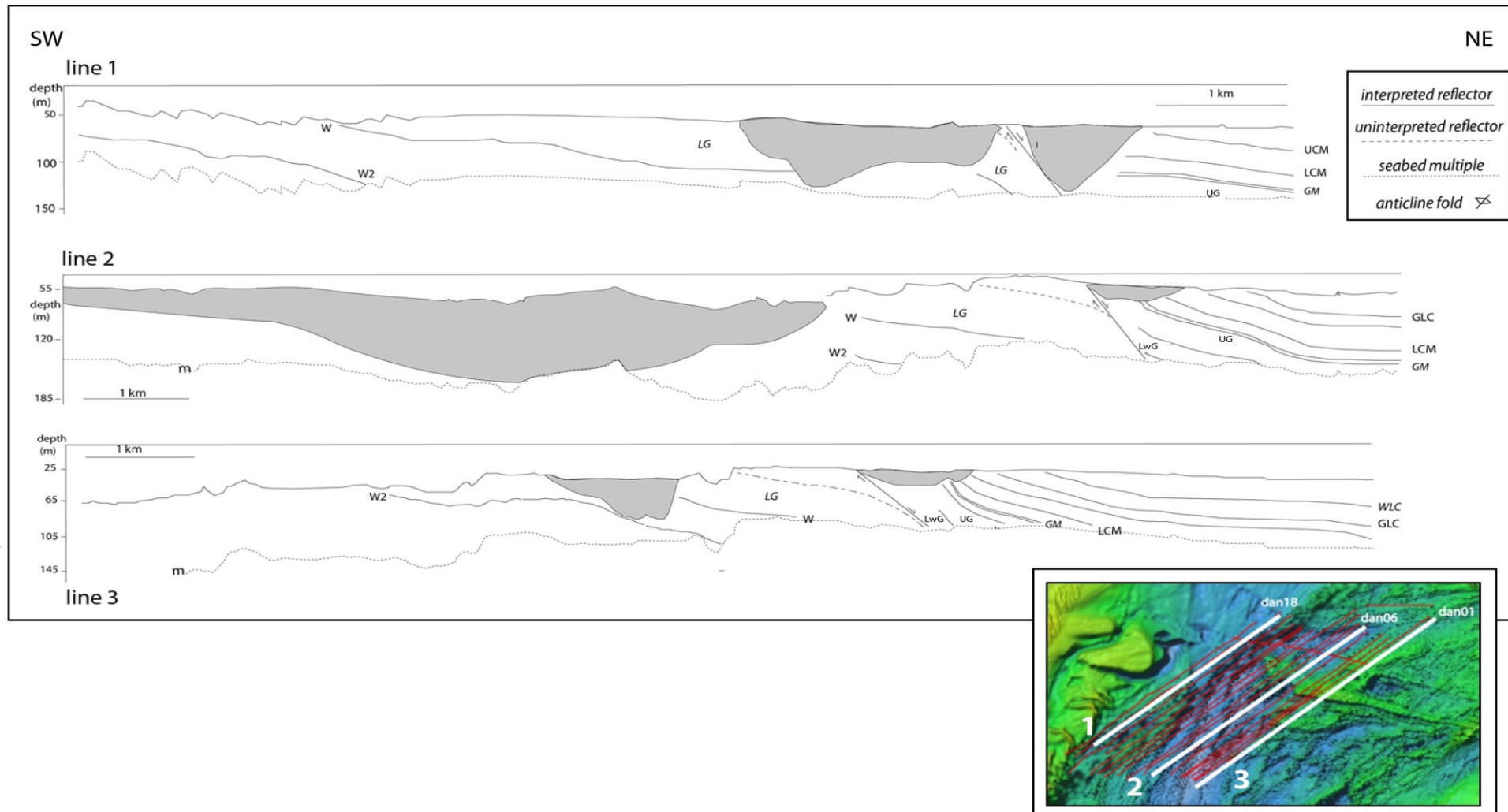


Figure 5.2: Line drawing of three seismic lines across the dataset showing the geometry of the main bedrock reflectors (labelled as in Table 1 in this chapter), the main interpreted structural elements and the extent of the Fosse Dangeard incisions (grey areas) that truncate the bedrock reflectors. The location of the lines used for the line drawing is indicated in the box, showing the seismic dataset with bathymetry in the background. Vertical exaggeration is 7. The letters relative to the names of the bedrock formations are the same as in Table 5.1 and Fig. 5.1. Dan01, 06, 18 indicate the original name of the lines of this dataset.

5.4 Characterization of bedrock incisions

5.4.1 Characterization of basal erosional surface (BES)

The study area is characterised by the distinct incisions eroded into Cretaceous bedrock. These depressions make up the Fosses Dangeard (Fig. 5.3).

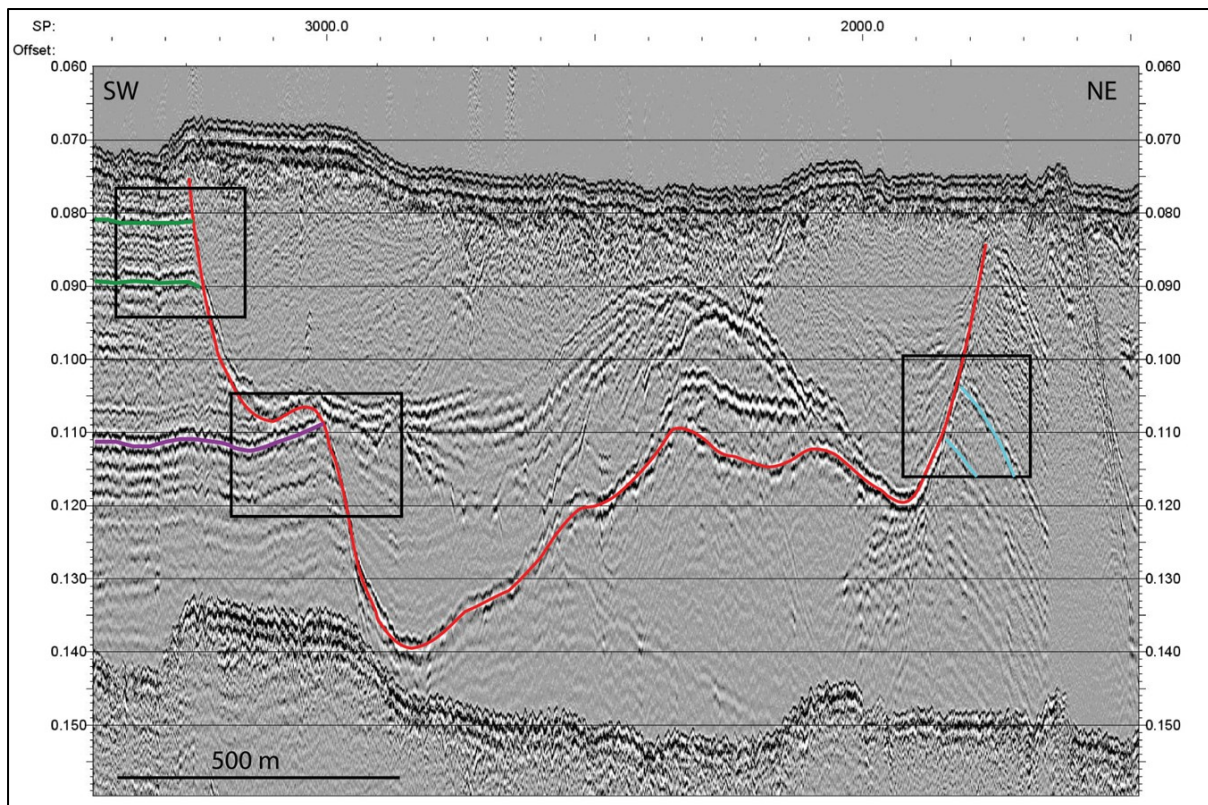


Figure 5.3: Seismic line displaying the sharp erosional contact between BES (red line) and the bedrock reflectors (for letters indicating the names of reflector see Table 5.1) on both sides of one of the incisions. The black boxes indicate the erosional truncation of the bedrock reflectors. The incisions display a basal surface characterised by an undulating geometry composed of more than one scour. VE is 15. See Fig. 5.19 for location of the seismic line.

The reflector that bounds the incisions represents their basal erosional surface and has been therefore labelled as BES (Basal Erosional Surface). BES is characterised by high amplitude, marking a sharp discontinuity with the bedrock reflections and abruptly cutting the underlying Cretaceous strata. The geometry of this surface varies throughout the dataset but the seismic character of the reflector remains constant. In the north-eastern part of the dataset the reflector forms a surface that displays a regular U-shaped geometry which, moving south-westwards, becomes less regular..

Locally, the basal erosional surface is defined by a more complicated geometry and, occasionally, more than one surface, as for the case of the incision imaged in Fig. 5.3. The variation in the geometry of the basal erosional surface and in the morphology of the incisions throughout the dataset is described in the following section through three example seismic lines.

5.4.2 Regional view of the incisions

Three seismic lines were selected across the dataset in order to describe the variation in the morphology of the incisions and their cross-section morphology on a regional scale. In order to identify the incisions described in the text, the features were named using letters from A to D.

In particular, Fig. 5.4 shows a NE-SW-oriented seismic line located at the Dover Strait and parallel to the Strait axis. The seismic line displays two large-scale incisions that sharply cut bedrock reflectors. In the north-eastern part of the line, incision A is 400 m wide and up to 80 m deep. The incision basal erosional surface has a U-shaped geometry and marks steep, almost symmetrical flanks of the incision carved on the northern flank of the fold into the Middle and Lower Chalk and Gault formations. The slope value of the incision flanks is respectively 18° and 15° for north-east and south-west flanks of the incision. The incision displays sediments completely filling it: the infill is characterized by a semi-transparent acoustic facies with two internal high amplitude reflectors truncating the infill, representing two erosional surfaces. Fig. 5.4 also shows that, where the Gault and the Lower Greensand formations are in lateral contact due to the tectonic deformation, a ridge composed of the more competent Lower Greensand rocks follows the axis of the monoclonal fold, outcropping at the seabed and displaying amplitude of up to 5 m.

On the other side of the fold, a second incision (B) is present, up to 60 m deep and 800 m wide. Its basal erosional surface has an irregular, complex geometry, being composed of more than one truncated erosional surfaces that display the same high amplitude acoustic character of BES. The incision is characterised by a steep NW margin (15°) that corresponds to the fold north-eastern flank, while the less steep south-western (downstream) flank is, on the other hand, not controlled by any structures and displays a gentler slope that varies from 4° to 1° moving downstream. These flank slope values represent a pattern that is observed

in most of the seismic lines where the incisions are present: a deeper upstream flank and a gentler downstream flank displaying two slope values. The variation in the slope pattern seems to be entirely due to variations in flow that carved the incision. No other factors such as lithology or tectonic elements control the morphology of the downstream flank of the incisions as no clear lithological or structural boundaries affect the morphology of the depression downstream flank. Fig. 5.4 shows that no lithological or structural element controls the slope pattern of incision B, carved into the Lower Greensand and Weald formations on both flanks and not bounded by any structural features on its downstream south-western flank.

Fig. 5.5 shows two other isolated incisions, located south-west of the seismic line displayed in Fig. 5.4. The narrow and steep incision (labelled as incision D) on the north-east of the line is bounded by a U-shaped basal surface that truncates the bedrock reflectors. Incision D is ~500 m wide and up to 50 m deep with almost symmetrical flanks that display slope values of respectively 8° and 7° for the north-east and south-east flanks. Its geometry is very similar to the geometry of incision A shown in Fig. 5.4. Incision D is filled with sediments that show a parallel, stratification that lies onlaps onto the incision flanks. These reflectors and that displays two seismic facies: almost transparent at the bottom, changing into a highly reflective facies in the upper part of the infill. The sediments are truncated by an erosional surface sub-parallel to the basal surface and to the infill reflectors, also onlapping on the incision flanks. This internal surface marks the observed change in the facies of the infill.

The second incision, named C, is present on the south-western part of the line and is carved into the Lower Greensand and Weald formations, measuring ~2 km in width and 40 m in depth. The infill displays smaller-scale cup-shaped internal erosional surface, present in the north-eastern portion of the incision. To the north-east, the incision becomes wider and deeper, as shown by Fig. 5.6, located just 300 m north-east of Fig. 5.5. In this area the incision is 4 km wide and up to 80 m deep and displays a complex basal geometry with steeper NE upstream flank (15°) and gentler downstream flank, showing slope values of 4° decreasing to 1° as for incision B described in Fig. 5.4. No lithological boundaries or structural elements controlling the slope pattern are present in this area, apart from small-scale folds deforming the Wealden layers.

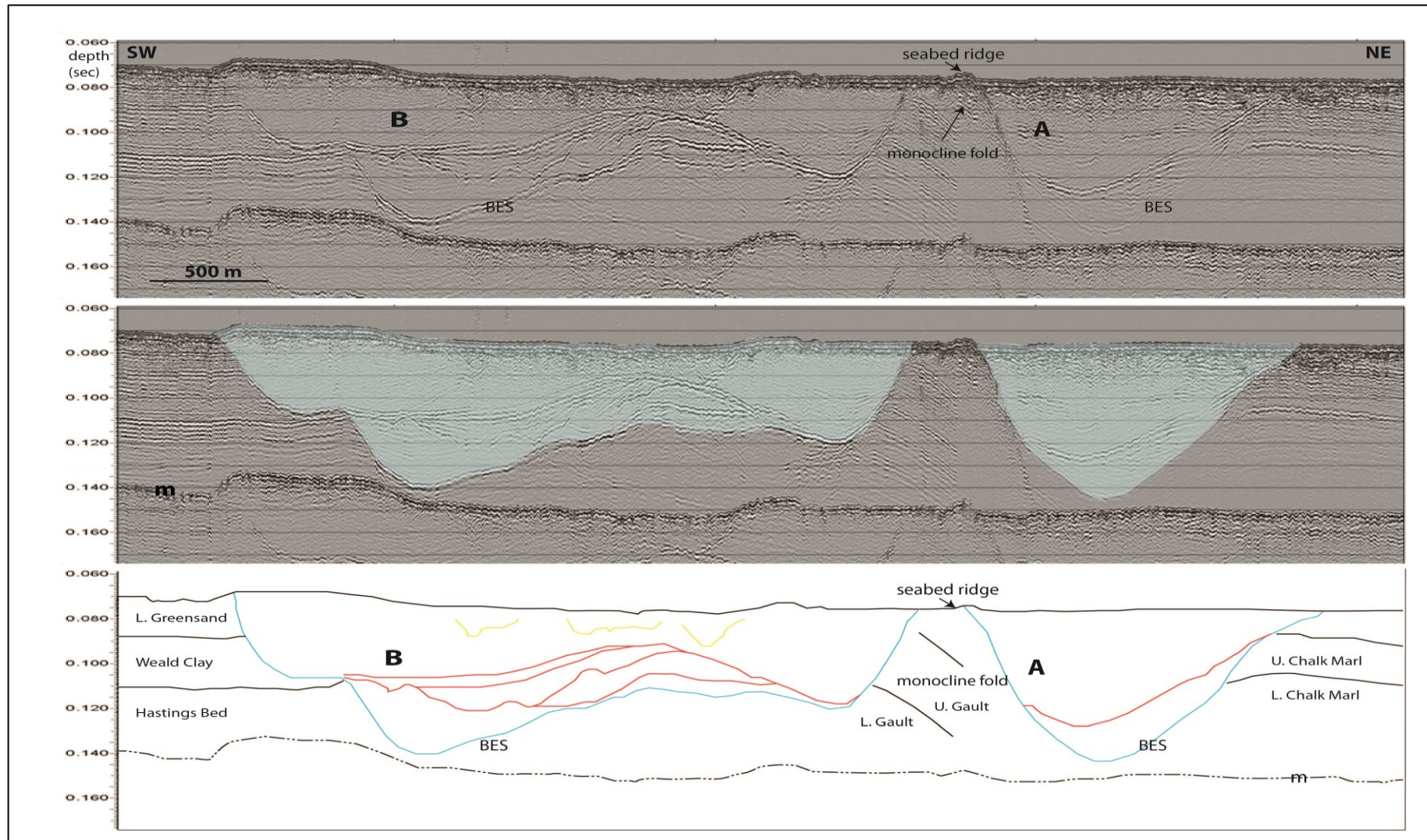


Figure 5.4: Figure showing incisions A and B from seismic line (upper and middle boxes) and line drawing (lower box) displaying the basal surface (BES), internal erosional surfaces (red lines) and **small-scale channels (yellow lines) present in the infill of incision B.** Solid black lines represent bedrock reflectors while the seabed multiple is indicated by the dashed line and labelled as m. VE is 13. See Fig. 5.19 for location of the seismic line.

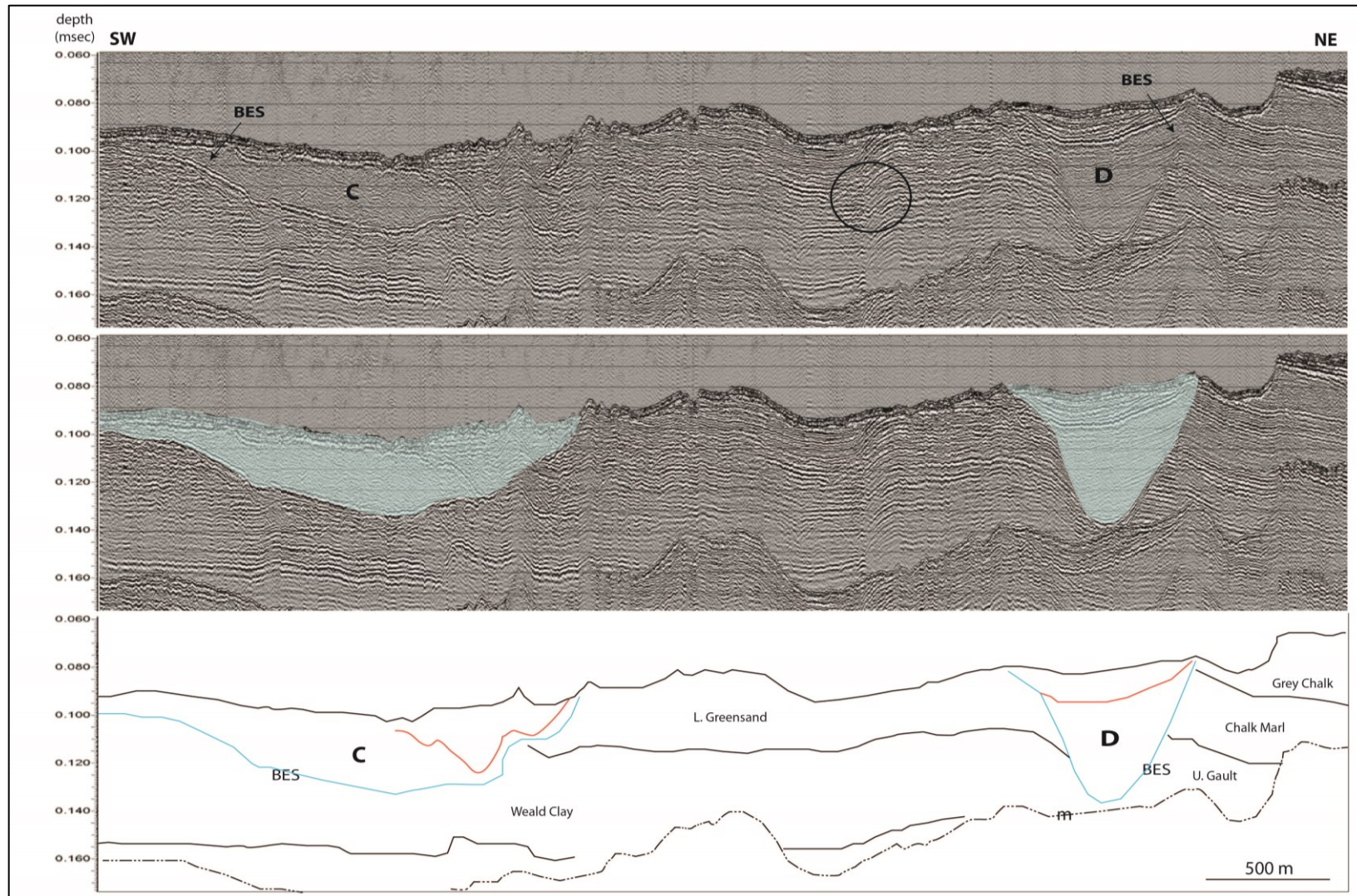


Figure 5.5: Figure showing incisions C and D from seismic line (upper and middle boxes) and line drawing (lower box) displaying the basal surface (BES) and internal erosional surfaces (red lines). Solid black lines represent the bedrock reflectors while the seabed multiple is indicated by the dashed line and labelled as m. VE is 13. See Fig. 5.19 for location of the seismic line.

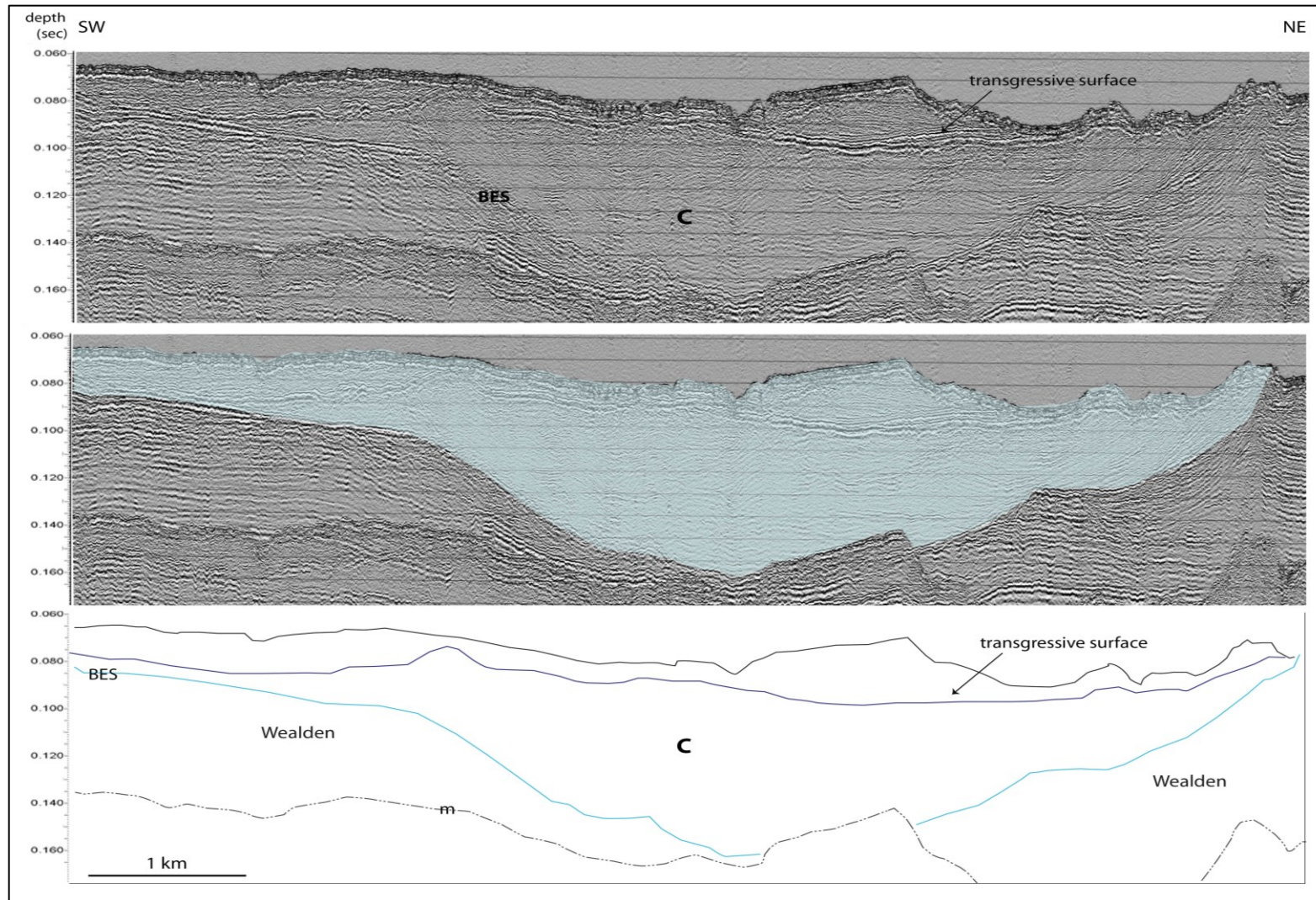


Figure 5.6: This figure shows incision C from seismic line (upper and middle boxes) and line drawing (lower box) displaying the basal surface (BES) and the transgressive erosional surface cutting the infill (blue). Solid black lines represent the bedrock reflectors while the seabed multiple is indicated by the dashed line and labelled as m. VE is 16. See Fig. 5.19 for location of the seismic line.

5.4.3 Plan view geometry of the incisions

Observations

The isopach map of the incisions was constructed using the thickness of the incision infill mapped from the seismic reflection lines. Because of the very close spacing of the parallel track-lines of the seismic dataset, it is possible to represent plan-view geometry of the incisions. The map displays four main isolated incisions carved into the bedrock, showing an infill thickness that varies from 30 up to more than 80 m (Fig. 5.7). All the incisions are completely filled by sediments. Because the seabed is sediment free, the infill isopach map also corresponds to the structure map of the bedrock surface, indicating the depth of the incisions respect to the seabed. The incisions are all located in an area of about 15 km². Incision A, located on the north side of the monocline fold, incisions B and C, located on the southern flank of the fold and the smaller incision D located south-east of incision C. All the mapped incisions are located south-west of the Middle-Lower Chalk Escarpment, as shown by Fig. 5.7.

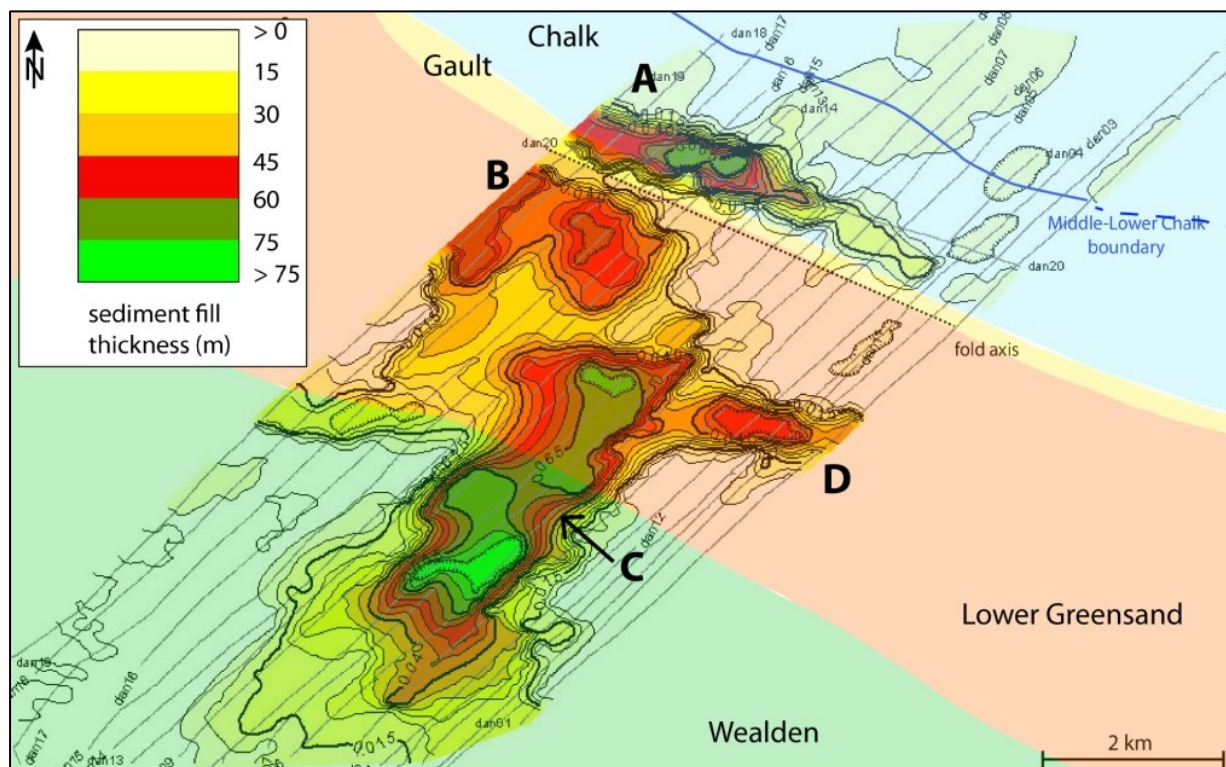


Figure 5.7: Isopach map showing the thickness of the incision infill and the depth of the basal erosional surface of the incisions below the seabed. The four main mapped incisions are labelled in the map. The axis of the monoclinial fold (dotted line) and the Middle-Lower Chalk escarpment (blue line) are also displayed in the figure. (See Section 3.2 for details on the TWT-depth conversion.)

Incision A

Incision A shows a maximum relief of ~80 m, although the deepest portion of the incision is partially masked by the presence of the seabed multiple on the seismic lines. In the Channel Tunnel seismic reflection data (see Chapter 3 for details), the incision reaches up to 100 m of depth below the seabed although the resolution of the data is much lower. The plan-view geometry of the incision displays an elliptical shape, elongated in a WNW-ESE direction, being 3900 m long and 900 m wide. The length/width and length/depth ratios are respectively 4 and 48 while width/depth ratio is 11. The incision is carved into the Chalk and Gault Clay formations: towards the south-east, this feature becomes and less wide and deep (< 10 m on the south-easternmost line of the dataset) in the area where the dip of the bedrock strata steeply increases due to the proximity to the fold, whose axis orientation is shown in Fig. 5.7.

Incision B

On the south-western flank of the monoclinal fold, another kilometre-scale incision (B) is present. This feature is less deep (50 m of depth) but larger than incision A, displaying almost circular shape (as indicated by the 1.3 length/width ratio), with a diameter of 1110 m, length/depth ratio of 21 and width/depth ratio of 18. The incision is carved into the Lower Greensand and Weald formations that outcrop on the south-west of the monoclinal fold.

Incision C

Incision C is the largest of the mapped features, measuring 4400 m in length in a NE-SW direction and 1400 m in width with a length/width ratio of 3, width/depth ratio of 17 and length/depth ratio of 55. The incision shows relief of up to 85 m below the seabed but, because of the seabed multiple hiding the deepest reflectors and the Channel Tunnel analogue seismic reflection data not being available in this area, it is not possible to precisely constrain its relief in its deepest part.

Incision D

Located to the south-east of incision C is incision D; this feature shows similar geometry to incision A as it is characterised by an ovoidal, WNW-ESE elongated shape. The incision has length/width ratio of 2.5 and width/depth ratio of 10 while its length/depth value is 20. It

is 1000 m long and 500 m wide with a depth that reaches a maximum value of 50 m below the seabed. The incision is carved into the Lower Greensand and Weald Clay formations, located just 2500 m south-west of incision (Fig. 5.7).

Table 5.2 summarises the geometrical parameters of the mapped incisions, showing that two of the four incisions (A and C) are more than 80 m deep and 4 km long.

Incision	Length (m)	Width (m)	Aspect ratio			Max depth (m)	Flank slope (deg)	Incised formations
			l/w	l/d	w/d			
A	3900	900	4	48	11	> 80	18/15	Chalk, Gault
B	1300	1100	1.3	21	18	60	15/4	LG, W
C	4400	1400	3	55	17	>80	15/4	LG, W
D	1000	500	2	20	10	50	8/7	LG, W

Table 5.2: Morphometric parameters measured for the four interpreted bedrock incisions. For all the features, length is considered the major axis of the incision (direction of elongation), width the minor axis. Slope is expressed in degrees; all the other parameters are expressed in meters. LG is Lower Greensand Formation, W is Wealden.

Interpretation

The morphometric analyses performed on the incision flank slopes indicate the presence of two main recognised patterns: smaller steeper features with almost symmetrical flanks (incisions A and D) and larger and wider features with strongly asymmetrical flanks characterised by variable slope values (incisions B and C).

In the case of incision A, which is bounded on the SW by the monocline fold, the incision displays an almost symmetrical geometry with steep flanks on both sides, the downstream flank being slightly steeper. The asymmetry can be explained by the structural and lithological control, due to the presence of the monocline fold, on the shape of the incision whose orientation follows the axis of the fold. Secondly, the incision is located at the lithological boundary between the Lower Chalk and the softer Gault Clay formations, forming a lithological boundary that could have acted as a preferential layer during the erosion of the incisions. The same geometry was also recognised in incision D which is nevertheless not laterally bounded by any tectonic structures and which is carved into different lithologies respect to incision A.

The second pattern is represented by incisions B and C: these features are characterised by a steeper NE flank, which is interpreted to be the upstream flank, and a downstream SE flank that is composed of two slopes: a steeper inner slope and an outer and gentler slope. As mentioned in the previous section, this change in slope value would only be controlled by the character of the erosional power of the flows that carved the incisions as no lithological boundaries or structural elements control the slope of the incisions in this area.

With regard to the incision spatial distribution, the isopach map of the incision infill shows a sub-circular to ellipsoidal plan geometry of the incisions that display elongation axes orthogonal to each other as for the case of the incisions A and C, indicating no preferential elongation direction. The incisions are scattered and patchy isolated bedrock features that display no connection to upstream or downstream channels and that are not connected to each other. In addition, the incisions do not appear to follow any structural patterns, apart from the presence of the monocline fold that only bounds two of the four mapped features. In the case of incision A, the decrease in depth of the incision on its eastern portion could be related to the increase in the dip (from $\sim 1^\circ$ up to 11°) of the bedrock strata that, in this area of the dataset, are strongly deformed in proximity of the monoclinal fold.

5.5 Infill acoustic facies and seismic stratigraphy

The sedimentary infill of the incisions is characterised by distinct seismic facies and by the presence of internal erosion surfaces whose geometry varies throughout the dataset. This section describes the acoustic facies and internal erosional surfaces interpreted in the fill of each incision: Table 5.3 summarises the interpreted acoustic facies and internal erosional surfaces present in the incision infill, whose acoustic character, geometry and significance is discussed in the following sections.

5.5.1 Observations

Incision A

The infill of incision A (Fig. 5.8) displays two distinct seismic facies, separated by an internal erosional surface. The lower part of the incision is characterised by an almost transparent seismic facies (A1) which shows weak internal reflections.

These are interrupted by a U-shaped, high-amplitude reflector (s1) that is eroded into the underlying sediments and displays a similar geometry and acoustic character to BES, representing an internal erosional surface.

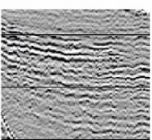
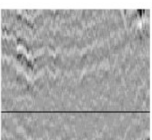
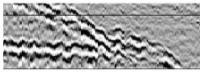
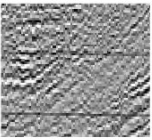
incision A	A1		transparent to semi-transparent facies
	A2		thin parallel reflectors
incision B	B1		transparent to semi-transparent facies
	B2		semi-transparent facies with high amplitude, u-shaped geometry reflectors
			undulated, complex geometry forming two saucer-shaped scours
	B3		semi-transparent facies with discontinuous, irregular geometry reflectors
incision C	C1		thin and dipping parallel reflectors
	C2		no internal erosional surface chaotic and discontinuous thin reflectors
incision D	D1		transparent to semi-transparent facies with parallel dipping reflectors
	D2		sub-parallel, high amplitude reflectors similar to s3

Table 5.3: Summary of the interpreted acoustic facies and erosional surfaces present in the incision infill.

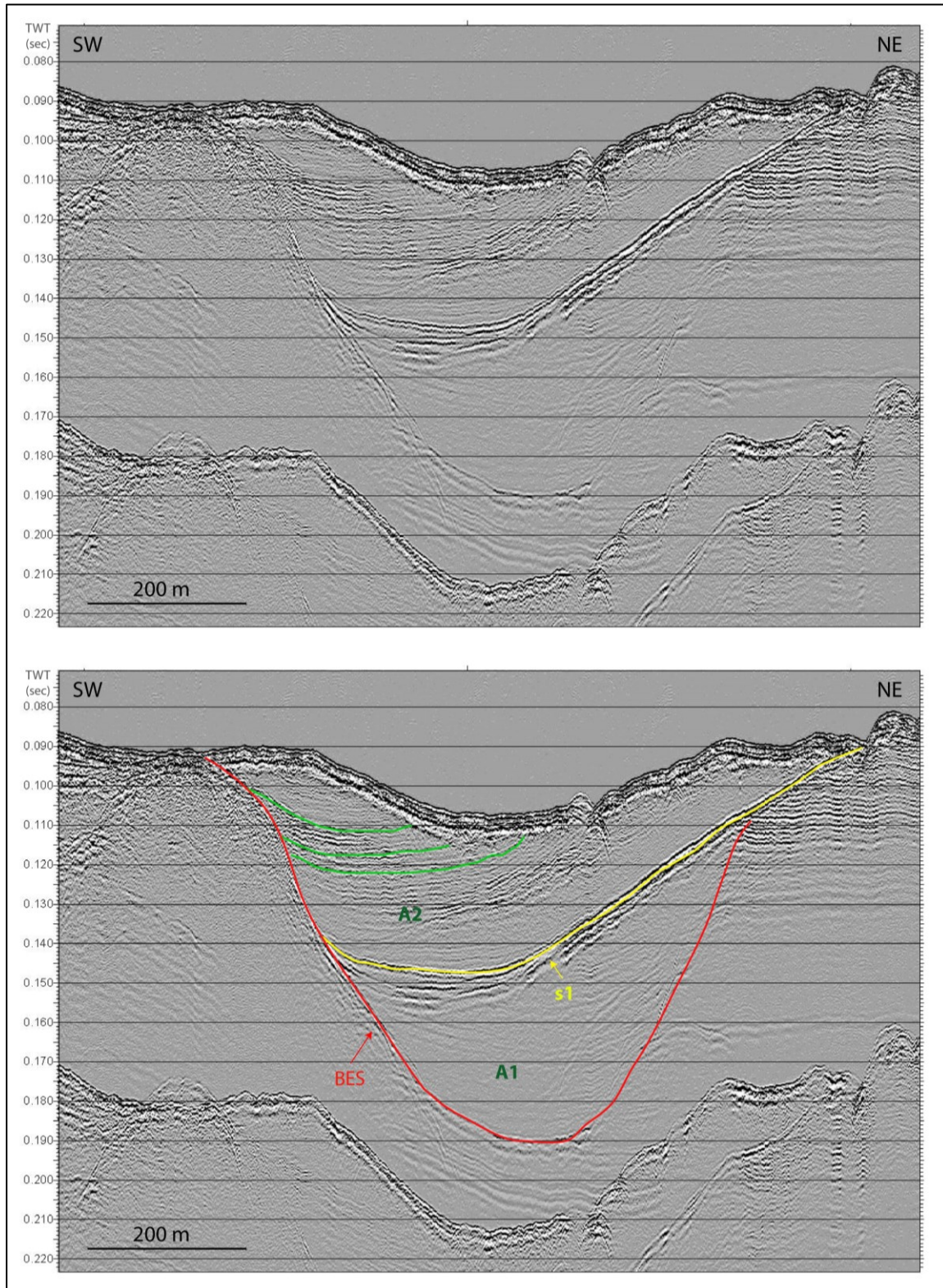


Figure 5.8: Uninterpreted (top) and interpreted (bottom) seismic image of two distinct acoustic facies (A1 and A2) present in the infill of incision A, separated by the erosional surface s1. The black circles indicate the onlap termination of s1 and reflectors of the upper facies on BES. VE is 5. See Fig. 5.19 for location of the seismic line.

Surface s1 displays a steep north-east and a gentler south-west flank and its depth varies laterally ranging from 40 to 55 m below the seabed. This internal erosion surface is overlain by a package of sediments whose facies (A2) is characterised by well stratified, thin horizons with regular geometry. These reflectors lie almost parallel to the underlying surface s1; the reflectors visible in the upper part of the infill (A2) display an onlap termination onto the south-west flank of the incision (Fig. 5.8). Fig. 5.9 shows the variation in the internal architecture of the incision infill. Facies A1 maintains the same transparent acoustic character throughout the dataset and its top is bounded by the erosional surface s1 that divides A1 and A2. The geometry of s1 changes substantially throughout the three shown seismic lines. The surface displays a more asymmetrical and shallow geometry in the central and eastern part of the dataset, also determining a change in the thickness of the upper facies A2 that varies from 0.015 to 0.050 sec as also indicated by the isochrone map. The acoustic character of A2 also varies, changing from parallel reflections that drape s1 and display onlap geometry onto the surface, to a more chaotic facies in the western area (line 3 in Fig. 5.9).

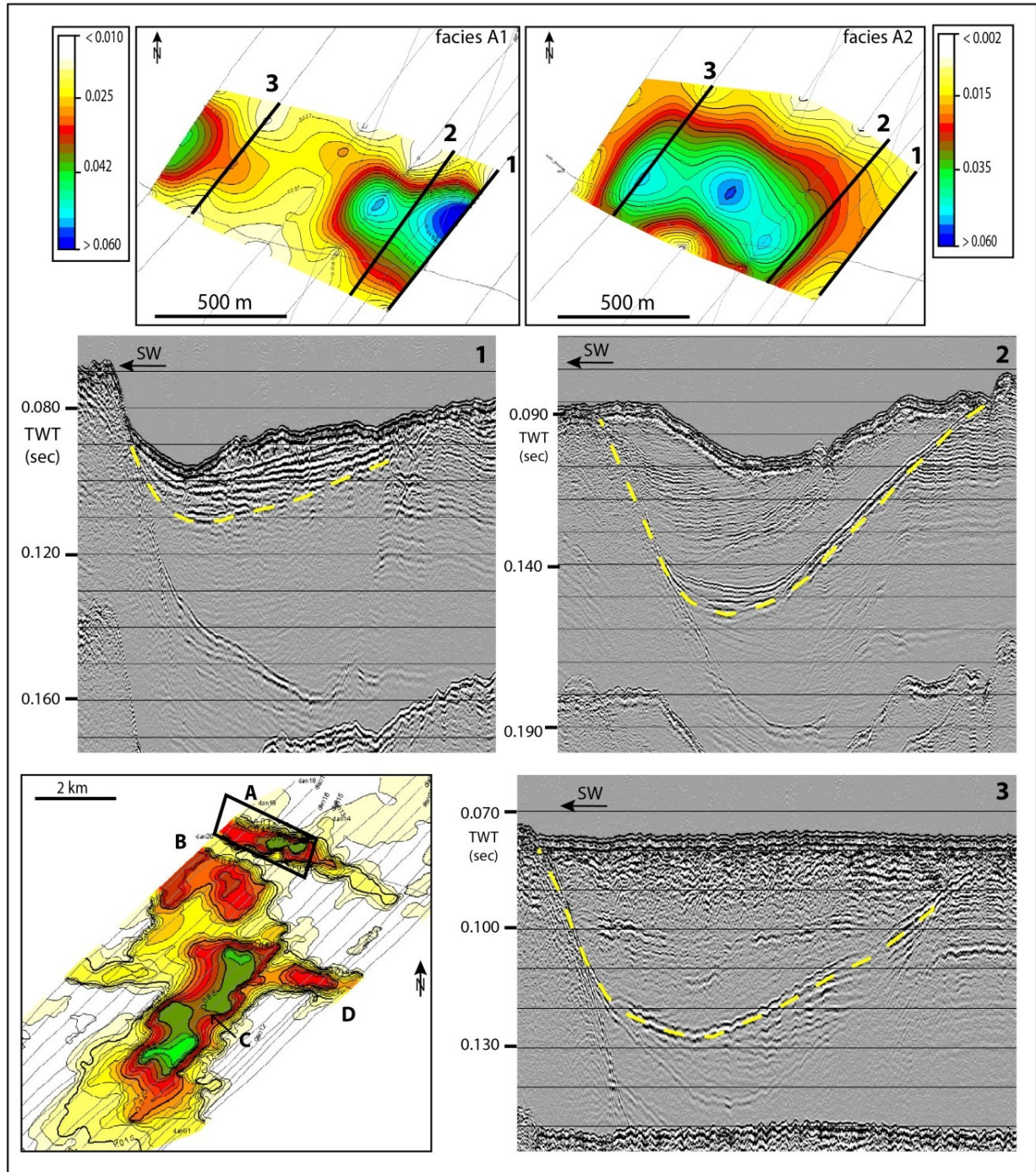


Figure 5.9: Three extracted seismic images showing the variation in the geometry and thickness of the infill of incision A. The dashed yellow line represents the internal surface s1 that separates the lower (A1) and upper (A2) facies of the infill whose thickness is represented in sec the isochrone maps.

Incision B

The seismic facies of incision B infill (Fig. 5.10) is characterised by irregular, discontinuous, low frequency reflectors. The incision infill is subdivided into 3 different scours (scours 1, 2 and 3 to which facies B1, B2 and B3 are associated) by erosional surface s2. This high amplitude reflector is characterised by undulating geometry: while the short wavelength undulating effect is due to ship heave artifacts, the longer wave-length irregular geometry of the BES is due to the amalgamation of erosional surfaces caused by multiple incisions (Fig. 5.10a). BES cuts the incision infill, sharply truncating the reflectors of scour 1, located at the base of the incision. s2 also locally reaches the base of the incision, merging with BES. The geometry of this internal surface is complex that marks two U-shaped upper scours (2 and 3) carved into the underlying sediments of scour 1. Scour 1 is 700 up to m wide and up to 30 m deep, showing asymmetric geometry with a steeper south-western flank represented by BES which is cut at the top by s2. The infill of the scour shows transparent acoustic facies (B1) with presence of weak reflections parallel to BES. In the upper part of scour 1, irregular and discontinuous high amplitude reflections are observed.

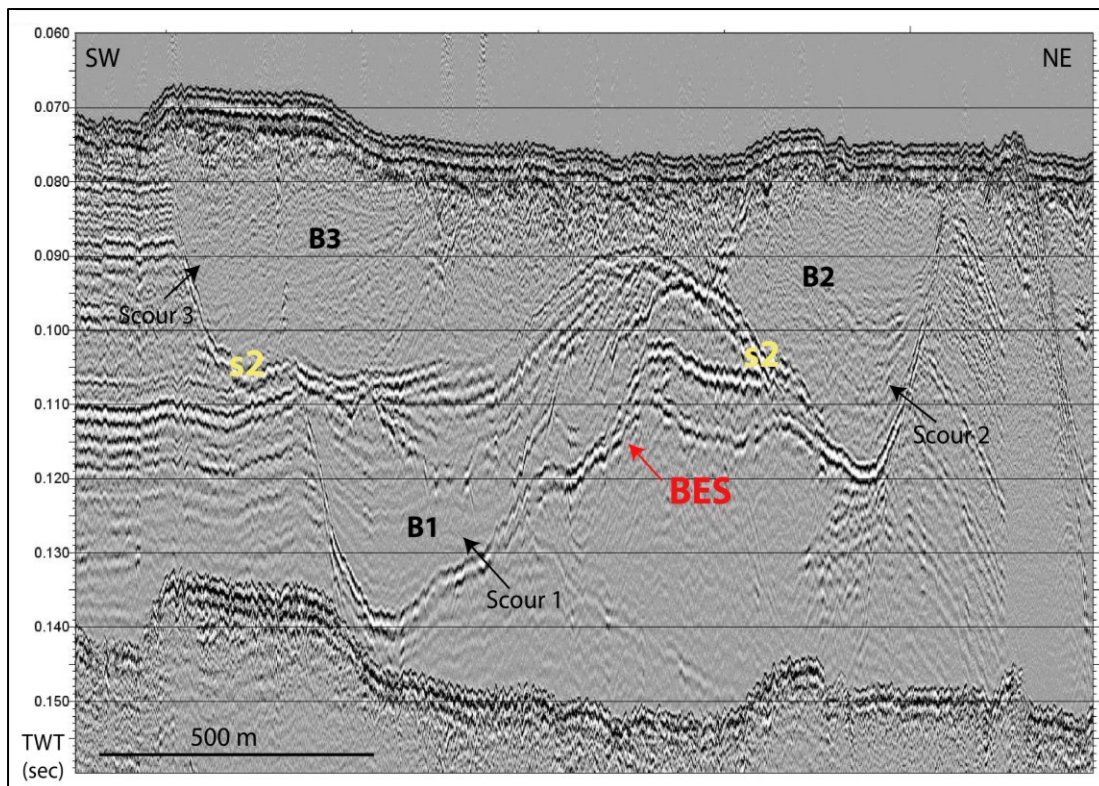


Figure 5.10: Seismic section showing incision B, composed of three scours, separated by the internal erosional surface s2. Each scour displays different seismic facies and a cup-shaped geometry very similar to the basal erosional surface of all the incisions, although on a smaller scale. VE is 13. See Fig. 5.19 for location of the seismic line.

Scour 2 is located in the north-eastern part of the incision and measures up to 400 m in width and 35 m in depth. Its geometry is almost symmetric with a slightly steeper north-eastern margin, bounded by the southwest flank of the monocline. The acoustic facies of the infill (B2) is characterised by a lower portion where parallel reflections are visible, showing onlap geometry on the basal erosional surface bounding the north-eastern flank of the scour (Fig. 5.10). The upper part shows semi-transparent facies with thin reflectors visible just below the seabed. On the southwest side of incision B, scour 3 is present, showing a seismic facies (B3) that has a very similar character to the upper part of scour 2.

Fig. 5.11 shows the variation in the architecture of the infill and erosional surfaces of incision B, showing how the geometry of the basal surface varies across the dataset. Fig. 5.11. also displays how the geometry of s2 changes significantly, determining a change in the geometry of the scours and their infill. In particular, s2 is deeper in the eastern and central portion of the incision, carving the sediments of facies B1 that displays a very limited thickness in this area as s2 overlaps with BES (lines 1 and 2 in Fig. 5.11). Moving westward, s2 is shallower but still characterised by saucer-shaped geometry, forming two scours into the underlying sediments. s2 is characterised by more than one high amplitude reflector that display cross cutting relationships (line 3 of Fig. 5.11). In summary, the depth and width of the scours varies locally, depending on the geometry of s2 that separates the acoustically transparent strata of scour (B1) from the semi-transparent strata present in scours 2 and 3 (B2 and B3).

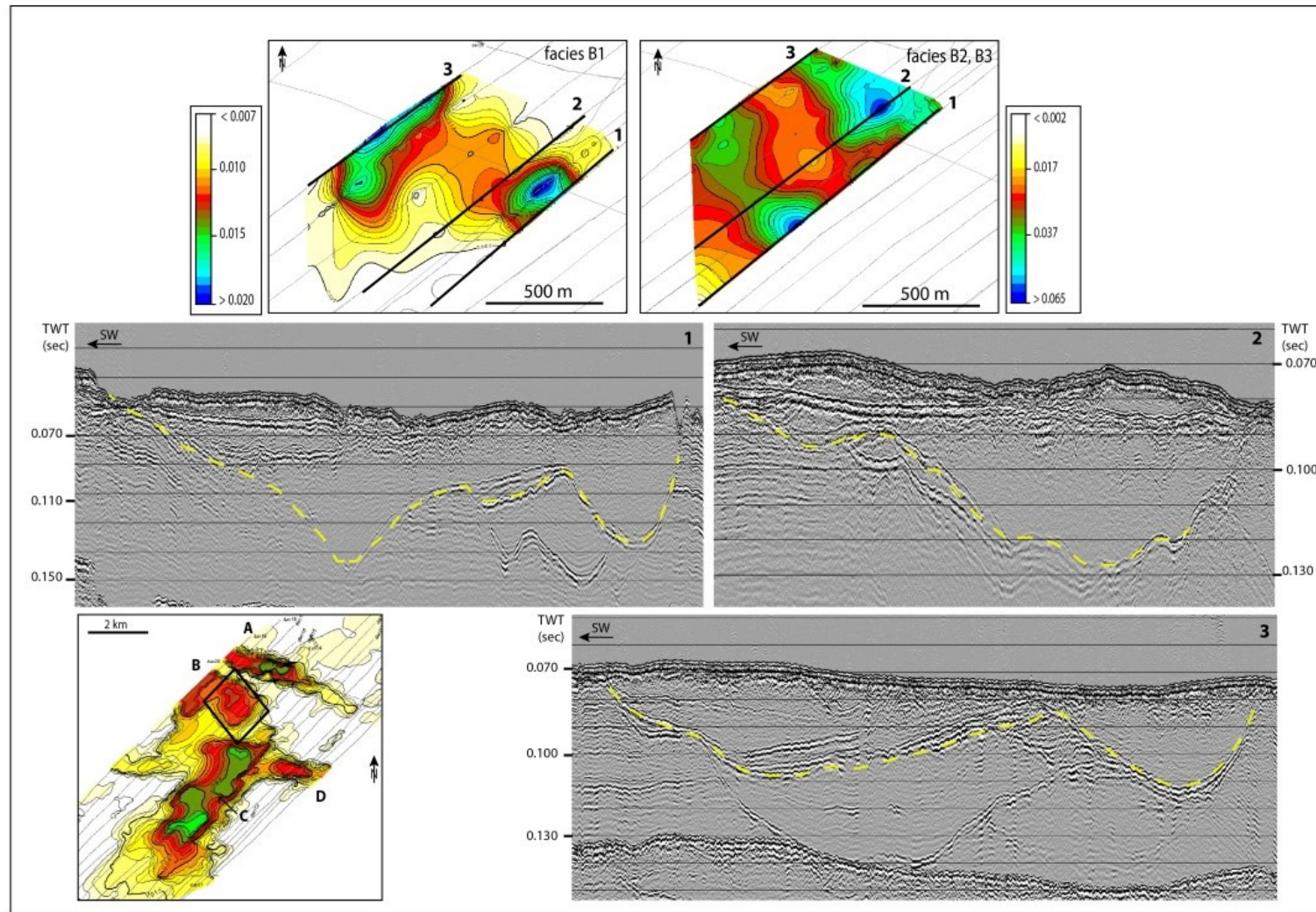


Figure 5.11: Three seismic reflection images showing the variation in the geometry and thickness of the infill of incision B, whose location is illustrated in the in the infill facies isochrones maps. The image in the lower-left corner is the incision map shown in Fig. 5.7, illustrating in which portion of the dataset the isochrones maps are located. The dashed yellow line represents the internal surface s2 that separates the lower (B1) and upper (B2 and B3) facies of the infill whose thickness is represented in TWT (sec) the isochrone maps.

Incision C

The infill of this incision is composed of two main seismic facies. As this feature is the widest of the mapped incision, measuring more than 4 km of width, the seismic character of the infill varies both vertically and laterally. The basal part of the infill shows, on the NE flank of the incision, thin and high-frequency sub-parallel reflectors dipping SW (facies C1), as shown in Fig. 5.12. Moving to the south-west, the termination of these reflectors is not visible as the reflections become weaker and much more discontinuous. Here, the infill displays a semi-transparent facies with some discontinuous, high-amplitude and low-frequency reflectors still visible. Moving further south-west the infill gradually becomes more transparent (infill facies C2) and no internal reflections are visible.

In contrast to the other mapped incisions, the infill of incision C does not display the presence of an internal high-amplitude surface. Nevertheless, moving upward in the infill stratigraphy, the incision is characterised by a change in the acoustic character of the sediments, revealing higher reflectivity and, locally, continuous and sub-parallel reflections.

Fig. 5.12 shows the change in the infill acoustic character in along strike-direction. Despite a change in the incision depth moving south-east, the infill character does not change significantly, showing two distinct facies (C1 and C2) on all the lines. The only exception is represented by the presence of a U-shaped surface that carves the sediments of the incision in its eastern portion, as shown by line 1 of the figure 5.12.

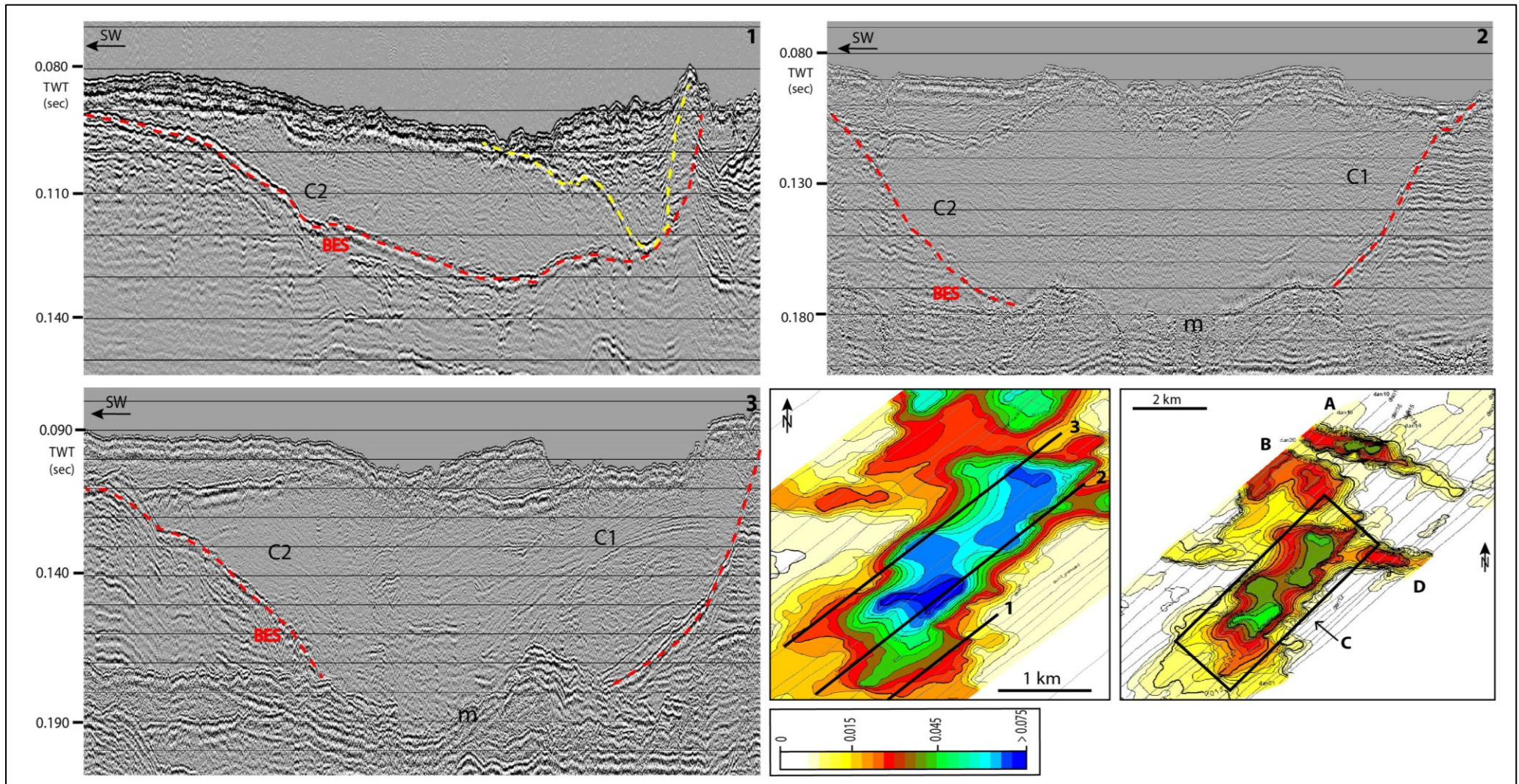


Figure 5.12: Three extracted seismic images showing the variation in the geometry of the basal erosional surface and of the infill of incision C. The two maps show the location of the incision (right box) and the thickness of the incision infill in TWT (sec) (isochrone map on the left). Dashed yellow line indicates the U-shaped erosional surface, m indicates seabed multiple.

Incision D

The infill of incision D (Fig. 5.13) displays two distinct facies, separated by an erosional surface (s3): the deeper portion comprises facies D1. The infill displays parallel reflectors with onlap termination on both flanks of the incision. The infill stratification is not parallel to the basal surface but lies at an angle of 45° on BES, gradually becoming more horizontal going upward in the infill. This is truncated by a high amplitude U-shaped reflector (s3) carved into the sediments and displaying symmetrical geometry and high energy acoustic character. The seismic facies of the sediments present above the high amplitude surface is named D2 and is characterised by high amplitude, well stratified reflections parallel to s3.

Fig. 5.14 shows how, moving from NW to SE across the seismic dataset, the depth of s3 ranges from 15 to 35 m below the seabed as the geometry of the incision also changes. Where the incision is deeper and BES displays a more symmetrical geometry, s3 also shows symmetrical geometry, almost parallel to BES. Here, the surface has lower relief and a less concave geometry respect to the central and eastern portion of the incision where s3 is only present on the south-eastern flank of the incision.

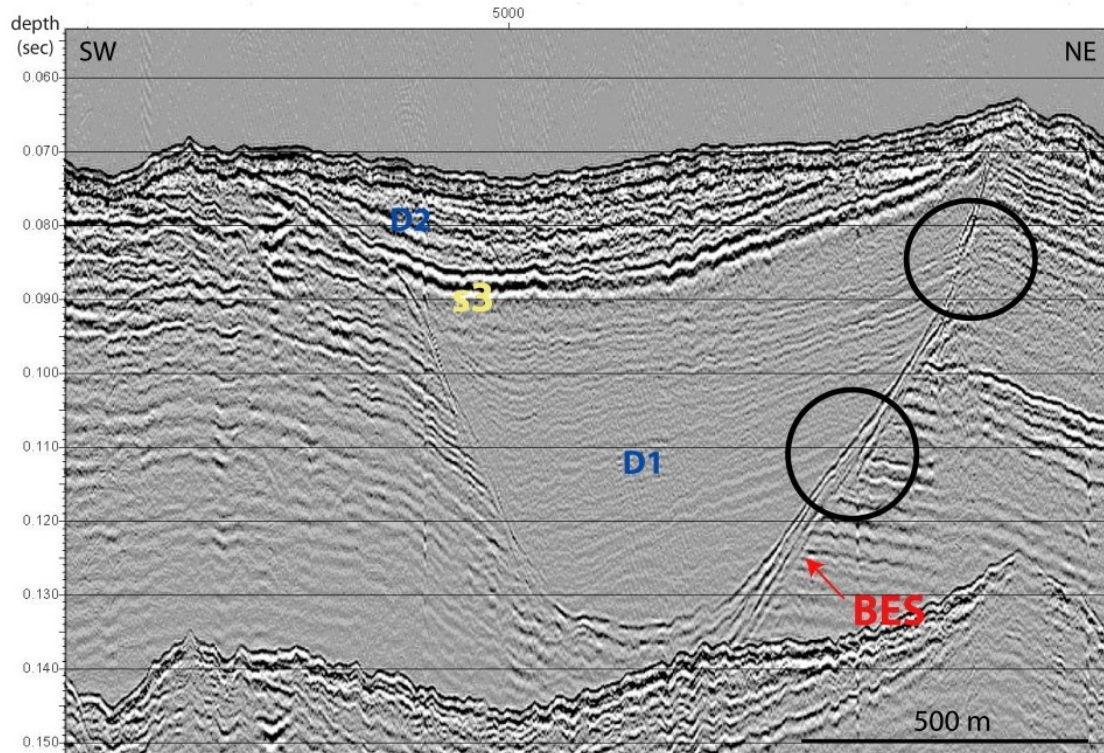


Figure 5.13: Image of incision D showing onlap termination (indicated by the black circles) of the inclined infill reflectors in seismic facies D1. Note the abrupt change from facies D1 to D2 in correspondence of the internal erosional surface s3. VE is 13. See Fig. 5.19 for location of the seismic line.

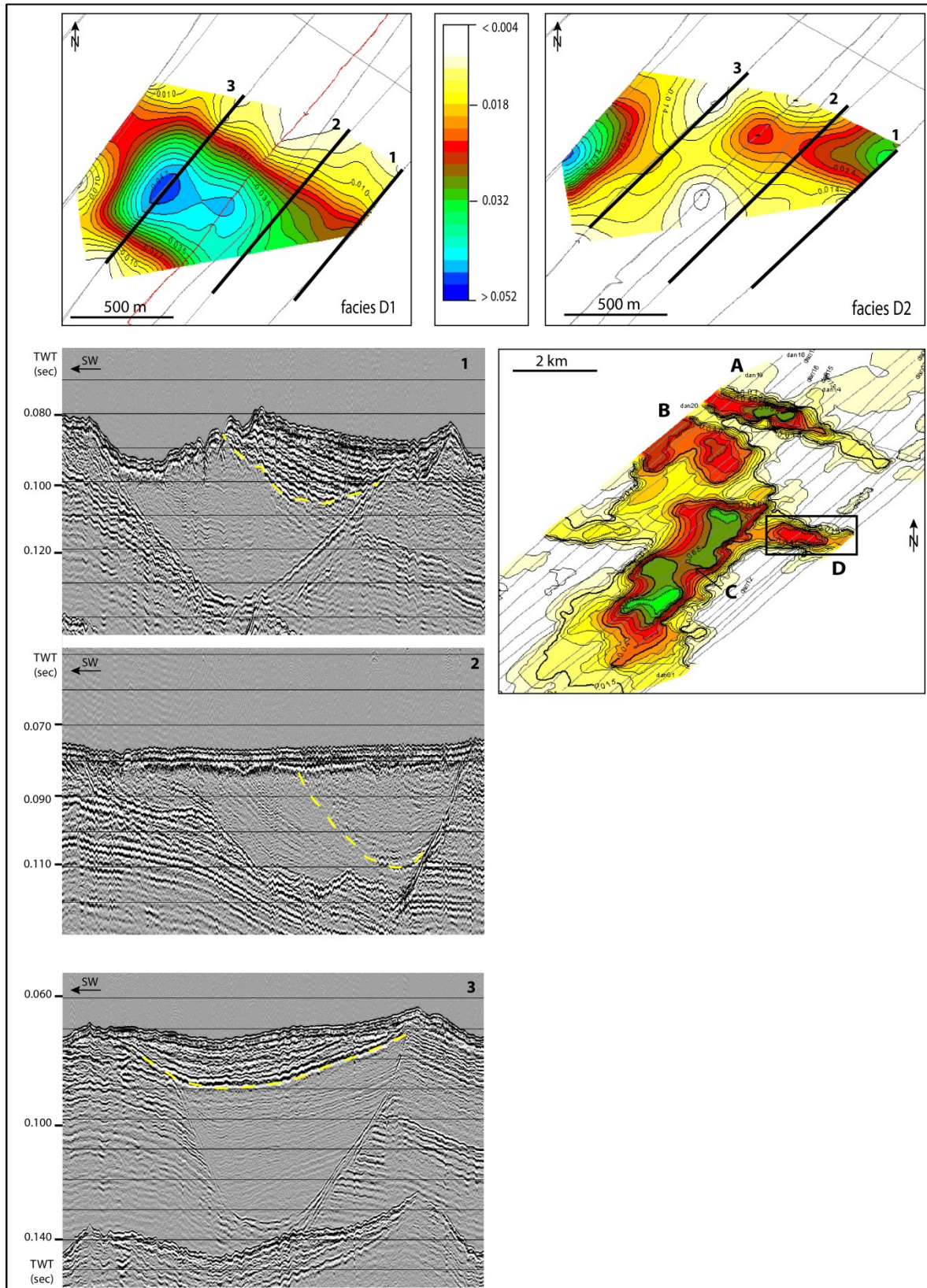


Figure 5.14: Three extracted seismic images of incision D showing the variation in the geometry and in the thickness of the infill. Two distinct seismic units compose the infill: D1 at the bottom and D2 at the top, separated by the internal erosional surface s3. The thickness of the two units is displayed in TWT (sec) in the isochrone maps in the top left (D1) and right (D2) corners.

Both incisions B and C are characterised by the presence of small-scale erosional channels carved into the infill (Figs. 5.10-5.12): these are 10 m deep and 50 to 300 m wide and are often accompanied by diffraction hyperbolae affecting the reflections present below the seabed. The dimension of these channels stresses the difference in scale between these features and the hundred of metres deep incisions that compose the Fosse Dangeard.

5.5.2 Interpretation

The analysis of the acoustic facies (Table 5.2) of the sediments that fill the incisions suggests the material composing the bottom part of the infill to be represented by unstratified material, as indicated by the transparent acoustic character of part of the infill. The presence of internal erosional surfaces (s1, s2 and s3) determines a change in the infill facies that generally becomes more reflective, showing more irregular reflections composing higher amplitude and less-transparent acoustic facies in the upper part of the incisions.

Where the internal stratification is visible and better imaged (incisions A and D) the infill stratification seems to have variable geometry, ranging from parallel to high angle stratification. The basal portion of the infill of both incisions A and D display high-angle stratification with SW dip of the infill reflectors (apparent dip up to 5°): this might be explained by the narrow geometry of the incisions, characterized by a low width/depth ratio, indicating a narrow and deep shape of the incisions. As features A and D are bounded by steep and narrow flanks, the sediments inferred as coming from the NE, might have been deposited with a high-angle geometry and downstream dip at the narrow base of the incisions. Conversely, the upper part of the incision infill displays an almost parallel geometry to the internal erosional surfaces with sediments that drape surfaces s1, s2 and s3.

The internal high amplitude surfaces (Table 5.2) present in the infill display the same high amplitude character as for the basal surface of the incisions, represented by BES, abruptly truncating the bedrock layers. A sharp erosional contact is visible between the infill sediments and the internal surfaces, whose dimensions and depth of erosion into the infill (up to more than 50 m deep) suggest high-energy character of the erosional events that carved the surfaces. This is also suggested by lack of gradual transition from deposition of the infill sediments to erosion of the internal surfaces. The similarity in the acoustic character and geometry of the internal surfaces and basal erosional surface suggests that s1, s2 and s3 were

perhaps carved into the incision infill by later erosional processes similar to the events that carved the basal surface. The presence of one (incision A and D) or more (incision B) internal surfaces suggests that the incisions were carved by at least two erosional events.

5.6 Spatial distribution of the Fosse Dangeard

The isopach map obtained from the seismic interpretation and showing the incision depth and the thickness of their infill was compared to the Fosse Dangeard contour map, previously collated by Destombes et al. (1975), in order to assess the correspondence of the two interpretations as the two datasets partially overlap. Although the location of the seismic lines used by Destombes is not known, the south-western area of Destombes et al. (1975) map, located in the French sector of the Dover Strait, overlaps with the area where the new dataset is located.

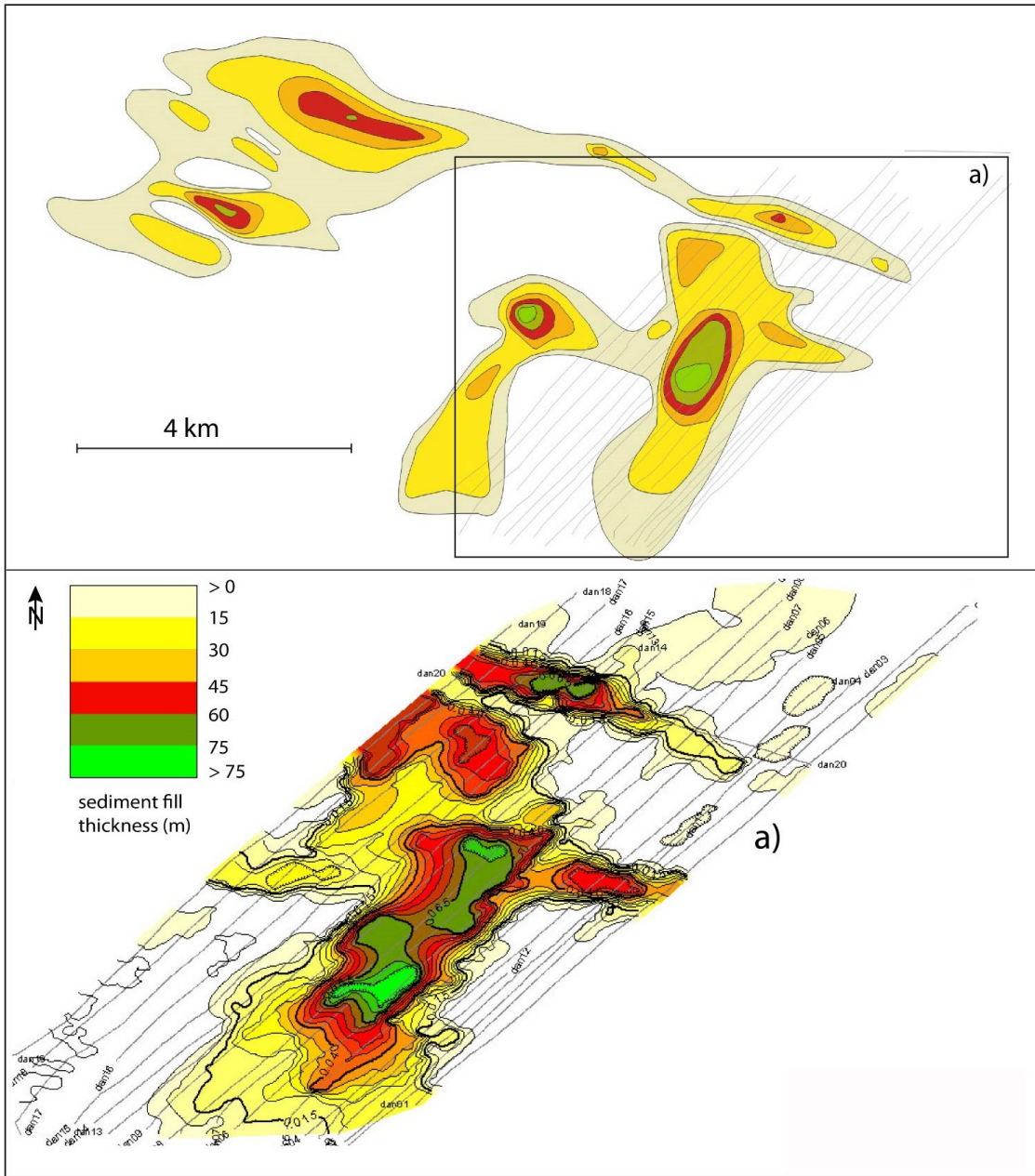


Figure 5.15: Comparison between Destombes map (upper contours) (Destombes et al., 1975) and the new interpreted contour map (lower contours) showing very good match between the interpreted features that display the same location and geometry in both maps. The new isopach map also shows significantly improved resolution of the mapped features. The extent of the area where the two maps overlap is shown by box a. The two maps are displayed using the same colour-scale.

The comparison between Destombes et al. (1975) map and the new isopach map is displayed in Fig. 5.15, using the same colour-scale for both maps. Firstly, the comparison reveals a good match between the location of the features interpreted by Destombes and those interpreted in this work. In addition, thanks to the improved quality of the high-resolution

dataset, the new map shows a strong increase in the resolution in the mapping of the geometry of the incisions.

As the interpreted incisions are in the same position in both maps, it is possible to infer that the incisions mapped by Destombes et al. (1975) in the UK sector where no new data coverage is available, may have been formed by the same processes that carved the group of incisions interpreted in the high-resolution dataset. This is consistent with the identification of bedrock incisions with the same acoustic facies of the features mapped in this work, located on the UK sector of the Dover Strait in the area where Destombes incisions are located (David Garcia-Moreno, pers. comm.). The incisions show an almost symmetrical distribution in both areas (Fig. 5.16), displaying same direction of elongation of the incisions in both sectors and confirming the presence of the incisions not only carved into the Lobourg Channel area but are also present on the bedrock platform that bounds the western margin of the channel (light-grey surface on the west of the sand bank in Fig. 5.16), as displayed by the map in Fig. 5.16 where the location of the incisions in both sectors of the Strait is shown.

The analyses and interpretation of the surfaces into which the incisions are carved, discussed in Chapter 6 of this work, allows for the reconstruction of the relative age of the surfaces and relative timing of the formation of the Fosse during the flood events.

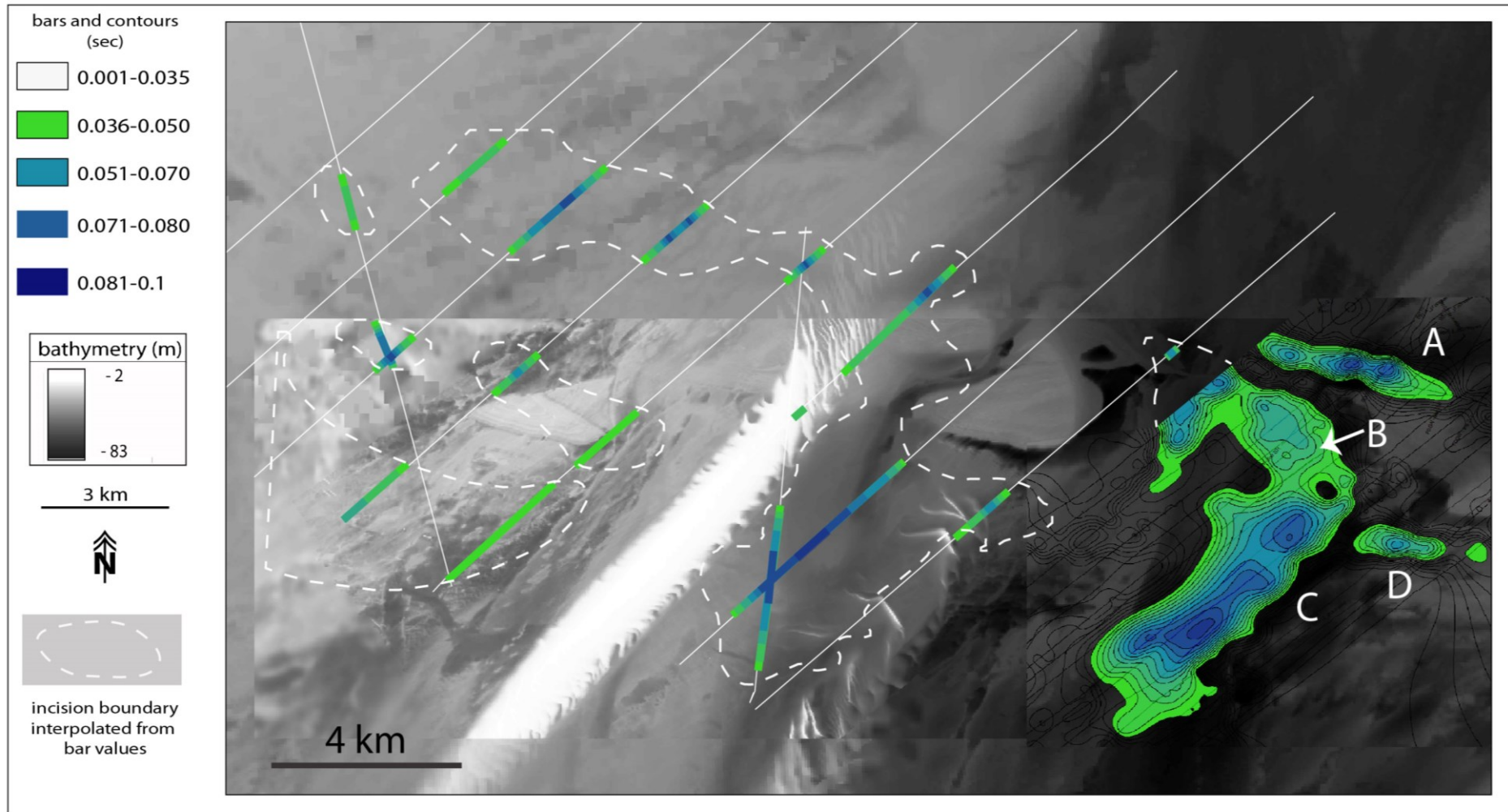


Figure 5.16: Contour map (on the French side) and coloured bars (on the UK side, depth provided by David Garcia Moreno) showing the depth of the basal surface of the incisions carved into bedrock at the Dover Strait, with bathymetry on the background (30 m bin size DTM). The dashed white lines indicate the boundary of the incisions obtained from contouring of depth values shown by the bars. The map shows how the incisions are located on more than one erosional surface, being carved into the bedrock platform (located on the west of the NE-SW elongated sand bank) present on the British side of the Strait, other than on the Lobourg Channel floor where the contours are shown.

5.7 Discussion: formation of the Fosse Dangeard

5.7.1 What is the mechanism responsible for the formation of the Fosse Dangeard?

Different mechanisms were proposed in the past as responsible for the formation of the Fosse Dangeard incisions, ranging from fluvial (e.g. Stamp, 1927; Hamilton and Smith, 1972; Gibbard 1985) to tidal (Hamblin et al., 1992) processes to tectonic control (Van Vliet-Lanoe et al. 2004). The analyses of the incision morphology and described in this chapter indicate that both fluvial and tidal hypotheses are not consistent with the observed dimension of the incisions. The process responsible for the erosion of the kilometre-scale observed incisions requires extraordinary high-energy and erosional power that cannot be produced by fluvial or tidal action (e.g. Rudoy and Baker, 1993; Baker, 2002b). Structural control is not consistent with the formation of the Fosse as the seismic reflection data show that the incisions are bounded at their base by sharp erosional contact rather than tectonic structures. Glacial processes were also previously interpreted as responsible for the formation of the depressions (Destombes et al., 1975; Kellaway et al., 1975) and similarities can be observed between the incisions and glacial tunnel valleys located in the North Sea (e.g. Laban, 1995; Praeg, 2003; Lonergan et al., 2006; Graham et al., 2007). Firstly, the infill is in both cases characterised by acoustically transparent facies (which, in the case of the tunnel valleys corresponds to fine grained material) and by two distinct seismic facies (Cameron et al., 1987; Huuse and Lykke-Andersen, 2000; Kluiving et al., 2003, Lonergan, et al., 2006; Krohn et al., 2009). In addition, both the Fosse Dangeard incisions and tunnel valleys show an irregular, rounded geometry of the basal erosional surface (Huuse and Lykke-Andersen, 2000; Praeg, 2003), which results from erosion by flows with elevated hydraulic pressure (Benn and Evans, 1998). Despite these similarities, the Fosse incisions have remarkably different plan-view geometry respect to tunnel valleys that show a typical arborescent and branched pattern (Ó Cofaigh, 1996; Praeg, 2003), often elongated in the direction of the flow. Conversely the incisions are isolated features that display an almost circular (incision B and C) to ovoidal shape perpendicular to the suggested flow direction (incisions A and D). Furthermore, in the case of the Fosse depressions, the tunnel valley interpretation is not appropriate as the Dover Strait area was never reached by the ice-sheet during Quaternary glaciations (e.g. Clark et al., 2004) and tunnel valleys only form under ice-sheets (e.g. Ó Cofaigh, 1996; Praeg, 2003).

The erosional character of the basal and internal surfaces and the observed morphology and dimension of the incisions require a high-magnitude event for their erosion. This is also suggested by the sharp contact between the saucer-shaped basal erosional surface and the bedrock reflectors, together with the isolated, sub-circular planar geometry of the incisions. The absence of headward erosion could also indicate a rapid, short-lived event to have carved the incisions (Meinsen et al., 2011). This evidence suggests a high-magnitude process to have carved the depressions. Typical fluvial processes do not have the required erosional power to carve such kilometre-scale features into bedrock in a short period of time (e.g. Baker, 2002b; Carling et al., 2009b).

The mapped incisions were compared to features previously interpreted in other megaflood terrains, showing strong similarity to bedrock scours carved into bedrock by high energy flows. These features, identified as plunge pools, are observed in several megaflood terrains on Earth, measuring tens of metre of depth. Plunge pools with similar scale to the mapped incisions are present at the Altai Mountains terrain, showing up to 70 m of depth (Rudoy, 2002) and in the Munsterland Embayment megaflood terrain, measuring 400 m of width and 35 m of depth (Meinsen et al., 2011). Other examples of plunge pools, carved by catastrophic, short-lived events are present at the Box Canyon (Lamb et al., 2008) and Channeled Scablands (Baker, 1978a). The Dover Strait incisions show the same saucer-shaped, sub-circular geometry typical of plunge pools, interpreted as the result of erosion by water-falling jets of water (Howard et al., 1992; Lamb et al., 2007), cascading from hydraulic jumps located at variable heights (e.g. 70 m for Altai Mountains and 35 m for Munsterland plunge pools). 2D models show that these features form, during megafloods, in upstream areas characterised by high discharge and flow velocities (Denlinger and O'Connell, 2010) by abrasion and cavitation processes (Whipple et al., 2000; Baker, 2009a).

In the case of the Fosse Dangeard, the incisions are scattered at the base of the Chalk Escarpment (Fig. 5.7), located on the northern flank of the Weald-Artois anticline. This fold has been interpreted to have impounded the Southern North Sea glacial lake from which the megaflood was originated (Smith, 1985; Gibbard, 1995; Gupta et al., 2007). This suggests that the presence of the escarpment would have acted as a structural dam for the lake and as a hydraulic jump for the cascading water, accelerating the water flow that carved the depressions into the Upper Cretaceous strata of the Dover Strait (Fig. 5.17).

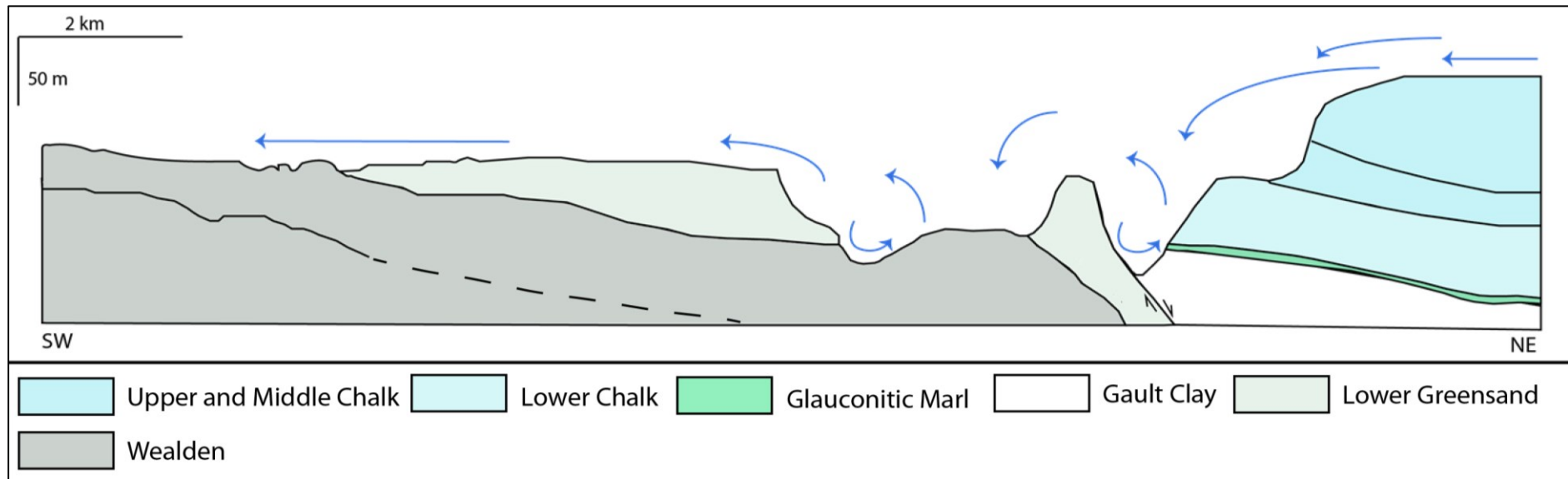


Figure 5.17: Cartoon of the plunge pool formation at the Dover Strait. Note the morphological step at Middle-Lower Chalk Escarpment. The blue arrows indicate the direction of the flow that carved the plunge pools, coming from the NE and flowing towards the SW into the English Channel. The two incisions would be separated by the Lower Greensand ridge (also visible from bathymetry and described in Section 6.2), composed of harder material. The geometry of the incisions would be in this case driven by lithology that followed the softer Gault Clay layers along the monocline.

It is only possible to estimate the elevation drop present at the Chalk Escarpment at the time of the formation of the incisions using the present-day elevation of the escarpment. At present, The Chalk Escarpment is ~100 m elevated above MSL, as indicated by profiles drawn on the available onshore SRTM topographic data (see Section 3.3 for details): as the incisions are 80 m deep into the seabed, an elevation drop of ~180 m can be estimated for the erosional event that carved the depressions, breaching the bedrock dam that impounded the south-western margin of the pro-glacial lake. The ~100 m hydraulic jump represented by the Chalk Escarpment would have accelerated the flood water that, cascading, carved the 80 m deep incisions into bedrock at the base of the palaeo-waterfall.

Comparison to artificial and natural plunge pools

The interpreted incisions were also compared to small-scale natural river (Comiti et al., 2005) and artificial laboratory (Pagliara et al., 2008a, 2008b) plunge pools, in order to analyse the geometry and slope pattern of these features at different scales (Fig. 5.18). The comparison shows a strong similarity, both in the plan view and cross-section geometry, of natural (Fig. 5.18a) and artificial (Fig. 5.18 b) plunge pools. The Dover Strait incisions were compared to metre-scale artificial laboratory plunge pools (Pagliara et al., 2008a, 2008b), both displaying steeper flank on the upstream active portion and gentle downstream passive flank, characterised by two slope values (Fig. 5.18b). Geometrical parameters calculated for the analysis of natural stream step-pool geometry (Comiti et al., 2005) were also applied to the kilometre-scale interpreted incisions, with the aim of recognising possible geometrical patterns in the geometry of the interpreted depressions.

As a result, two main conclusions can be drawn from the analyses on plunge pools; firstly, there is a strong morphological similarity between the plunge pools, independent from their scale and nature, especially in the flank slope pattern. Secondly, despite the limits in resolution and difference in scale, the identification of geometrical patterns represented by characteristics slope values of upstream and downstream flanks of the plunge pools. These results indicate that parameters used for small-scale artificial and natural pools can also be used to characterise and describe the geometry of bigger scale features, in order to identify geometry and slope patterns. In particular, the measured parameters describe characteristics such as elongation, depth/width ratio and flank slope. Finally, as the two observed patterns in the geometry of the Dover Strait incisions cannot be explained by lithological or structural

control (see Section 5.5.3), the experiments conducted by Pagliara et al. (2008a, 2008b) suggest that the change in geometry could be due to the variation of hydraulic factors, such as jet height or angle of impact of the flows that carved the plunge pools during the flood.

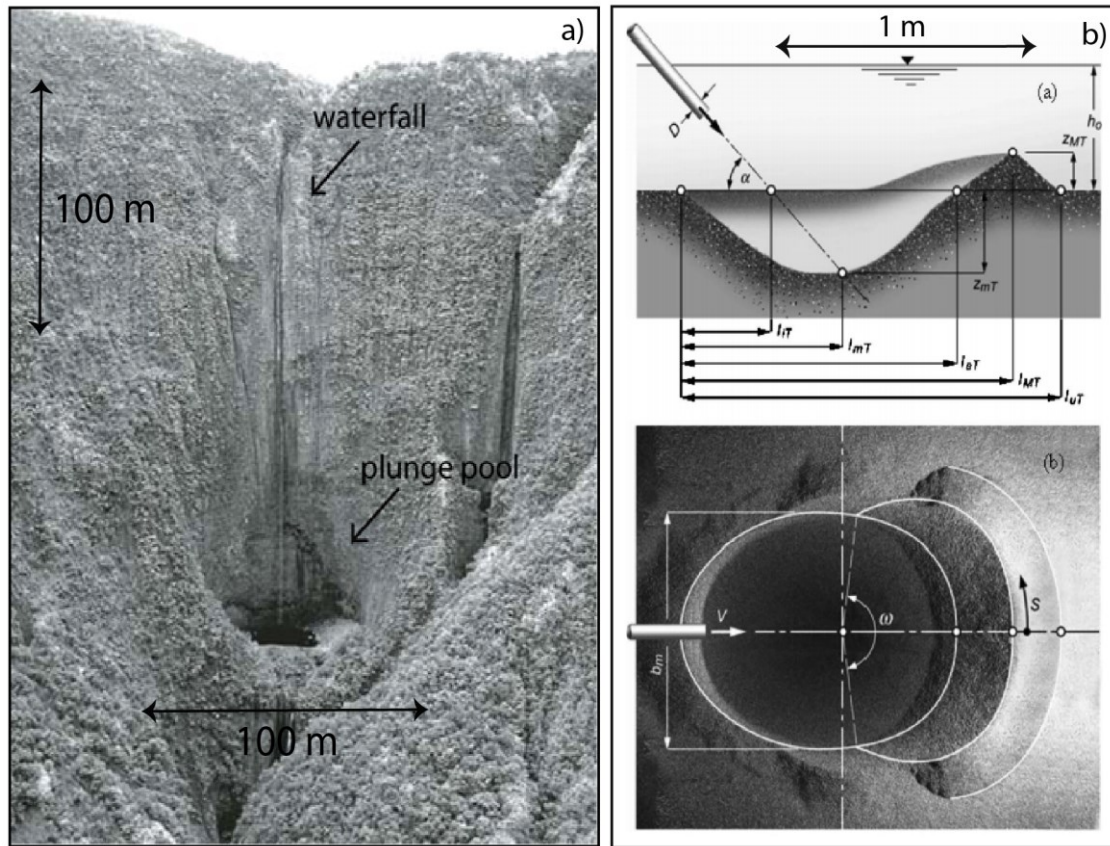


Figure 5.18: Examples of natural (box a, from Lamb et al., 2007) and laboratory (box b, from Pagliara et al., 2008b) plunge pools showing the same sub-circular plan-view geometry at different scales. The upper figure of box b displays the cross section geometry of an artificial plunge pool, showing strong similarity to the geometry of incisions B and C described in the text. Despite the difference in scale of up to three orders of magnitude, both Dover Strait and laboratory plunge pools display steeper upstream flank and gentler downstream flank characterised by two slope values.

5.7.2 When and how was the infill deposited?

The analysis of the depression infill revealed the sediments to be composed of fine grained, laminated material with different levels of stratification. In particular, incisions A and D display a highly stratified infill with internal layers, parallel geometry and lateral onlap contact on the flank of the incision. The acoustic character of the incision infill shows, on seismic reflection data, strong similarity to the infill of tunnel valleys which is characterised by fine-grained material. (Cameron et al., 1987; Huuse and Lykke-Andersen, 2000; Kluiving

et al., 2003; Lonergan, et al., 2006; Krohn et al., 2009). This, would suggest the sediments to be composed by fine grained material. This hypothesis is consistent with the interpretation of the infill to have lacustrine origin, as for the case of the upper, fine-grained facies of tunnel valleys infill and for the sediments filling the plunge pools interpreted by Meinsen et al. (2011) in north-western Germany. Nevertheless, this interpretation is purely speculative, as no direct sedimentological evidence is available for the incision infill. The only available sedimentological data is detailed by Destombes et al. (1975), who performed analyses on an incomplete core (VO50) located in the British Sector of the Dover Strait (See Section 3.2 for location). The core is incomplete (33 % recovery) and log data are only available for the upper 15 m of the infill where the analysed sediments were composed by reworked fragments of Cretaceous bedrock with no basal lag deposits and presence of microfossils, indicating an estuarine cold environment (Destombes et al., 1975). The palynological analyses performed on the core indicate a late Brorup interstadial age (63 to 61 kyr) of the sediments (Morzadec-Kerfourn, 1975). If a pre-Weichselian time for the erosion of the Fosse Dangeard is assumed, consistently with the Quaternary history of the Strait (e.g. Gibbard, 1985, 2007; Smith, 1985; Gupta et al., 2007, see Section 2.4) the upper part of the infill would therefore have been deposited in a later phase respect to the erosion of the incisions.

Due to the lack of sedimentological data, the absolute age of the basal and internal erosional surfaces of the incisions cannot be determined or correlated across the four incisions, as there is no chronological constraint on the age of the erosional events. Despite this, all reflectors display the similar acoustic character. This indicates that the erosional processes that carved the surfaces were likely to have been similar and were, if not synchronous, produced by related events in more than one pulse. The relative timing of the erosional events can be inferred from the presence of multiple erosional surfaces that suggest a first event to have carved the basal surface, followed by a second event responsible for the erosion of the internal surfaces into the infill. In addition, the change in the facies of the infill in correspondence of the internal erosional surface could possibly reflect a change in the depositional conditions or in the source of the sediments that fill the incisions, whose age remains unconstrained.

As the interpreted incisions are located on two different surfaces (Fig. 5.15), and are interpreted as the result of multiple erosions, it is necessary to reconstruct the relative and absolute timing of the surfaces into which the incisions are carved. As the Fosse Dangeard is located at the proposed megaflood spill-point, the age of the formation of the incisions is

strictly related to the reconstruction of the megaflood history and opening of the Dover Strait. The analysis of the age of the erosional surfaces on which the incisions are located is detailed in Section 6.2 of this work.

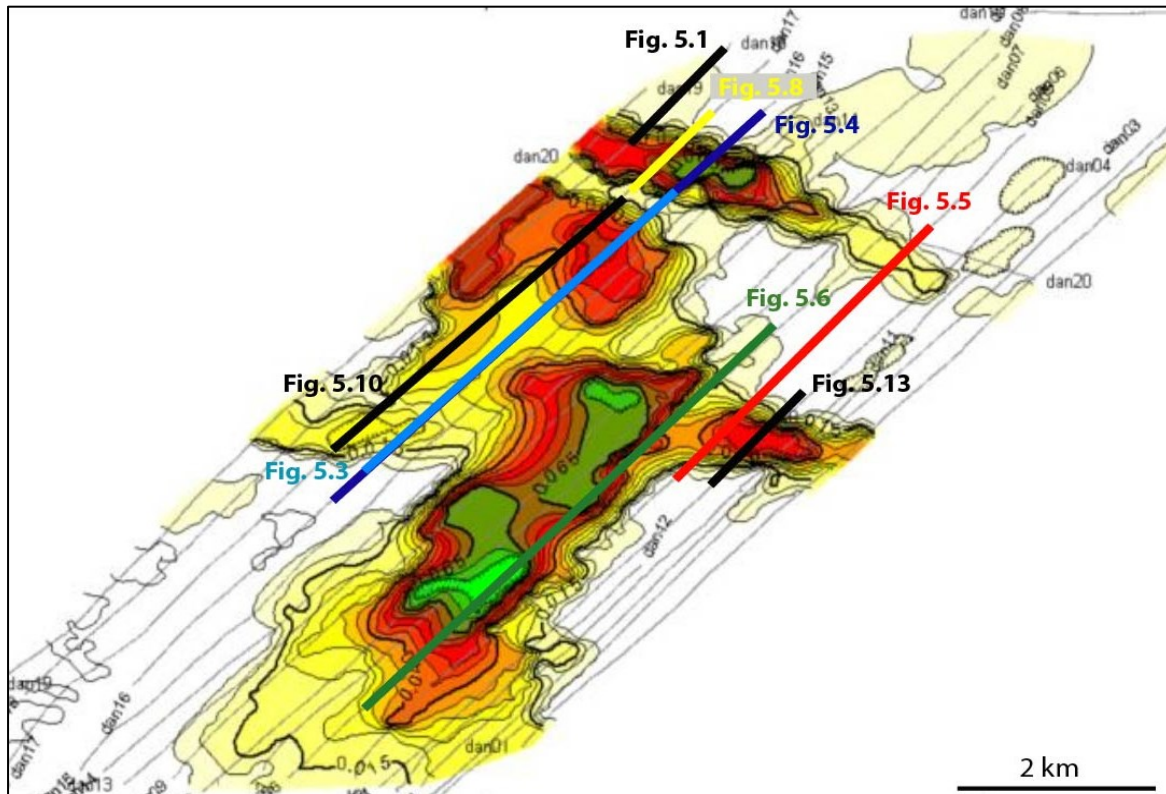


Fig. 5.19: Location of figures shown in this chapter. Colours indicate different figure number,

5.8 Conclusions

The following conclusions can be drawn from the performed morphometric and seismo-stratigraphic analyses of the Fosse Dangeard:

- The incisions that compose the Fosse Dangeard are interpreted as plunge pools carved into the bedrock by cascading water flows, accelerated by the elevation drop formed by the Chalk Escarpment. The depressions show strong similarity to the saucer-shaped, sub circular plunge pools mapped in other megaflood terrains on Earth, as well as small- scale natural and artificial plunge pools.
- The infill of the incisions is composed of (possibly fine-grained sediments) that display multiple acoustic facies and scoop-shaped internal erosional surfaces, whose geometry varies locally. Although it is not possible to correlate the erosional surfaces across the dataset, their presence indicates that the incisions and their infill were carved by multiple erosional events.
- While the lower part of the infill is acoustically semi-transparent facies, possibly indicating finer-grained sediments, the upper facies display higher reflectivity and more stratified geometry, onlapping on the basal erosional surface. The change in the character of the infill could indicate a change in source or depositional conditions of the infill during multiple events.
- Given the present elevation of ~100 m above MSL of the Chalk Escarpment, the up to 80 m deep interpreted plunge pools indicate a total elevation drop of at least 180 m for the flows that breached the Chalk dam located at the Dover Strait. Further analyses on the elevation and relative age of the surfaces on which the incisions are carved are detailed in Chapter 6.

6 . Geomorphology of megaflood features in the Northern Palaeovalley

6.1 Introduction and background

Following the interpretation of the Fosse Dangeard bedrock incisions at the Dover Strait in Chapter 5, this chapter illustrates the analysis of the morphology of the Northern Palaeovalley, extending from the Dover Strait to the Central English Channel. The merged bathymetry, composed of a variety of datasets including regional scale to high-resolution grids (see Chapter 3), allows for detailed geomorphological analyses the bedforms observed in the Northern Palaeovalley. The quantitative characterisation of the flood features allows for a comparison to similar flood bedforms present in other megaflood terrains and for the interpretation of the nature and magnitude of the processes that produced the mapped features.

In order to understand the processes that led to the formation of these bedforms, the comparison to other megaflood terrains has been made through the analogy between flood-bedform associations described in literature and the erosional features interpreted in the study area. In particular, previous studies on megaflood terrains (e.g. Baker, 1973a; 1978a; Herget, 2005; Lamb et al., 2008; Meinsen et al., 2011) show that flood-related bedforms are usually found in typical associations of channel-scale features. These associations of features are typically are composed features such as of kilometre-wide channels, streamlined erosional remnants, retreat cataracts, hanging valleys. The megaflood complexes previously studied in literature can be directly observed and studied in great detail as they are located in emerged areas. Kilometre-scale valleys present in megaflood terrains commonly indicate major flood drainage routes (Kehew and Lord, 1986; Kehew et al., 1990; Carrivick et al., 2004; Gupta et al., 2007; Meinsen et al., 2011), while palaeo-flow path is interpreted through the analyses of palaeo-flow indicators such as erosional grooves (Baker and Nummedal, 1978) and streamlined islands (Baker, 1978a; Rudoy, 2002; Lamb and Fonstad, 2010; Meinsen et al., 2011). Carved into the flood channel floor retreat cataracts, plunge pools and potholes (Baker and Nummedal, 1978; Rathburn, 1993; Rudoy, 2002; Carrivick et al., 2004) are usually found, carved by high magnitude flood flows through plucking, abrasion and cavitation (Sklar and Dietrich, 2004, 2006; Whipple et al., 2000; Baker, 2009a).

The analyses of flood bedforms will also focus on the characterisation of their spatial distribution and geometry in relation to bedrock geology, following the geological mapping detailed in Chapter 4. The cross-cut relationships of the interpreted flood surfaces into which the bedforms are carved allows for the reconstruction of the relative timing of the flood events, reconstructing the relative chronology of the flows that carved the observed flood surfaces and bedforms, including the Fosse Dangeard bedrock incisions described in Chapter 5. Finally, channel cross-profiles will be used for palaeo-hydraulic calculations, in order to estimate the magnitude of the palaeo-flood and compare the obtained peak discharge to those of major terrestrial megafloods, such as the Channeled Scablands (e.g. Clarke et al., 1984; O'Connor and Baker, 1992), Lake Bonneville (e.g. O'Connor, 1993), Altai Mountains (e.g. Baker et al., 1993; Herget, 2005; Carling et al., 2010) and Lake Agassiz (e.g. Clarke et al., 2004) megafloods.

6.1.1 Aims and Objectives

The aim of the present chapter is the mapping and interpretation of the kilometre-wide palaeo-channels and associated features carved into the English Channel seafloor. In order to do so, the chapter has the following objectives:

To map and interpret the role of the Lobourg Channel, located at the Dover Strait, and of the connected channels located at the proposed spill-point area, where the Fosse Dangeard is located.

- To characterise the geometry of the Northern Palaeovalley channel network and of their tributary valleys, analysing the channel cross-sectional profiles, from the breach-point to the downstream reaches. The channel profiles also permit a palaeo-hydraulic calculation of the peak discharge of the flow(s) that carved the observed flood-features.
- To map the morphology of the flood features present in the downstream sector and understand their relation to the Dover Strait area. A detailed analysis of the distal area is crucial in understanding the flood bedform spatial distribution and relation to bedrock geology.

- To characterise the distribution and geometry of palaeo-flow indicators such as streamlined islands and longitudinal lineations. This will allow for the reconstruction of the palaeo-flow path from the spill-point to the downstream area and for a comparison to similar features identified in other megaflood terrains on Earth.
- To understand the cross-cutting relationships between the mapped channels and related erosional surfaces, in order to constrain the relative timing of the flows that carved the Northern Palaeovalley channel network.

The objectives allowed for the following questions to be answered: 1) Was the Northern Palaeovalley and its erosional bedforms carved by a catastrophic flood? 2) Are the observed features similar to the flood-related bedforms interpreted in other megaflood terrains on Earth? 3) What is the palaeo-flow path and what is the relation between the upstream Lobourg Channel, located at the spill point area, and the channels in the downstream reaches? 4) Does the seabed geology and tectonics control the geometry and spatial distribution of the interpreted megaflood features? 5) What is the relative timing and number of events that carved the Northern Palaeovalley? 6) Can we estimate the magnitude of the flow and the peak discharge involved in the megaflood flows and compare it to other flood events occurred on Earth?

6.2 Geomorphology of the Northern Palaeovalley

The merged bathymetry permits a detailed mapping and analysis of the channel network and bedforms carved into the Northern Palaeovalley. The geomorphology of the mapped channels and seabed features is described in this chapter, together with the geomorphic analyses performed on the interpreted features whose line drawing is shown in Fig. 6.1. The first step of the performed analyses consisted of the identification of the channels that compose the Palaeovalley network (Section 6.2), in order to detail the channel network morphology and geometry, interpreting the palaeo-flow path. Secondly, the erosional surfaces were mapped, using their cross-cut relationships to constrain the relative chronology and number of events that carved the Northern Palaeovalley. In addition, the chapter presents the analysis of spatial

distribution and geometry of flood-features present on the channel floor, such as streamlined islands and grooves (Sections 6.3 and 6.4). Finally, the cross-profile geometry of the channels permitted an estimation of the peak discharge of the events (Section 6.5) that carved the Palaeovalley network through palaeo-hydraulic calculations.

The palaeovalley system is divided in three main areas: the Lobourg Channel area, the Central Channel area and the Channel Basin area. The Lobourg Channel area is the area that lies between the Southern North Sea, the Dover Strait and a fault, trending WNW-ESE, and bounding the platform present on the UK sector (Fig. 6.1) of the Strait. This area extends for ~100 km in NE-SW direction, parallel to the path of the channels. It has a higher seabed slope (up to 0.7°) respect to the downstream areas. Moving downstream to the Central area, the channel system trends E-W with a lower slope, ranging from 0.01 to 0.1° . This is associated to different lithology of the seabed in this area with respect to the Lobourg Channel, as the Central area is located in the Hampshire Basin, where Palaeogene rocks outcrop (see Chapter 4). This portion of the Palaeovalley is 90 km long and it is the first segment with streamlined erosional remnants.

Further south-west, the Channel Basin area is the portion in which the channels display anastomosed pattern, forming streamlined bedrock remnants between the channels. The merged dataset is, as illustrated in Chapter 3, subdivided into a British and French sector: while the British sector displays very high quality of the bathymetric data, the French sector is characterised by poor quality of the data, especially in the Channel Basin area (Fig. 6.1). For this reason, it was only possible to map the path of the main channel present in the area and represented by Channel C as the quality of the data does not allow for more detailed analyses.

Associated to the mapped channels, seven erosional surfaces (Fig. 6.1) were identified from the mapped channels that compose the Palaeovalley network. Surfaces were numbered from 1 to 7 according to their depth, starting from shallowest surface (interpreted as pre-flood platform) to the deepest (inner channels). Same number was given to more than one surface interpreted to have been originated during the same flood event. A brief description of each surface is given here.

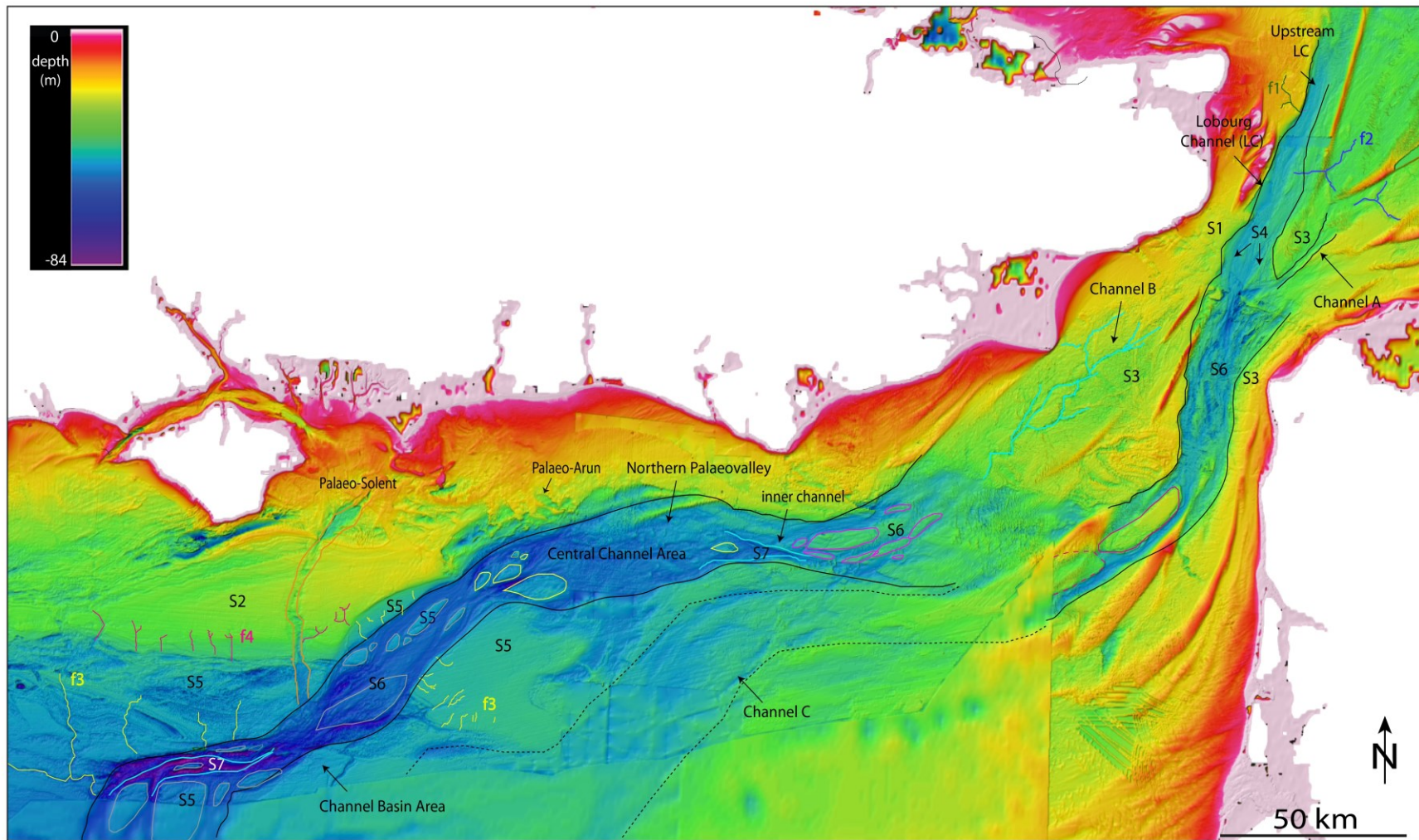


Figure 6.1: Line drawing of the interpreted channels and related mapped erosional features displayed with bathymetry in the background. The detailed description of all the labelled features is given in the text.

The description of the areal extent and elevation (depth) of each surface is given later in the text where each surface is described in detail:

- **S1 and S2:** These represent the areas interpreted as pre-flood surfaces. These are the offshore platforms present both upstream (S1) and downstream (S2).
- **S3:** This represents the outer margins of the Lobourg Channel. It is present on both flanks of the channel.
- **S4:** Inner margins of the Lobourg channel, present on both flanks of the channel.
- **S5:** Margins of the Northern Palaeovalley in the downstream area. S5 also includes the island-shaped remnants, carved by the flood into the same surface.
- **S6:** This is the Lobourg Channel and Northern Palaeovalley floor.
- **S7:** Identifies the inner channels carved into S6.

6.2.1 Lobourg Channel (LC)

The Dover Strait area is dominated by the presence of the Lobourg Channel (LC), interpreted as erosional surface S6 and crossing the Strait with a NE-SW direction. LC extends from the Southern North Sea up to the Central Channel area and displays kilometre-scale dimensions for all its length, as indicated by the two profiles in Fig. 6.2. A change is observed in the character and geometry of LC across the Strait, as also shown by the profiles in Fig. 6.2: two different areas (Upper LC and Lower LC) were therefore identified, divided by Middle-Lower Chalk boundary (Fig. 6.3).

Upper Lobourg Channel

Channel geometry

The Upper Lobourg Channel includes the portion of LC extending from the North Sea to the Middle-Lower Chalk Escarpment. Its most upstream portion (Fig. 6.1) in the Southern North Sea area is carved into Palaeogene rocks and is partially masked by seabed sediments: for this reason the channel is relatively shallow in this area, only changing to high-relief geometry as it enters the Strait where it is carved into the Middle and Lower Chalk formations (Fig. 6.2).

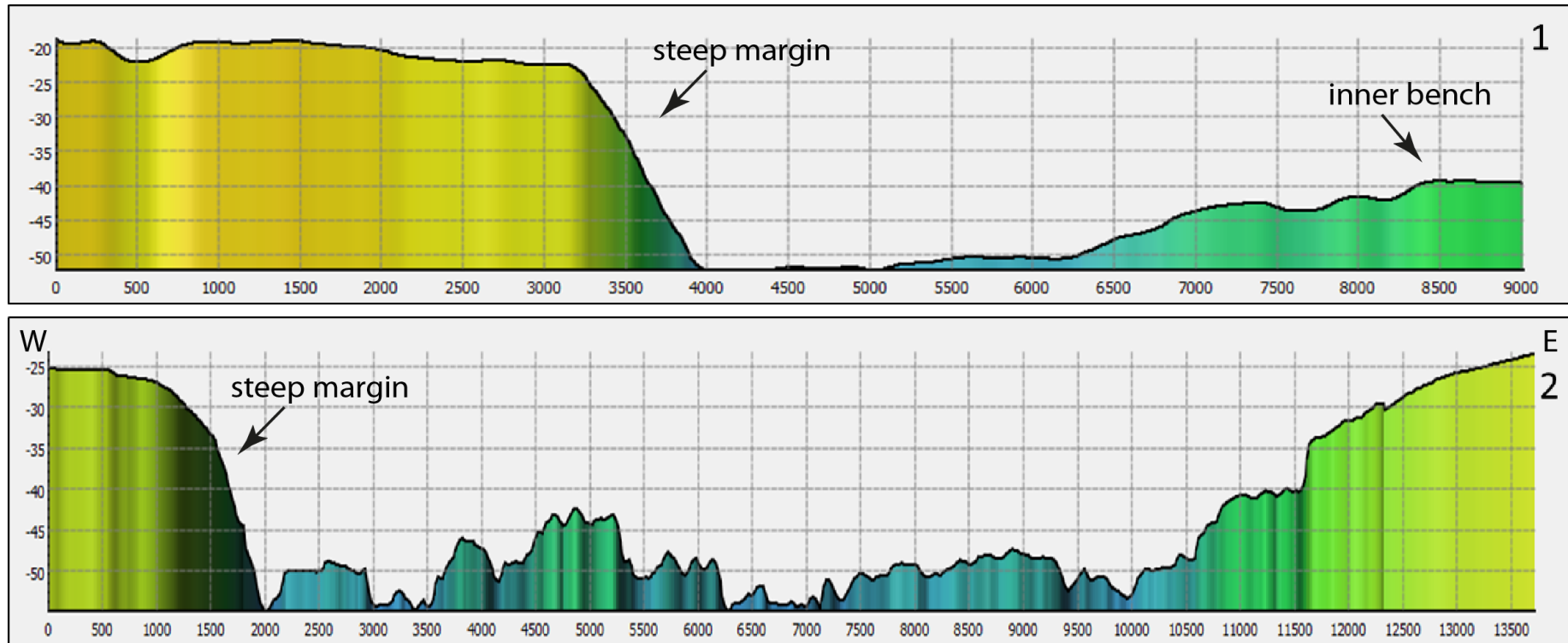


Figure 6.2: Two cross-profiles drawn across the Lobourg Channel, showing how the geometry of the channel changes from the upper to the lower portion of the Dover Strait. Profile 1 (Upper LC) shows a very steep western margin of the channel and a very poorly defined eastern bench, only represented by a bedrock remnant on the channel floor. The channel floor shows a very smooth profile. Profile 2 (Lower LC) displays a more symmetrical channel section with steep flanks. The channel floor shows a rough and irregular profile if compared to Profile 1. The location of the profiles is shown in Fig. 6.4. Colours indicate depth, ranging from yellow (shallow water, 20-25 m) to blue (50 m).

In the upstream area of the Strait (Fig. 6.2-6.4), the Lobourg Channel is 44 km long, 3.5 km wide, 30 m deep and displays rectilinear geometry. Its orientation is NNE-SSW, lying at a depth of 50 to 51 m.

The channel is generally characterised by width that is orders of magnitude larger than depth, as for the typical kilometre-scale megaflood channels (Table 1). Upper LC channel displays inner and outer benches represented by distinct erosional surfaces that are described in the following sections.

Channel margins – outer flanks

The flanks of Upper LC are strongly asymmetrical (Fig. 6.2) as the north-western flank is much steeper (slope value of up to 4°) respect to the south-eastern flank that displays a 10 m relief and a maximum slope of 0.8° . The erosional surfaces that represent the flanks of Upper LC in this area have been interpreted as S1 (north-western flank) and S3 (south-eastern flank), as shown in Fig. 6.3. S1 lies at a depth that ranges from 19 to 30 m and has a width of 5 up to 16 km in E-W direction, extending for 55 km across the.

While the western sector of the pre-flood shelf (S1) does not show evidence of flood-related surface, its eastern sector of the shelf shows evidence of scouring and high-energy erosion with same erosional pattern as the flood surface located further downstream and representing the bedrock platform that bounds Lower LC), of which it lies at the same elevation (~ 28 m). In addition, a set of erosional lineations indicating high-energy erosion are located on the bedrock remnant at the Middle-Lower Chalk boundary (Fig. 6.5). For detailed description of the erosional lineations see Section 6.4 of this chapter. The platform is carved by a series of small-scale channels labelled as f1 (an example is shown in Fig. 6.3). The channel has been interpreted as dendritic post-flood drainage channel as it displays the same geometry as other small-scale channels previously interpreted by Gupta et al. (2007) in the downstream area. (Fig. 6.1). Four sets of drainage channels, including those in the downstream portion of the Palaeovalley, have been identified in the dataset and labelled from f1 to f4. All the drainage channels have very small scale when compared to the kilometre-scale flood channels.

The south-eastern margin of the channel is formed by erosional surface S3 (Fig. 6.3) which consists of the 13 km-wide bedrock remnant that lies between the Lobourg Channel and channel A. Channel A flows into the Lobourg Channel in the eastern Dover Strait area,

flowing with a NE-SW direction parallel to LC. Its length is 25 km, its width ranges between 800 m and 1500 m and its relief is much lower than those of the main channel, ranging from 4 up to 10 m where the channel joins the Lobourg Channel. A set of two small-scale channels labelled as f2, with character very similar to channel f1, is visible on the north of channel A, although the channels are partially covered by sand banks that drape the surface extending north-eastward for more than 40 km across the northern Dover Strait. The two channels interpreted as f2 are composed of a first rectilinear tract that bifurcates into two branches: the westernmost channel flows with a NE-SW direction for 8 km and then bifurcates into two 6 to 7 km long branches that are up to 3 m deep. The western branch of the channel seems to be carved into more than one surface, cutting across the outer (S3) and inner (S4) margins of the Lobourg Channel and then disappearing on the Lobourg Channel floor (S6). The cross cut relationship of the surfaces indicates that f2 has probably been carved after the erosion of the surfaces that are now the channel inner and outer margins, associated to the formation of the Lobourg Channel.

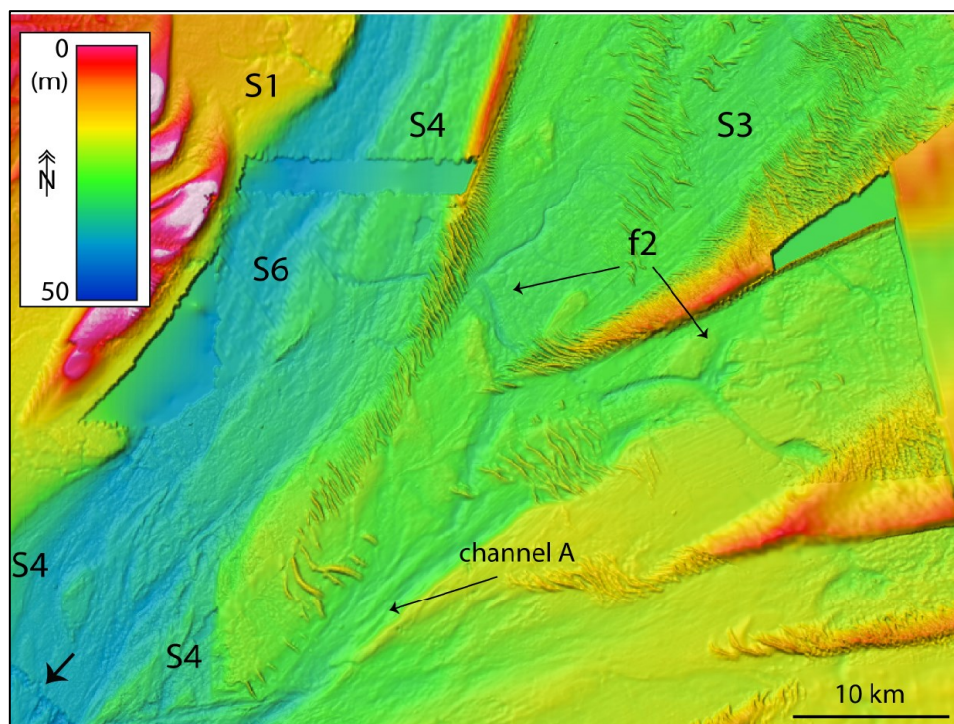


Figure 6.3: Bathymetric image of the outer and inner margins of the Upper portion of Lobourg Channel, bounded on the south-west by the Middle-Lower Chalk boundary visible in the lower-left corner of the figure and pointed by the black arrow. See Fig. 6.16 for location of this figure.

Channel margins – inner benches

Two inner benches are present on both flanks of the Upper LC, located upstream of the Middle-Lower Chalk boundary and represented by surface S4 on both flanks of the channel (Fig. 6.3). The north-western (inner) margin lies 15 m below the platform (outer margin) at a depth that ranges from 43 to 47 m, and 10 m above the channel floor, forming a narrow bench (10 km of length and 1 km of width). The south-eastern margin is less defined and no outer and inner margins are present. The channel flank is here identified by as a series of discontinuous bedrock remnants that lie at a depth of 43 to 45 m. The remnants are all elongated in the direction of the channel axis (NNE-SSW), the biggest of which is 7 km long, 2 km wide and lies at the same elevation of the bench on the opposite flank of the channel.

Lower Lobourg Channel

Channel geometry

Just downstream of the area where the tributary channel joins the Lobourg Channel, a step is carved into the Lobourg Channel floor (Figs. 6.3 and 6.4): this scarp, visible at the seabed, represents the offshore expression of the Middle-Lower Chalk escarpment, marking an up to 5 m-high step at the seafloor and located at the boundary between the two formations.

Downstream of the escarpment, LC shows a different orientation, geometry and seabed acoustic texture respect to the Upper portion, due to the change in lithology from the more competent Upper and Middle Chalk to the more erodible Lower Chalk. Here, the channel trends NNW-SSE for 25 km south of the Chalk Escarpment, changing into an ENE-WSW direction for 30 km in its final tract, where LC merges into the Northern Palaeovalley. Lower LC is 10 km wide and 15 to 25 m deep, displaying a characteristic irregular acoustic texture of the seafloor that contrasts with the smooth acoustic texture observed in the upstream portion where the Middle Chalk Formation outcrops at the channel floor (Fig. 6.4).

Channel margins

Lower LC is bounded on both sides by erosional surfaces identified as the outer margin in the Upper portion. On the north-western margin this surface is a bedrock platform that is 25 m elevated from the channel floor. This surface is the downstream continuation of the platform and lies at a depth of 28 to 35 m. On the south-eastern, the channel flank (Fig. 6.1) is poorly imaged as it is located in the low resolution portion of the dataset. Nevertheless, a small

portion of the channel flank is visible in higher resolution data, as shown by the profile in Fig. 6.2. Here the channel flank has the same elevation as the north-west channel margin, lying at ~30 m depth. A detailed description of the bedrock platform is given in Section 6.2.2 of this chapter.

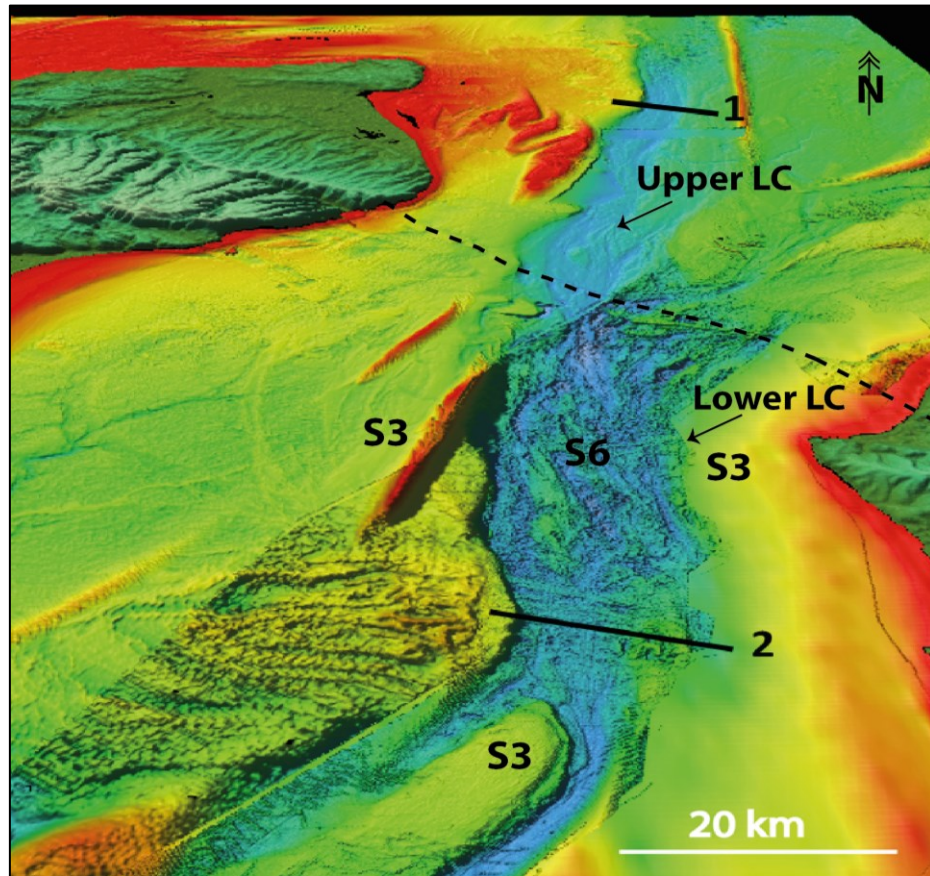


Figure 6.4: Location of profiles shown in Fig. 6.2 and boundary between Upper and Lower LC separated by the Middle-Lower Chalk escarpment (dashed line). This figure also shows surface S3, representing Lower LC flanks. Colour-scale is the same as in Fig. 6.1.

Lobourg Channel flood features

As the Lobourg Channel floor is sediment-free in its upper and central portions, multiple erosional features measuring kilometres of length are observed on the channel floor (Fig. 6.5). From the high-resolution dataset, multiple sets of longitudinal lineations composed of trough and ridges are visible at the seabed (Fig. 6.5), located upstream of the Chalk Escarpment and carved into the Middle Chalk in Upper LC. As shown by Fig. 6.5, the lineations are carved into two different surfaces represented by the channel floor (surface S6)

and on the channel north-western flank (s1). A detailed description of the longitudinal lineations is given in Section 6.4 of this Chapter.

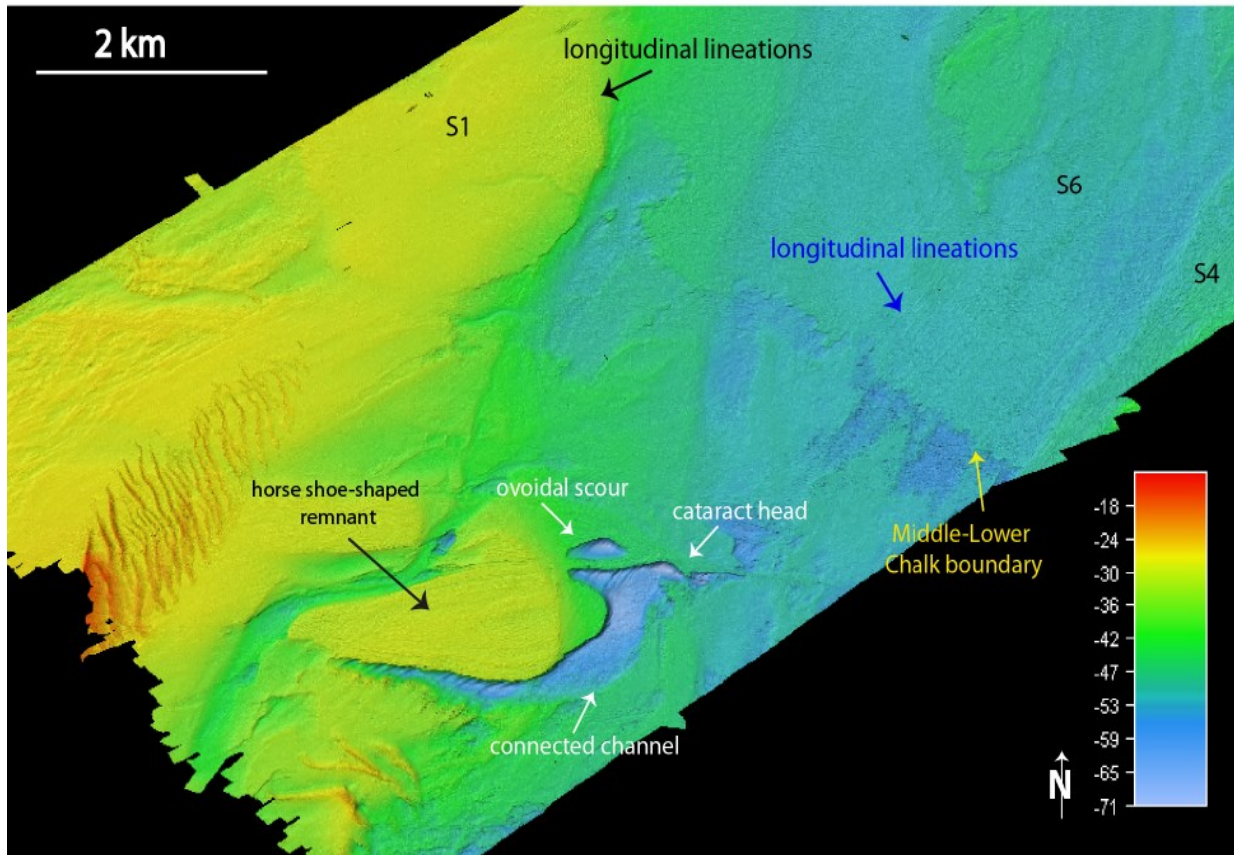


Figure 6.5: Image of the cataract complex located just downstream of the longitudinal lineations at the base of the Middle-Lower Chalk boundary. Depth is in m. See Fig. 6.16 for location of this figure.

In addition to the longitudinal lineations, 3 km downstream of the escarpment at 45 m of depth, another kilometre-scale erosional feature is visible on the channel floor (Fig. 6.5). This is a >20 m vertical bedrock wall carved into the Lobourg Channel floor: the head of the incision is 1200 m wide and runs parallel to the NW-SE oriented Gault-Lower Greensand contact that is also visible from bathymetry (Fig. 6.5) and follows the axis of the monoclinical fold present in this area (see Section 5.2). The headwall of the incision shows a theatre-shaped geometry, neatly carved into the channel floor. Connected to the headwall is a 3.5 km-long, 600 m-wide and 20 m-deep channel that bounds a horseshoe-shaped bedrock remnant on its western margin (Fig. 6.6).

The channel trends N-S, taking a westerly direction at its end where it terminates into a rounded-shaped incision into the bedrock. Here, another channel is present (Figs. 6.5-6), located on the north-western flank of the horseshoe-shaped remnant (Fig. 6.5) and displaying a NE-SW direction. This second channel is 1.8 km long, 350 m wide and up to 12 m deep, bifurcating to the north-east into two branches that also show theatre-shaped heads (Fig. 6.6). In particular, the westernmost branch displays stubby head represented by a 200 m wide and 14 m deep elongated scour that suggests plunge pool erosion at the base of the headwall.

The profile in Fig. 6.6 across the head of the vertical incision shows evidence of a plunge pool carved at the base of the headwall and probably formed by plunging water. No upstream channels are present above the headwall, indicating that the 20 m deep incision was formed by erosion of retreating flows. Headward migration of the cataract headwall formed the two channels. An ovoidal scour is present at the head of the cataract, located upstream of the deeper channel. The scour is located 120 m upstream of the cataract headwall and is 600 m long, 300 m wide and 20 m deep with its major axis that parallels the cataract.

This cataract complex, given its steep vertical wall carved into the channel floor and the theatre-shaped geometry of the headwall, is very similarity to the Dry Falls cataracts observed in the Channeled Scablands megaflood terrain (Baker, 2009a, 2009b). These consist of two, 200 meter-high, retreat cataracts carved by retreating flood flows that display kilometre-scale longitudinal lineations carved at its head and plunge pools at their base (Baker and Nummedal, 1978). Both Scablands and Lobourg Channel cataracts, despite the difference in scale, lack upstream channel connection. This, taken together with the presence of erosional lineations present on the channel floor just upstream of the cataract, and plunge pools located at their base, indicates that the Lobourg Channel cataract is likely to have been carved by the same process that carved the Dry Falls, formed by headward (upstream) migration of the cataract head during high magnitude flow erosion (Baker, 1978b). In addition, available seismic data (David Garcia-Moreno, pers. comm.) collected where the cataract is located show that both the headwall and the plunge pool at its base are carved into bedrock. No sediments are present at its base. This, taken together with the presence of bedrock plunge pools, indicates that high-energy flows would have eroded the cataract as the formation of the 20 m deep and 1.2 km wide headwall into bedrock can only be explained as product of high magnitude, catastrophic flows with powerful erosional forces.

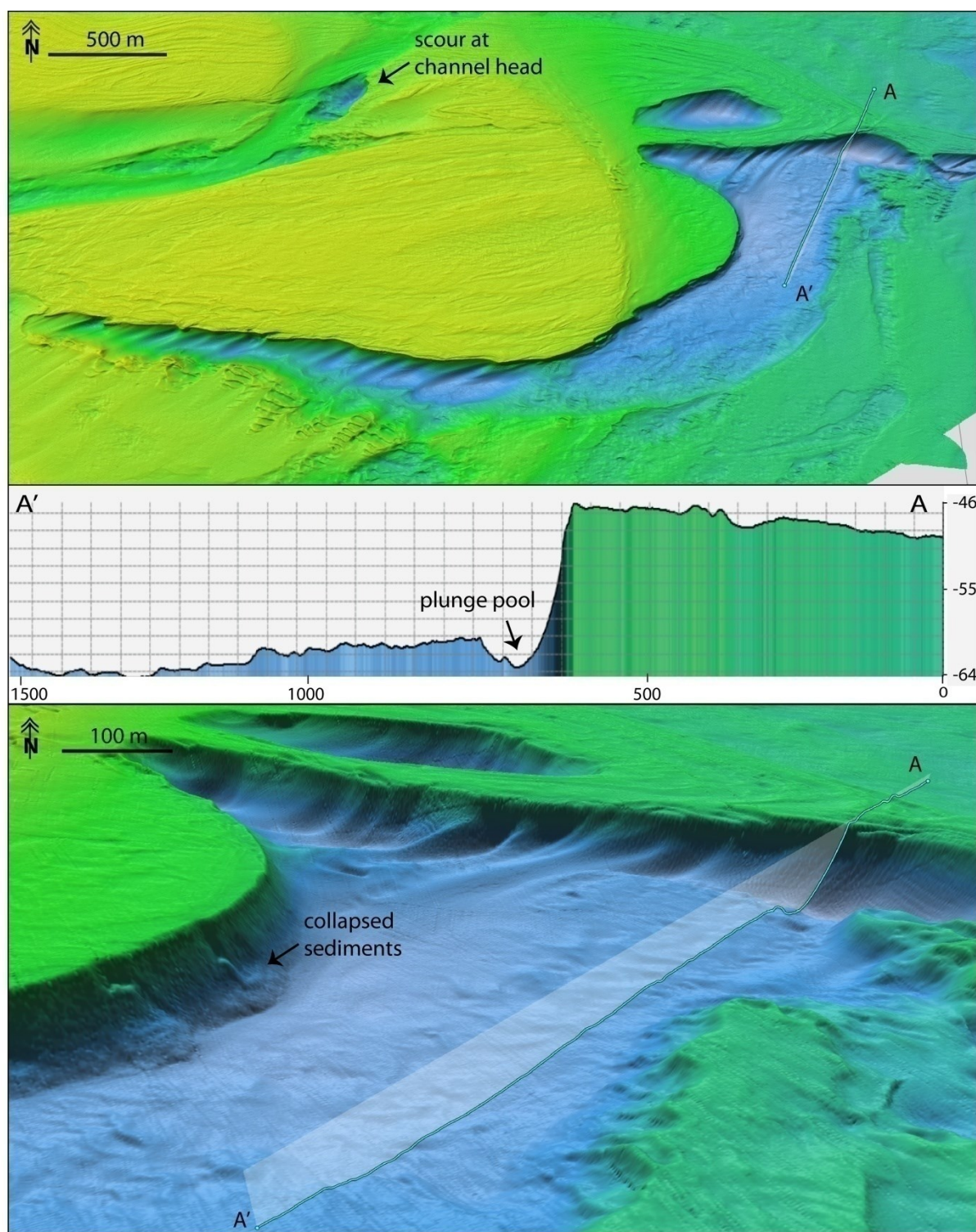


Figure 6.6: Image of the cataract and the connected downstream channel carved into bedrock, showing a profile across the 20 m high cataract head. Collapsed sediments are present at the base of the cataract wall and at the base of the channel north-western margin. Depth and distance on the profile are in m. Colour-scale is the same as in Fig. 6.5.

Bathymetric expression of the Fosse Dangeard incisions

The portion of the Lobourg Channel floor or where the cataract is located also corresponds to the area where the incisions that compose the Fosse Dangeard are present, as shown by the contour map in Fig. 5.16 (Chapter 5, where the incisions composing the Fosse Dangeard are detailed). The high-resolution bathymetry analysed in this chapter was overlapped to the contours of the Fosse Dangeard shown in Fig. 5.16, in order to analyse the acoustic character of the seabed and the incisions (Fig. 6.7).

Although no clear bathymetric expression of the incisions is visible, as they are completely sediment filled, a contrast in the acoustic character of bedrock and sediment filled depressions is expected to be visible from bathymetry. As demonstrated in Chapter 4 of this work, lithological boundaries are usually marked by strong changes in the seabed acoustic character that should therefore reflect the passage from bedrock to the incision infill sediments.

On the French sector of the Strait (eastern sector), the 2 m bin size bathymetry displayed in Fig. 6.7 shows how the depressions are located in those portions of the Lobourg Channel floor that show a rough and irregular surface. The perimeter of the depressions is not clearly recognizable at the seafloor, which is characterised by an alternation of elevated bedrock remnants and more depressed areas that mark scarps with amplitude of 0.5 to 1.5 m at the seafloor. The incisions are located north (incisions A) and south of a bedrock ridge that is visible from the bathymetry and that corresponds to the Gault-Lower Greensand contact. This boundary forms a ridge which is up to 10 m high at the seafloor, outcropping at the near the top of the monocline fold present in this area (see Section 5.2), with a ENE-WSW parallel to the fold axis. The monocline, as shown from the seismic data, bounds the south-western flank of depression A and the north-eastern flank of depressions B. However, not all the depressions show a correspondence between their boundary and bedrock ridges or remnants visible at the seafloor. As the incisions are completely sediment filled and the seafloor displays a very irregular acoustic texture, the contrast at the seafloor is only visible in some areas and it would not be possible to map the incisions from bathymetric data only.

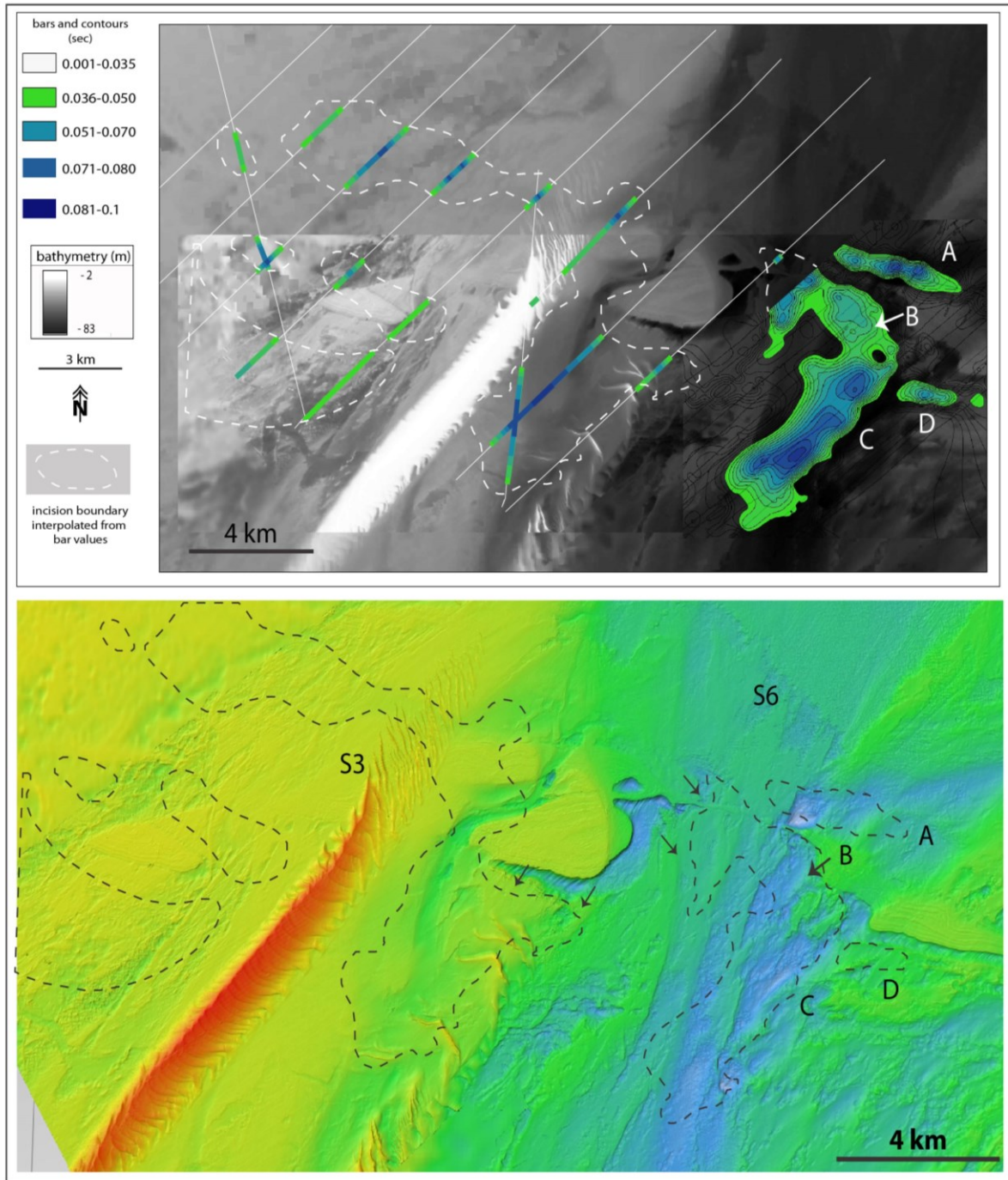


Figure 6.7: Contours (French sector) and coloured bars (British sector) of the incision basal surface depth (top) and 2 m bin size bathymetry overlapped to the incision boundary (lower box). The depressions in the French sector are labelled with the same names as in Chapter 5. The black arrows in the lower box indicate how the margin of some of the depressions is bounded by bedrock erosional remnants visible at the seafloor. Colour-scale is the same as in Fig. 6.5, indicating that S3 lies ~25 m above S6.

As illustrated in Chapter 5, the seismic data collected in the same area in the British side of the Strait displays presence of incisions with the same character as the Fosse Dangeard (David Garcia Moreno, pers. comm.). This suggests that the formation of the observed erosional features carved into the Lobourg Channel floor, and the channel itself, postdate the erosion of the Fosse Dangeard incisions. This is also indicated by the Lobourg Channel floor truncating the incision infill, as shown by seismic data interpreted in Chapter 5. This is consistent with the incisions on the British side of the Strait being located on the bedrock platform (as displayed in Figs. 5.16 and 6.7). Bathymetric data indicate that the bedrock platform lies at least 25 m above the Lobourg Channel floor, where the incisions on the French sector are located. The presence of the incisions on both the platform and the channel indicates that the incisions were carved into the older surface (platform) before the Lobourg Channel was eroded. This is also consistent with seismic evidence showing the presence of a bedrock incision located at the head of the retreat cataract and truncated by the Lobourg Channel surface that was eroded after the bedrock incision formed.

6.2.2 Lobourg Channel-Northern Palaeovalley junction

Further downstream into the Lobourg Channel, a giant island (Fig. 6.8) composed of bedrock and characterised by a streamlined geometry is present on the channel floor. It is the biggest streamlined island mapped in the study area. This feature is 16 km long, 4 km wide. It stands up to 25 m above the channel base, elongated in an ENE-WSW direction, parallel to the channel axis. A detailed description and interpretation of the streamlined islands present in the Northern Palaeovalley is given in Section 6.3 of this chapter. The island has a 6 m scarp on its surface coincident with the fault that crosses the channel floor in this area (Fig. 6.8). The fault, which separates the Wealden and the Hampshire basins, also determines the contact between the Wealden and London Clay formations, characterised by different levels of resistance of erosion. Carved into the Wealden Formation, the north-eastern side of the island lies at ~27 m of depth and has higher amplitude at the seabed respect to the more eroded southern side of the island that lies at ~32 m and is carved into Palaeogene rocks of the Hampshire Basin. Moving towards the south-west, another bedrock remnant with the same acoustic character and elongated geometry of the streamlined island seems to be present (dashed pink line in Fig. 6.1), lying at almost the same elevation (~34 m) as the south portion of the streamlined island. This feature is at least 10 km long, although it is not possible to

observe its southern and western margin, as the area is covered by poor-resolution data and is almost completely draped by modern seabed sediments. Here, the channel floor becomes gradually filled with modern seabed sediments composed of sediment waves and sand banks with up to 35 m relief (Figs. 6.1 and 6.8). Despite the thick sediment cover, the NE-SW orientation of the observed bedrock remnants and the geometry of the Lobourg Channel show that this area represents the junction of the Lobourg Channel and the Northern Palaeovalley.

6.2.3 Channel B and bedrock platform

The bedrock platform that further upstream bounds the western margin of the Lobourg Channel, is cut by channel B (Fig. 6.8). The platform displays 32 km of width at its widest point and extends for more than 38 km in NE-SW direction, dipping southwest and representing the steepest of the mapped surfaces with slope value of 0.2 up to 0.4°. The eastern margin of the platform lies at a depth of 26 to 35 m and displays the same elevation as the giant streamlined island located in the Lobourg Channel floor, whose surface lies at 27 to 32 m depth, and of the bedrock remnant that lies at 35 m depth. The profile in Fig. 6.9 displays how the island and the platform show the same elevation, suggesting that the elongated remnant and the island could represent erosional remnants of the bedrock platform. This erosional surface probably formed a wider bedrock platform across the Strait until it was partially eroded by the Lobourg Channel, entering the Palaeovalley from the east, as suggested by the eroded the remnants whose orientation is parallel to the channel floor axis.

On the western sector of the bedrock platform, channel B is carved into the platform for more than 40 km (Fig. 6.8), at a depth that ranges from 30 to 35 m parallel to the Lobourg Channel. The channel is characterised by an almost rectilinear geometry and bifurcating branches. Its upstream portion is composed of two main tributaries, joining into a single bigger channel and displaying no connection to an upstream or onshore river drainage system.

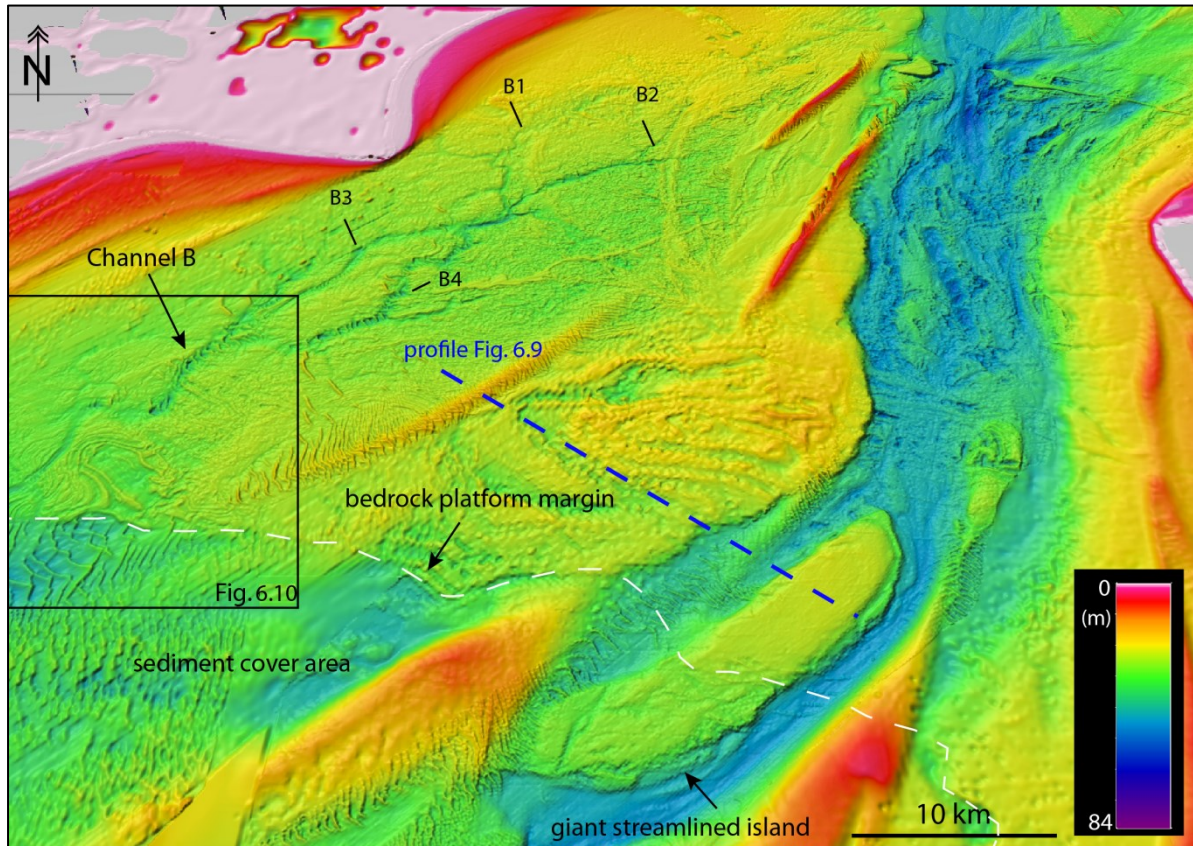


Figure 6.8: Southern view of the Lobourg Channel turning westwards into the Northern Palaeovalley in the area dominated by seabed sediments, south of the fault (dashed white line) that bounds the bedrock platform.

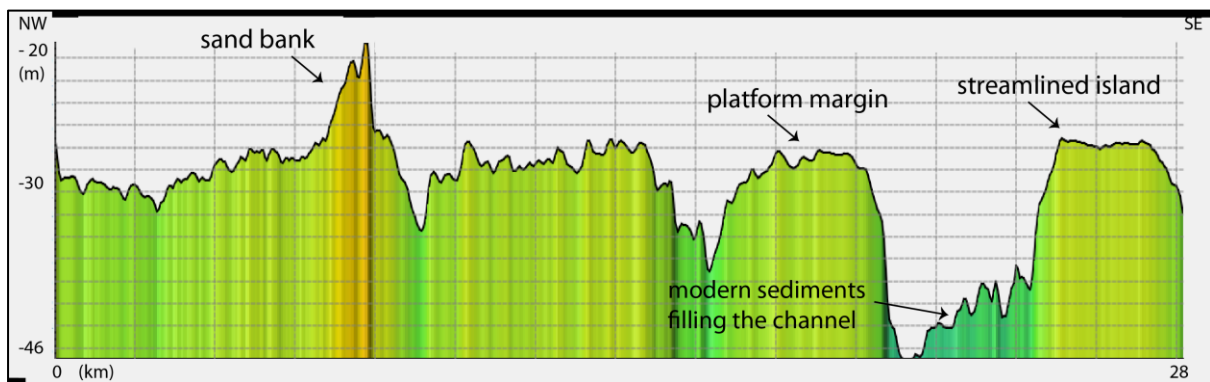


Figure 6.9: Depth profile of the bedrock platform and the streamlined island, indicating that both surfaces lie at the same elevation. The location of the profile is shown in Fig. 6.8.

The seismic reflection data of the Central Channel dataset (Fig. 6.10) show that sediments are present under the base of channel B, deposited into a palaeo-channel. Fig. 6.10 gives evidence of the presence of an older, sediment-filled channel, up to 30 m below the seabed, at the base of channel B. Fig. 6.10 displays the 1 km wide and 15 m deep sediment filled

channel subsequently reoccupied by the modern channel carved into the sediments. The available seismic also indicates that the modern channel is incised into the palaeo-channel for all its length, following the same path. This indicates that palaeo-channel B, like the modern channel, has no upstream onshore connection to the river drainage system in its upstream tract.

The upstream portion of channel B is composed of two branches (Fig. 6.8): B1, which is 7.5 km long and has a 5 m relief, and B2, which extends for more than 12 km and has an up to 8 m relief. These two channels are connected to two minor branches that are incised into the bedrock and show a variable relief that ranges from 3 to 5 m and display a length of 2 to 3 km. Where B1 and B2 join, the channel shows a rectilinear geometry for almost 2.5 km after which it divides again in two main branches, flowing separately for 12 km with an almost parallel direction. Although the two branches are likely to have been active more than once, as suggested by the presence of a palaeo-channel below the incision visible at the seabed, it is not possible to establish if the two branches were active at the same time. This is due to lack of cross-cutting relationships indicating a relative timing of the two erosional surfaces. A bedrock remnant is bounded by the two channel branches, B3 (up to 7 m relief) and B4 (5 m relief). Both branches have a tributary channel coming respectively from NW and SE and joining the main channel.

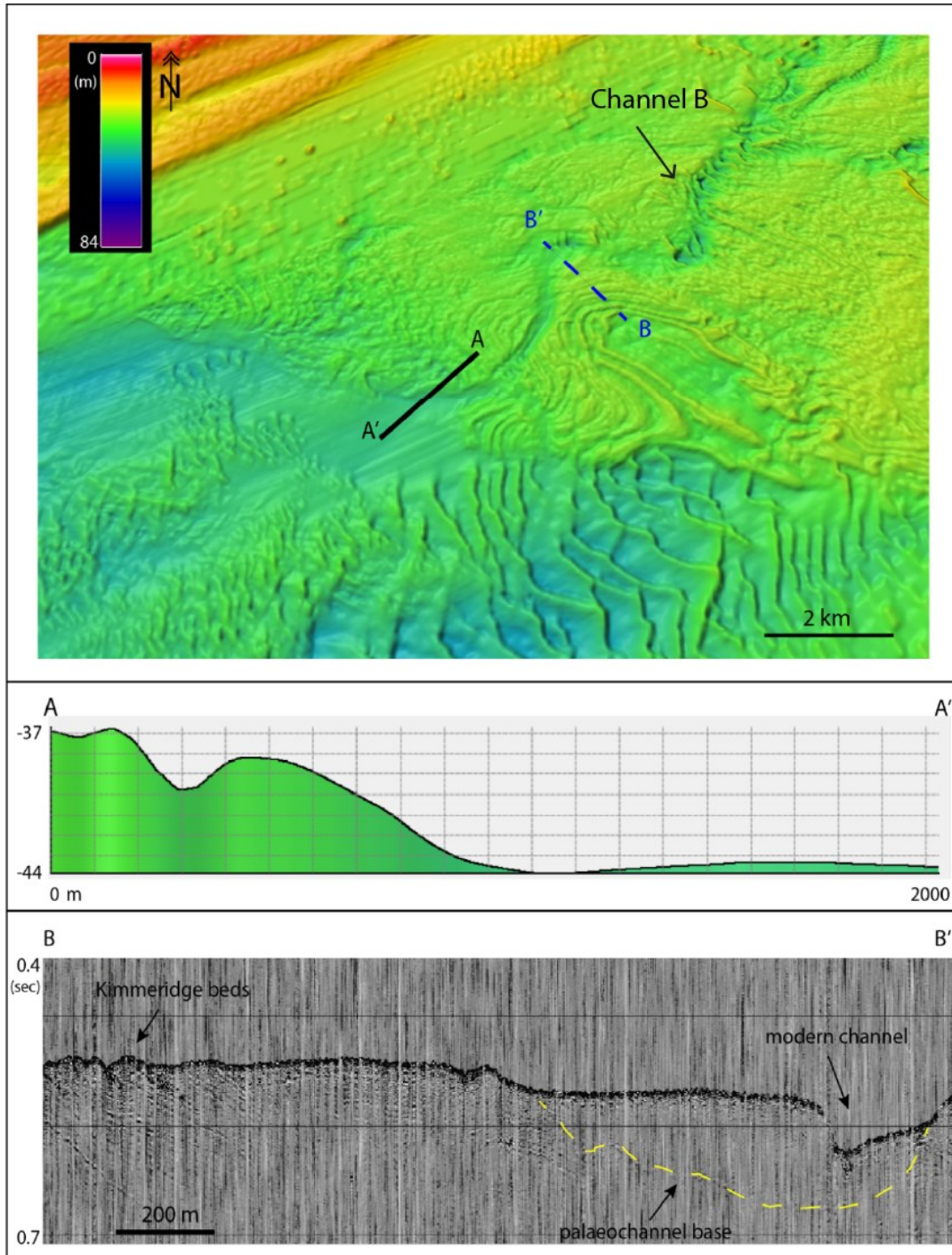


Figure 6.10: Bathymetric image of the lower portion of channel B. The two profiles show a cross profile of the bedrock platform margin (A-A'), showing the truncated outlet of the channel B and the seismic profile (B-B') across the reoccupied palaeo-channel.

The final tract of channel B is characterised by a more anastomosed geometry of the channel that develops for 11 km, after which the channel ends at the fault with NW-SE direction that bounds the southern portion of the bench. Here, the channel seems to be truncated by the

Lobourg Channel that joins the Northern Palaeovalley (Figs. 6.8 and 6.10) at the termination of the platform, as suggested by the 5 m step between Channel B and the Lobourg Channel floor (Fig. 6.8). The channel path in this area was probably driven by the presence of the fault that bounds the platform, creating a preferential direction for the Lobourg Channel to enter the Palaeovalley and truncating the platform and channel B at the structural boundary between Wealden and Palaeogene rocks. Truncated channel B represents an example of typical hanging valley (Baker, 1978a; Carling et al., 2009b). Hanging valleys represent truncated channels that lie on elevated surfaces, truncated by younger erosional surface, as in the case of channel B, eroded into surface S3 that lies 5 m above the Palaeovalley floor, truncated by the Lobourg Channel that joined the palaeovalley from the east (Fig. 6.11).

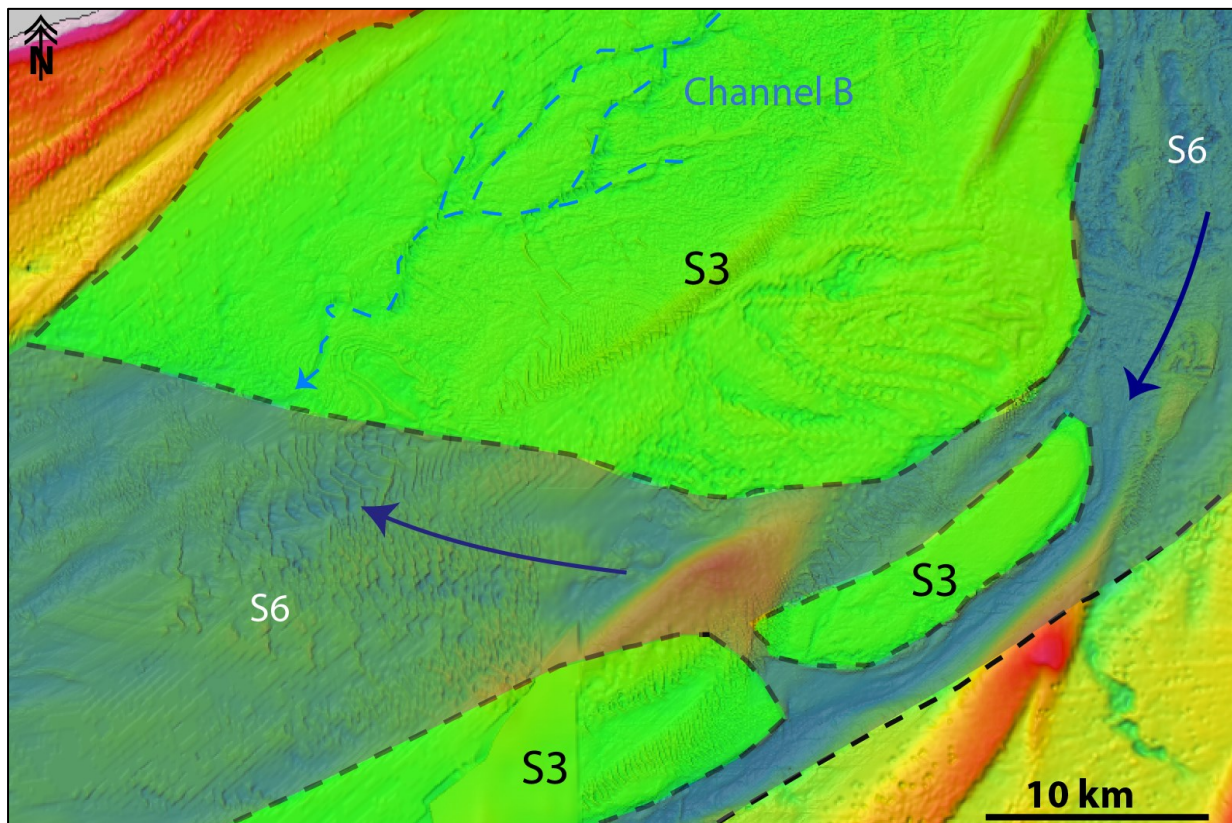


Figure 6.11: Interpretation of the erosional surfaces in the area where the Lobourg Channel enters the Northern Palaeovalley. Bedrock platform (S3) represents an older flood surface, on which post-flood Channel B was eroded after the flood events. A second erosion event is represented by the Lobourg Channel (blue surface and blue arrow indicating the direction of the palaeo-flow) carving the bedrock remnants and truncating channel B that lies 5 m above the Palaeovalley floor. Colour-scale for the bathymetry is the same as in Fig. 6.10.

The presence of channel B and of the two elongated remnants, also carved into the bedrock platform, indicate that the Lobourg Channel (S6) entered the Palaeovalley, eroding the remnants into the bedrock platform and truncating channel B (Fig. 6.11).

Although Channel B is interpreted as an hanging valley, there is no evidence of the channel being a flood channel. Its dimension would more likely indicate a fluvial origin, representing a normal, integrated drainage channel carved sub-aerially on the bedrock platform as for the case of drainage channels f1-f4. Both these smaller-scale channels and Channel B display lack of connection to the onshore river system as their most recent incision post-dates the erosion of the flood surface on which they are carved. In the case of Channel B, evidence of reoccupied palaeo-channel would indicate multiple erosion, possibly at different times. As two flood events are suggested, Channel B may have been carved at least twice, with its latest incision post-dating the two floods.

Importantly, the interpretation of Channel B as drainage channel suggests that it can be used as a flood high-water mark for the estimation of the platform channel depth during the flood.

6.2.4 Central Channel Area (CCA)

Channel geometry

This portion of the Northern Palaeovalley (Fig. 6.1) is characterised by a sharp change in the channel geometry and in the seabed character, due to a change in the seabed geology as the CCA is located in the Hampshire Basin where Palaeogene rocks outcrop at the seabed. The change from Chalk to Palaeogene rocks leads to a sudden change in the geometry of the channel and of the morphology of the bedforms carved into the channel floor. The CCA, which is 90 km long and 12 to 30 km wide, is the central portion of the Northern Palaeovalley: the channel here is much wider than in the Lobourg Channel area. The channel here has the same maximum depth (25 to 30 m) but is wider (20 to 25 km). Other than being much wider, the channel also has a flat floor, with almost no slope gradient. The direction of the flow, coming from the north-east through the Lobourg Channel and from the Tributary channel, also changes shifting from NE-SE to an E-W direction. The eastern sector of the CCA shows a thick sediment cover that extends for 30 km on the eastern side of the CCA, composed by the sediments of the sand wave field deposited in this area of tidal-dominated

sediment convergence (Fig. 6.8). West of the sand waves the channel floor is sediment free and displays a group of streamlined islands (Fig. 6.12) carved into the bedrock (see Section 6.3 for a detailed description of the streamlined islands).

Channel margins

In the Central Channel area, both flanks of the channel are represented by inner bench of the Palaeovalley, almost entirely eroded. These surfaces (S5, Fig. 6.1 and 6.12) are heavily eroded in comparison to the same surface present in the adjacent Wealden (on the east) and Channel (on the west) basins, where more competent rocks outcrop at the seabed. On the north, a group of elongated bedrock remnants is present on the east of the area where the Palaeo-Arun flows into the main channel (Fig. 6.12). The remnants lie at 45 to 47 m depth and are flat and smooth character if compared to the underlying channel floor. Their dimensions (the bigger measuring 7 km of length and 2.5 km of width) and acoustic texture is very similar to the streamlined islands present in the area (located at the same elevation as the remnant and therefore labelled as surface S5), although the remnants display a rectangular, elongated geometry rather than a teardrop shaped-geometry typical of the islands.

The southern margin (S5) is composed of bedrock remnants, lying at 46 to 50 m of depth. Here, the bedrock remnants divide the Northern Palaeovalley from Channel C that flows in the French sector.

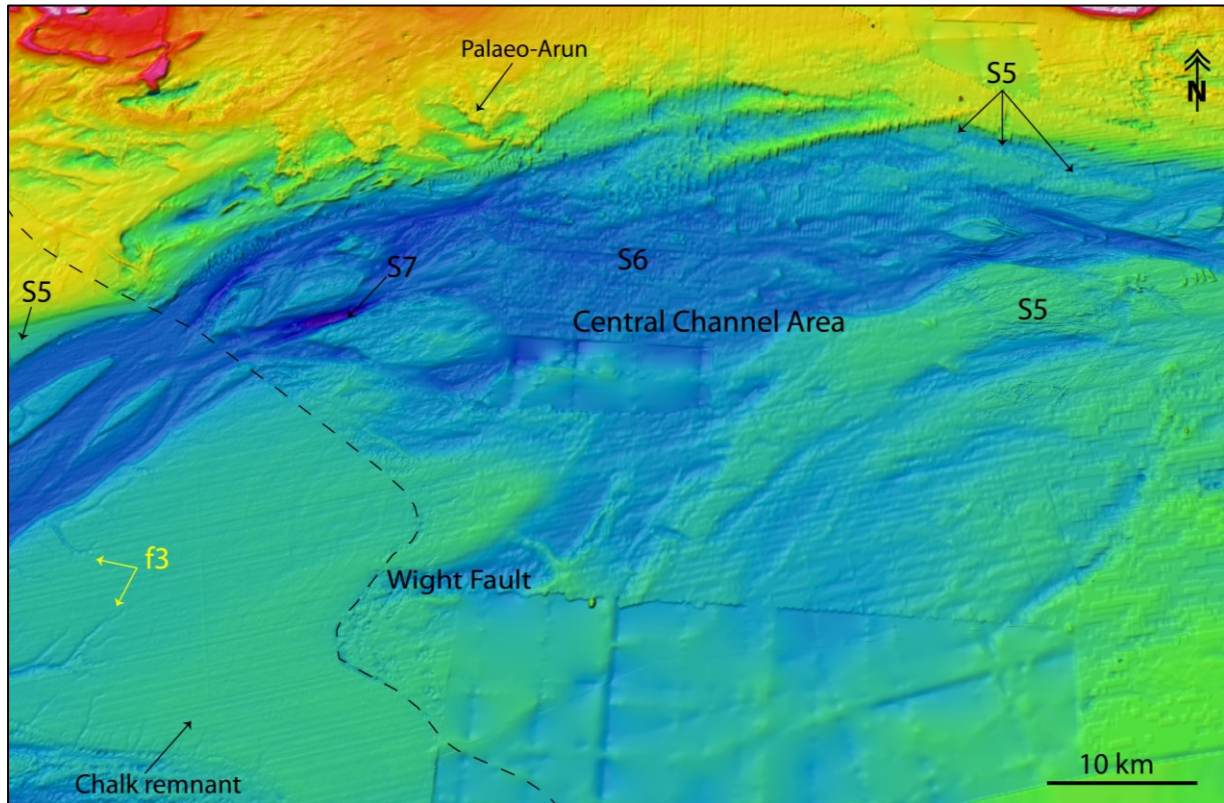


Figure 6.12: Bathymetric image of the Central Channel Area where the Northern Palaeovalley is bounded by surface S5 represented by bedrock remnants in the Hampshire Basin and by a channel bench in the Channel Basin, separated by the Wight Fault. Colour-scale is the same as in Fig. 6.10. See Fig. 6.16 for location of this figure.

6.2.5 Channel Basin Area

Channel geometry

This area corresponds to the south-western portion of the Northern Palaeovalley (Fig. 6.1), located on the west of the Wight Fault (Fig. 6.12), determining a further change in the channel geometry respect to the Channel Basin Area (see Chapter 4 for detailed description of the seabed geology and tectonic structures). A change in the Palaeovalley direction is observed in this area respect to the Hampshire Basin, changing to a NE-SW direction for 40 km and switching again to an E-W direction for ~38 km in the westernmost sector of the Palaeovalley (Fig. 6.13). In this area, the channel geometry narrows down to a maximum width of 6.5 to 12 km against 30 km in the Channel Basin Area, maintaining a depth of 20 to 35 m. This area is generally characterised by narrow and deep channel geometry of the palaeovalley that suggests high-energy erosive flows to have carved the channels, as revealed by the presence of deep inner channels carved into the palaeovalley floor (Baker, 1978b;

Gupta et al., 2007; Baker, 2009a, 2009b; Carling et al., 2009b). The observed inner channels are present in the central and western sector of the Channel Basin area and show a relief of 2 to 25 m. Here, the palaeovalley shows an almost E-W direction and a group of bedrock remnants carved in its floor, separating the bifurcating inner channels (Fig. 6.13) and forming a typical anastomosed and bifurcating geometry typical of flood channels (Baker, 1978a, 1978b; Carling et al., 2009b).

Channel margins – inner bench

The Channel Basin area is characterised by the presence of well-preserved inner benches of the palaeovalley, represented by surface S5. The north-western bench lies at a depth of 48 to 53 m and is 3.5 km wide and 25 km long, extending in a NE-SW direction. The bench displays exactly the same elevation as the cluster of streamlined islands present in the palaeovalley in this area, also showing same acoustic character, suggesting that the bench and the islands represented a single surface (S5) before the erosion of the palaeovalley floor (S6).

A big scale feature underlies S5 on the channel south-eastern margin (Figs. 6.1, 6.12-13). This feature has the same Chalk typical smooth acoustic texture as the inner bench on the other side of the channel, also showing evidence of late drainage channels. The bedrock remnant measures 25 km width and 23 km length. It lies 16 m above the channel floor. This surface is almost flat as the depth only varies between 46 and 48 m across the remnant. The southern portion of the Wight Fault crosses this surface, marking a strong change in the acoustic character of the seafloor in correspondence of the transition between Chalk and Palaeogene rocks. As for the northern bench of the channel, the Chalk remnant also shows the same acoustic character and elevation of the streamlined islands, supporting the hypothesis of the island is a bedrock erosional remnant of surface S5.

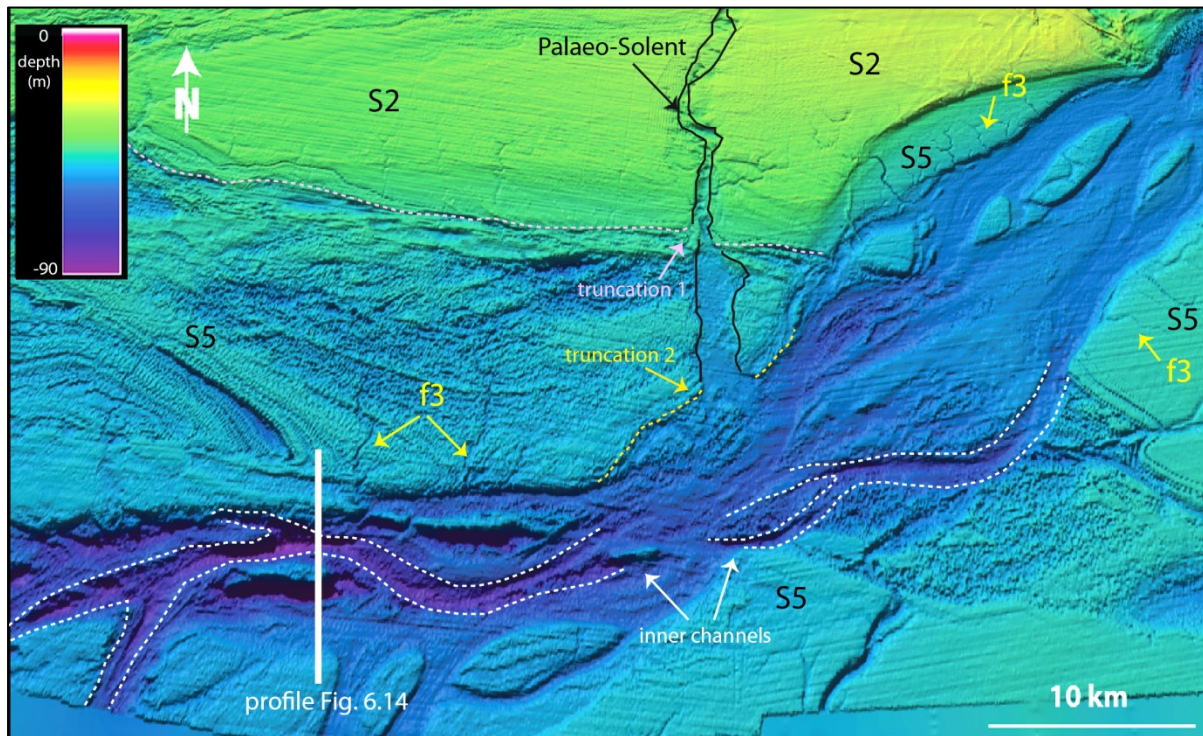


Figure 6.13: Bathymetric image of the Channel Basin area of the palaeovalley, showing up to 25 m deep inner channels incised into the channel floor, separated by elongated bedrock remnants. The figure also shows the double truncation (indicated by pink and yellow arrows and labels) of the Palaeo-Solent River, carved into two erosional surfaces elevated from the Palaeovalley floor. See Fig. 6.16 for location of this figure.

S5 displays, on both margins of the palaeovalley, the typical Chalk smooth acoustic character and is incised by a set of small-scale channels (f3) carved into the Chalk. These represent another generation of sub-aerial, post-flood drainage hanging valleys, indicating that the erosion of these channels into the flood surfaces postdate the time of the flood incisions. Small scale channels labelled as f3 are also present on the surface of the streamlined islands, supporting the hypothesis that the islands represent a remnant of the surface that forms the Paleovalley margin (S5). The small-scale dimensions of these channels indicate that they were carved on the channel benches by normal fluvial processes and not by high magnitude flows, as part of the fluvial drainage system during emerged times. The preservation of these dendritic channels supports the hypothesis that the channels were eroded in sub-aerial conditions. Finally, the comparison between these shallow channels and the kilometre-scale megaflood channels stresses the difference in size between the flood-related features and the typical river drainage features.

Further downstream, where the palaeovalley takes a more westerly direction, its flank is more than 60 km long and 25 km wide, displaying a large-scale fold visible on its bedrock surface (Fig. 6.13, on the west portion of the channel bench), determining a series of changes in lithology and consequently in the resistance to erosion of the surfaces. Another set of streamlined island showing the same elevation and acoustic character of the bench (Figs. 6.13-14) is present on the channel floor, consistently with the hypothesis that the islands were carved by a later flow into a former flood surface, that now represented the channel flank. As shown by the bathymetry, the area that is dominated by the multiple lithological boundaries due to the presence of tectonic structures (see Section 4.4). Drainage channels present on the east are still visible, measuring 0.5 to 2 km of width and 1 to 5 m of depth and flowing for 7 to 25 km with N-S direction.

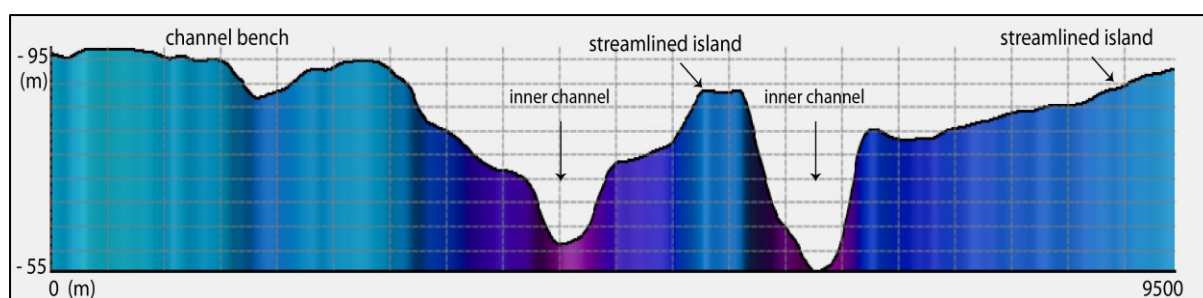


Figure 6.14: Bathymetric profile showing that the channel bench and the two streamlined islands, separated by the up to 25 m deep inner channels (S7), show the same elevation as they were carved by flows into an older flood surface (S5).

Channel margins – outer bench

The outer margin of the Palaeovalley (S2, Fig. 6.1) is only visible on the north, as on the south channel C flows in the French sector of the Channel. The northern margin is represented by a wide bedrock shelf (identified as surface S2), located south of the Isle of Wight area (Fig. 6.1). The surface has width of up to 26 km at its widest point in a N-S direction and extends for more than 90 km at a depth that ranges from 24 m (on the east) to 35 m on the west of the Palaeo-Solent River (Velegrakis et al., 1999; Westaway et al., 2006) that is carved into its central and western sectors. In this area, the bench shows higher slope values with respect to the central and upstream area, with slope up to 0.06° . The surface shows small-scale drainage channels carved in the southern portion and labelled as set f4 (Fig. 6.15). These channels have a relief of 0.5 to 2 m, a width of 0.5 to 1 km and extend for

up to 5 km with N-S direction on the bench. The termination of the channels is truncated in correspondence of a scarp that lies ~10 m above the underlying channel margin, described above as the palaeovalley inner bench.

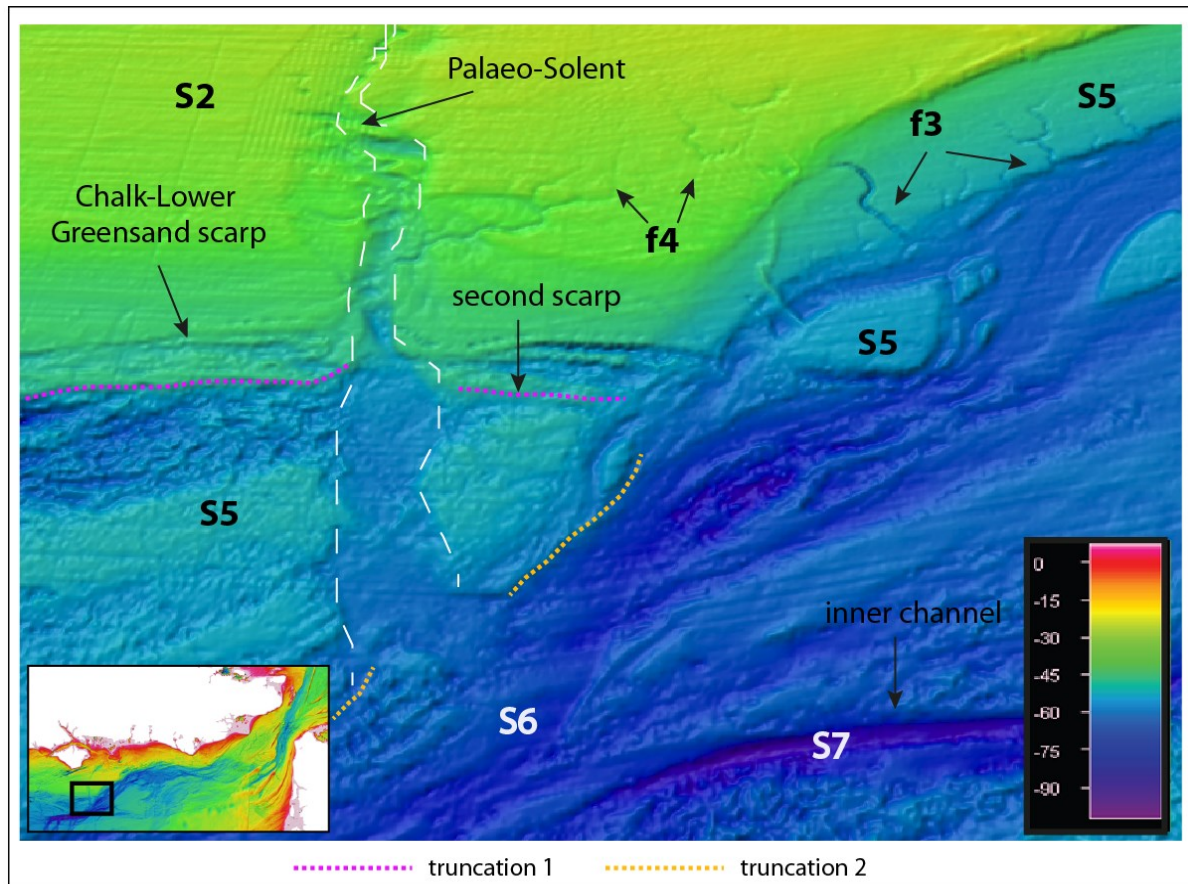


Figure 6.15: Bathymetric image of Palaeo-Solent (dashed white line) flowing through S2 and S5 and being truncated two times by the flow that carved the channel inner bench (S5) and the channel floor (S6). The sub-aerial drainage channels f4 and f3 are also shown in the figure, located at the bench margins. Depth is indicated in m.

6.2.6 Palaeo-Solent River

There is no evidence of flood erosion on the wide bedrock shelf (S2) into which the Palaeo-Solent River is carved. The channel flows through the shelf for 32 km with N-S direction. The Palaeo Solent shows a narrow geometry and is filled by modern seabed sediments that locally almost fill the channel. The Palaeo-Solent modern channel floor is 5 to 16 m below its benches and is sharply truncated at the margin of the surface the shelf (Figs. 6.13 and 6.15), representing another example of hanging valley, as indicated by the figures that display two clear knick-points present along the channel. The first corresponds to a marked scarp present

at the southern margin of the shelf, up to 12 m high, represented by a clear lineation visible on the bathymetry (Fig. 6. 15) that corresponds to the boundary between the area where the Chalk and the Lower Greensand formations outcrop. The presence of the lithological boundary could have created a preferential direction for the erosion of the flood surface (that now represents the channel margin) into the shelf,, truncating the Palaeo-Solent. The second knick-point is associated with the scarp at the southern margin of the channel bench, up to 15 m elevated from the palaeovalley floor, and truncating the Palaeo-Solent River for a second time (Fig. 6.15).

6.2.7 Channel C

This channel is located in the French sector of the merged dataset (Fig. 6.1), which shows very low resolution as the data were interpolated using depth point digitised by available seismic (see Chapter 3). It is therefore not possible to describe the morphology of Channel C with the same level of detail as for the other channels mapped in the dataset. Nevertheless, the eastern portion of the dataset displays a strongly improved data quality in the final merge, showing that Lobourg Channel seems to be connected to Channel C in the Central Channel Area.

Where the Lobourg Channel takes a westerly direction into the Northern Palaeovalley, Channel C and the palaeovalley are separated by a series of bedrock remnants that represent the south-eastern margin of the palaeovalley but also bound the north-western margin of Channel C. The area where the Lobourg Channel flows into the Channel C is not well imaged in the southern sector because of the presence of a thick sediment cover, other than the lower resolution of the data, making detailed mapping of the channel not possible in this area. An almost continuous sediment cover drapes this portion of the seabed, with sand banks up to 25 m high. Channel C seems to follow the same path of the Northern Palaeovalley, showing a sharp change in direction in the Central Area where the channel has an E-W direction for 50 km and successively changes into a south-western direction. This change in the channel orientation is likely to be due to the presence of the big-scale Chalk remnant that separates the Hampshire and the Channel basins. Here channel C diverts, running with a SW direction for 50 km (Fig. 6.1), almost parallel to the Northern Palaeovalley that also change its direction. The channel width ranges from 10 to 20 km with depth of 10 to 15 m, being

shallower than the palaeovalley: this difference in depth could nevertheless be due to the limited resolution of the data that does not allow for a precise measurement.

The limited resolution of available data does not allow for the investigation of flood features in Channel C. Channel C represents the southern branch of pre-flood Channel River, of which the northern branch was the pre-flood Northern Palaeovalley. High resolution data could allow for the identification of possible flood features carved on Channel C surface, as well as for the mapping of multiple erosional surfaces as for the Northern Palaeovalley.

6.2.8 Summary of channel network morphology and comparison to pre and post-flood morphology

In order to fully understand the effect of the flood flows on the topography of the English Channel and therefore be able to quantify the magnitude of these events, it is necessary to distinguish flood-related features from pre-flood morphology.

Pre flood topography can be mapped on the dataset from the Dover Strait to the downstream reaches, represented by the English Channel River System, described in Chapter 2. The English Channel River System is composed of two main branches: the Northern Paleovalley and the Southern branch, which is the offshore continuation of Seine and Somme Rivers. Both are still present on the seabed of the English Channel, modified by the floods and with erosional features superimposed on the pre-flood features. A number of fosses (e.g. Hurd Deep, St Catherine's Deep), are also present in the downstream area of the Channel River System, representing its deepest point. These are also interpreted as pre-flood topography as there is no evidence of their catastrophic erosion (see Section 2.3, Lericolais et al., 2003).

The Northern Paleovalley, previously representing the main north-west European river drainage system, would have been carved by at least two flood flows, changing its morphology and dimensions and transforming Arun and Paleo Solent rivers into hanging valleys. The Lobough Channel would have only existed after the second breaching of the Dover Strait. Erosional remnants formed as a consequence of the high energy flows that transformed the Northern branch of the Channel River into a flood channel.

Table 6.1 summarizes the geomorphologic characteristics of the channels that compose the Northern Palaeovalley network, indicating their dimensions, geology of the seabed and their interpretation in relation to the megaflood events. The analysis of the channel geometry and of the related erosional surfaces shows how the channel network displays all the typical characteristics of flood-generated channels and erosional surfaces.

The Lobourg Channel and the Northern Palaeovalley are a network of kilometre-scale anastomosed channels carved into bedrock, suggesting exceptionally high-energy erosive flows to have carved the channels into the Cretaceous and Palaeogene rocks for hundreds of kilometres. The associated erosional surfaces are associated with the modern, sediment-free channel floor and by the channel margins and inner benches. Maximum erosion depth can be estimated for the bedrock platform carved by drainage Channel B in sub-aerial conditions after the flood. Smaller-scale drainage channels also post-date flood erosion, suggesting that flood surfaces on which they are carved can be used for palaeo-hydraulic estimations for the two flood events.

Channel name (erosional surface)		Carved into surface	Length (km) (flow direction)	Depth (m)	Width (km)	typical width:depth	Lithology of the seabed (sediment cover if present)	interpretation
Lobourg Channel	Upper LC (S6)	S4 (LC inner bench)	85 (NE-SW)	10-30	2.5-8	250	Middle Chalk to Wealden	flood channel
	Lower LC (S6)	S4 (LC inner bench)	30 (ENE-WSW)	10-25	10	500	Wealden and Kimmeridge (sand banks 0 – 35 m)	
Channel A		S3 (LC outer bench)	25(NE-SW)	4-10	0.8-1.5	150	Middle Chalk(sand waves 1 m)	tributary channel
Channel B		S3 (LC outer bench)	38(NE-SW)	5-8	0.5-1	100	Wealden (Sand waves 5 m)	drainage channel
Central Channel Area (S6)		S5 (Palaeovalley margin)	90 (E-W)	25-30	20-25	1000	Wealden and Palaeogene (Sand waves and banks 0-25 m)	highly eroded flood area
Palaeo-Solent River		S2 (pre-flood bedrock shelf), S5 (Palaeovalley margin)	32(N-S)	5-16	0.8-2	150	Middle Chalk to Wealden	hanging valley
Downstream Palaeovalley (S6)		S5 (Palaeovalley margin)	40 (NE-SW)/40 (E-W)	20-35	6.5-12	350	Upper Chalk to Wealden	downstream flood channel
Inner channels		S6 (Palaeovalley floor)	30 (E-W)	2-25	1-2	100	Upper Chalk	inner flood channels
Post drainage channels	f1	S1 (pre-flood platform, LC outer margin)	45 (NW-SE)	0.5-1	0.5-2	500	Upper Chalk	post-flood drainage channels
	f2	S3 (LC outer margin)	8 (NE-SW/NW-SE)	3-5	0.5-1	200	Middle Chalk	
	f3	S5 (Palaeovalley margin)	5 (N-S)	0.5-1	0.5-2	500	Middle Chalk	
	f4	S2 (pre-flood shelf)	25 (N-S/E-W)	1-5	0.5-2	500	Palaeogene and Middle Chalk	

Table 6.1: Summary of the geometrical parameters of the mapped channels and interpretation of the mechanisms responsible for their formation.

Post-flood drainage Channel B is sharply cut by the Lobourg Channel, entering the Palaeovalley from the East and truncating the channel that lies on the platform, 5 m elevated from the channel floor, representing a typical example of hanging valley (Baker, 1978a). The Palaeo-Solent River also is a hanging valley, carved into two surfaces (S2 and S5) that are respectively the palaeovalley inner and outer northern benches. These surfaces are elevated from the channel floor (surface S6) of respectively 30 and 15 m. Finally, the interpreted erosional surfaces are incised by four distinct sets of small-scale channels representing post-flood sub-aerial drainage channels (Gupta et al., 2007). The comparison between the scale of the flood and the fluvial-related drainage channels is a further indication of the exceptional dimensions of the palaeovalley channel network, carved for hundreds of kilometres into bedrock.

As the channels are entirely carved into bedrock, the relation between their geometry and the seabed geology defines the flood terrain. Where the seabed is dominated by more competent Cretaceous rocks (Lobourg Channel and Central area, respectively in the Wealden and Channel Basins) the channels show rectilinear geometry, well-preserved benches and bifurcating geometry with inner channels separated by elongated bedrock remnants. Where the palaeovalley is carved into more erodible Palaeogene rocks (Hampshire Basin), the channel changes its direction, flowing E-W, and showing a much broader geometry with width up to 25 km. The inner channels and bedrock remnants present in this area show lower relief, while the channel benches are highly eroded and only present as bedrock remnants, other than well-preserved erosional surfaces.

Sections 6.3 and 6.4 detail the morphometric analyses performed on the sets of streamlined islands and longitudinal lineations, located in the upstream area of the dataset on the Lobourg Channel floor. Through detailed analyses and analogy to similar features interpreted on Earth, the geometry, nature and formation mechanisms of these features will be analysed with the aim of testing a flood-generated origin of these features.

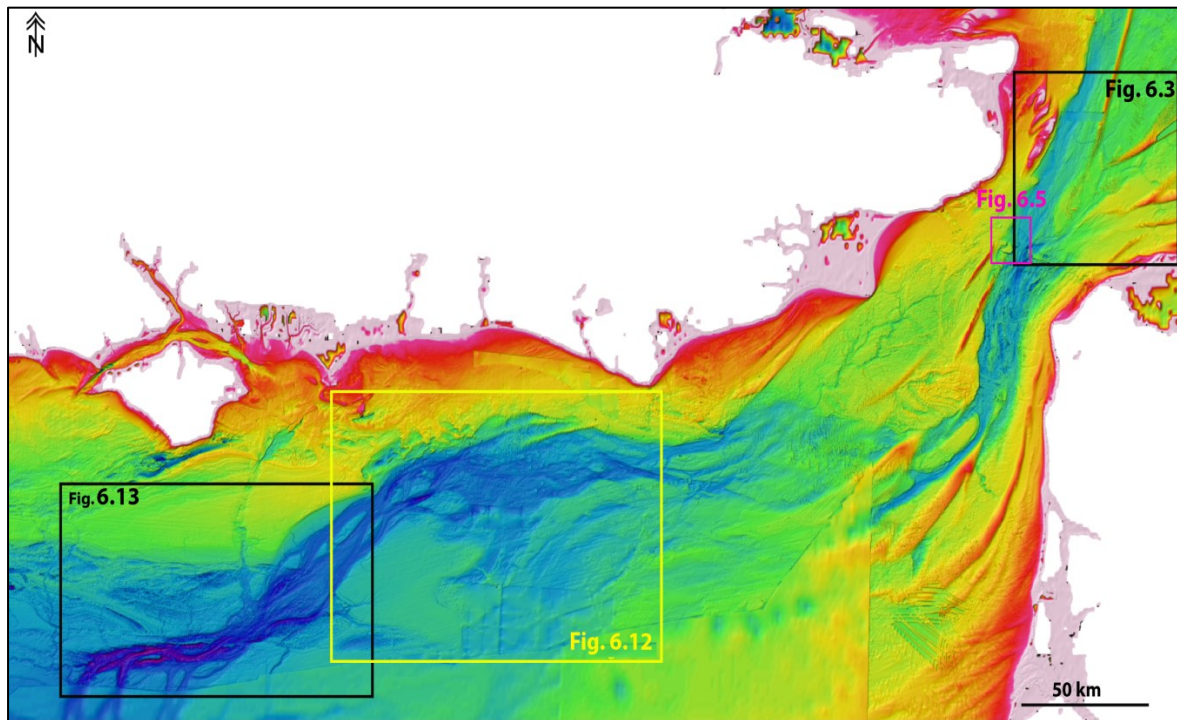


Figure 6.16: Location of figures shown in this section.

6.3 Streamlined islands of the Northern Palaeovalley

Following the interpretation of three distinct sets of streamlined islands present in the Northern Palaeovalley, this section details the geomorphologic and morphometric analyses performed on the mapped features using bathymetry and seismic data.

6.3.1 Background

Streamlined geometry

Streamlined islands are erosional remnants, formed by high magnitude flows and characterized by typical teardrop geometry that develops as the islands are carved into erodible material by a process of minimizing fluid drag when subjected to flow during their formation (Baker, 1978b).

Bretz (1923) first recognized streamlined forms in the Cheney-Palouse area of the Channeled Scabland terrain, identifying teardrop-shaped isolated hills of loess. Streamlined islands are often present in clusters, identified as component of braided channel systems, where they are presents in groups of hundreds of isolated elements (e.g. Trevena and Picard, 1978; Baker, 1978a; Komar, 1983; Leverington, 2004) as for the case of the Channeled Scablands. As they are observed in several megaflood terrains (e.g. Bretz, 1923; Baker and

Nummedal, 1978; Kehew and Lord, 1986; Gupta et al., 2007; Meinsen et al., 2011; McClenagan, 2013) and represent one of the most distinctive flood-related features carved by megaflood floods, several analyses have been conducted in order to understand the mechanisms responsible of their formation. In particular, due to their characteristic teardrop-shaped geometry, streamlined islands have been compared to the shape of lemniscate loop (Baker, 1978a; Komar, 1983, 1984), similar to airfoil geometry (Fig. 6.17) and defined in polar coordinates by the formula:

$$r = l \cos(k \theta) \quad \text{with } k = \pi^2 l^2 / 4A$$

where l is length, A is area and k is the elongation factor. In order to compare the geometry of streamlined islands to the lemniscate loops, geometry analyses were performed on islands and airfoil-shaped laboratory features (Komar, 1984; Baker and Kochel, 1978), finding a direct relationship between length-to-width (l/w) ratio and k values of both lemniscate loop (Fig. 6.17) and streamlined island geometry (Komar, 1984).

In order to fully understand the mechanisms and hydraulic conditions that lead to the formation of streamlined islands, Komar (1983) also performed a series of flume laboratory experiments on small-scale islands comparing the geometry of lemniscate loop, symmetrical airfoils, drumlins, and streamlined islands of different shapes formed by water flows.

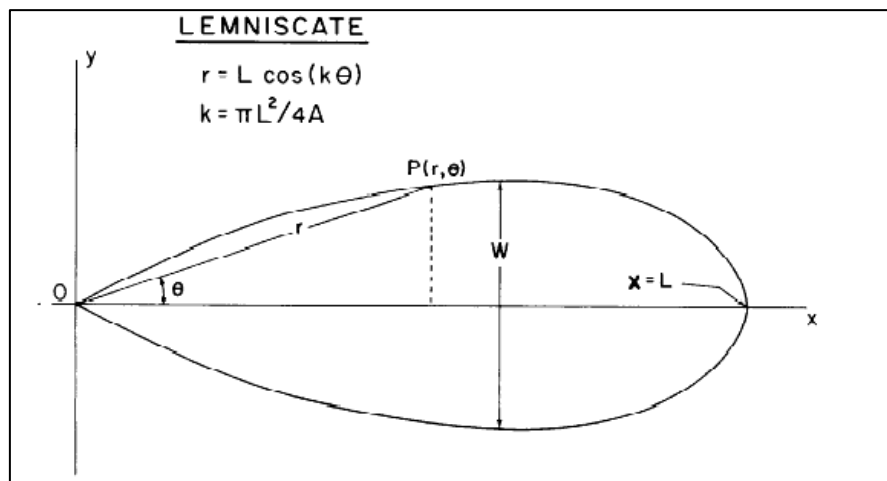


Figure 6.17: Properties of a typical lemniscate loop, showing its polar (r, θ) and Cartesian (x, y) coordinates (from Komar, 1984).

Komar's analyses focused primarily on the length-to-width ratios of these landscape features and their consistency with the analytical lemniscate shape. Earth and Mars streamlined islands resulted to have nearly the same average length-to-width ratios, as well as other dimensions, and are consistent with the lemniscate and airfoil shapes (Baker and Kochel, 1978; Komar, 1984).

The explanation for the observed l/w ratios of streamlined geometry can be explained as the result of minimisation of the island drag when subjected to a flow. The total drag of an object with respect to a fluid is given by the sum of pressure (or form) drag and frictional (or skin) drag (Shapiro, 1961). Form drag is dependent on the form of the object while skin drag depends on its surface. Komar's experiments, together with Baker's (1978a, 1978b) observations on Scablands islands, demonstrated that the minimum total drag is achieved and maintained by the islands when they reach a length/width ratio of 3 to 4, which is considered the equilibrium form that represents the best compromise between skin and form drag. In particular, when the equilibrium form is reached by the island, some three-fourth of the island original area has been eroded away (Komar, 1983).

Erosion, which is at first very fast, rapidly decreases its rate and the island shape is maintained. This is the reason why islands that reach equilibrium geometry are more likely to be preserved.

With regard to the conditions under which these islands are formed, Komar (1983) demonstrated that erosional streamlining is most effective in carving lemniscate-shaped islands when the islands are submerged by flow. In particular, highest efficiency is reached when water depth is sufficient to top the island surface, with flow over the island becoming supercritical because of decreased flow depths and high velocities (Komar, 1983). Patton and Baker (1978a) also recognized that the Scablands islands formed under flow conditions displayed better shaping, find a very strong similarity between the observed islands and those used by Komar in his submerged and short-flow condition experiments (1983).

Application of streamlined geometry in geomorphology

Streamlined islands are used in geomorphology for two different aims. Firstly, the islands give evidence of erosional processes that involve presence of water and can be used as palaeo-flow indicator as the flow that carves these elongated features has a parallel direction

to their main axis (e.g. Baker and Kochel, 1978; Baker and Nummedal, 1978; Komar, 1983; 1984). In addition, streamlined islands represent reliable flood-water level indicators that are used for palaeo-hydraulic analyses. Islands can be in fact be used to constrain the high-water levels during flood, as their surface is often cut by features such as divide crossings, indicating high water surface elevation (Baker, 1978b). Divide crossings consist of troughs carved into the islands by flood water that spilled out of the water- filled channels and are interpreted to establish lower limits for high water surface (Bretz, 1923; Baker, 1978a, 1978b; Patton and Baker, 1978b). The error in the measurement of the water level is inversely proportional to divide depth as estimates are more accurate for those divides that were carved by shallow water (Baker, 1978a).

In the case of the English Channel population the analyses on the mapped islands was performed in order to answer the following questions:

- What is the material in which the islands are carved and what is the relationship between the seabed geology and their geometry?
- Is the island morphology consistent with the typical airfoil geometry observed for islands in laboratory experiments and other megaflood terrains?
- What was the mechanism responsible for their formation? Is it consistent with the megaflood hypothesis?
- How can these features be used in the reconstruction of palaeo-flow path and hydraulics of the flood?

6.3.2 Morphology of the mapped streamlined islands

A population of 36 islands (Fig. 6.18) was mapped using the merged bathymetry dataset: this enabled detailed morphometric analyses of the geometry and distribution of the islands, allowing for a comparison with populations of islands mapped in other megaflood terrains. The streamlined island population interpreted in the study area shows a variable range of dimensions and geometries. The island's lengths range from 1.5 km to 20 km and the width varies from 350 to 6800 m. The surface area of the islands is also variable, ranging from 6 up to 83 km² while their elevation above the seabed ranges from 4 to 30 m. Despite the variability in the size and geometry of the islands, there is no definite pattern in their

dimensions as bigger and smaller islands are distributed both in the upper and downstream portions of the dataset. Three main sets of islands were identified, depending on their location and on the lithology of the bedrock in which they were carved: from upstream to downstream, they were divided in sets A, B and C. In order to understand the relationship between their location, the bedrock geology, and their geometry, the three sets of islands are described and compared in this section.

Set A

Set A is composed of 13 islands, located in the Wealden Basin and displaying an elongated geometry (higher l/w values respect to sets B and C). The islands display flanks with gentle slope (1 to 2°) and amplitude at the seafloor of 4 to 30 m. Their amplitude is generally less than 10 m, with the exception of the biggest island present in the dataset, located in the Lobourg Channel (Fig. 6.19a). This island is the northernmost feature present in the dataset and measures 20 km in length, 5 km in width and 30 m in amplitude above the seafloor, showing a bench (that lies 5 m below the top of the island) on its north-eastern margin. The bench is only visible in the northern portion of the island as it shows more definite geometry and lower level of erosion respect to its southern portion, as illustrated in Section 6.2 of this chapter. The islands displays a 5 m deep cross-over channel (Fig. 6.19a) that cuts the surface with an orientation of 40 to 60° respect to the island main axis. This channel represents a typical feature observed on the surface of streamlined islands, interpreted in literature as the result of oblique wave on the island surface (Baker, 1979; Komar, 1983).

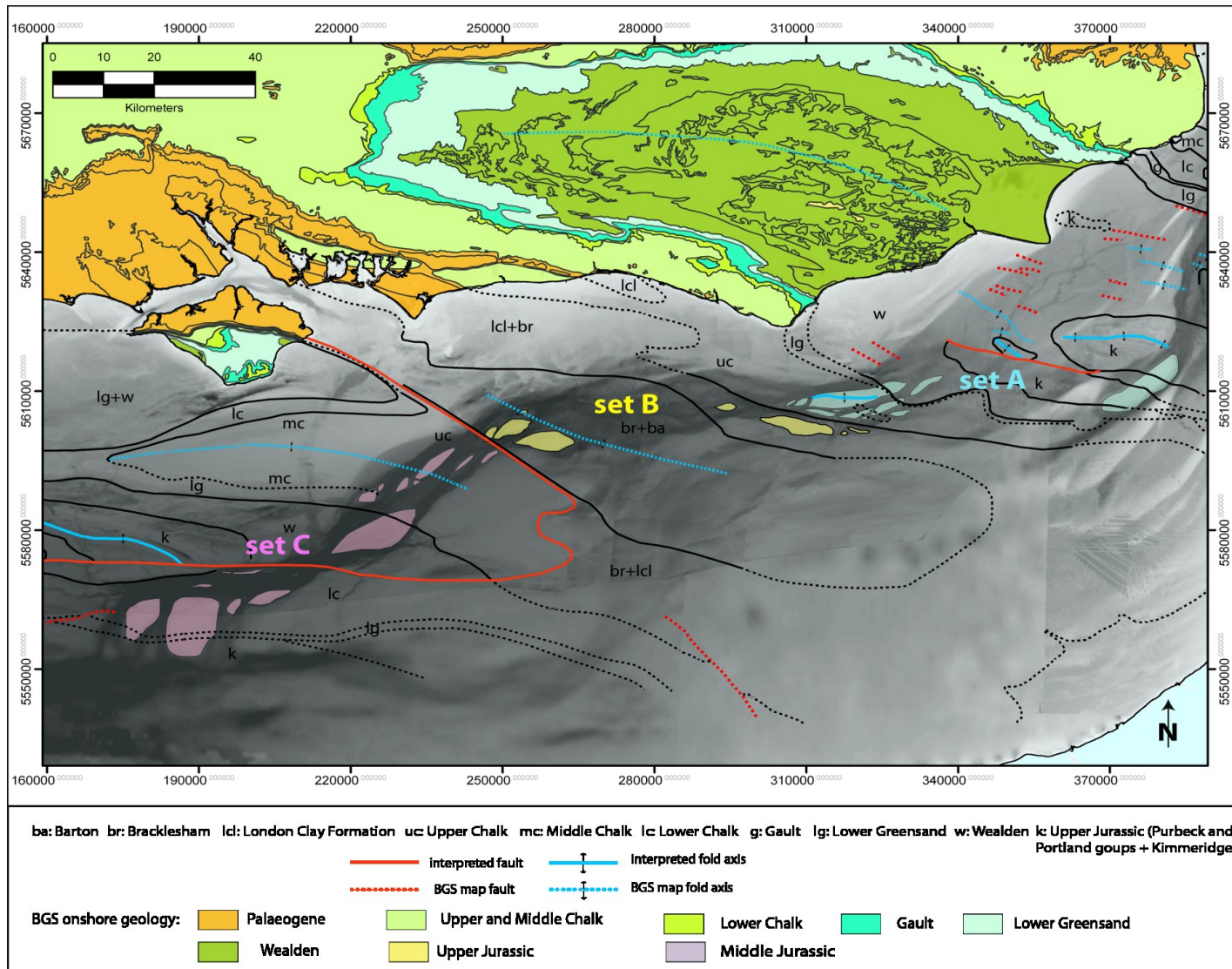


Figure 6.18: Map of the interpreted streamlined islands, subdivided in three groups: set A (light blue), set B (yellow) and set C (pink). The grey-scale bathymetry is displayed in the background together with the geological map of the area (See Chapter 4 for details).

Further downstream in set A (Figs. 6.18-19b) a cluster of islands (shown in Fig. 6.19b) also shows differential erosional on its surface, displaying two, 2 m-deep incisions running parallel to the main axis of the biggest of the islands, almost 20 km long.

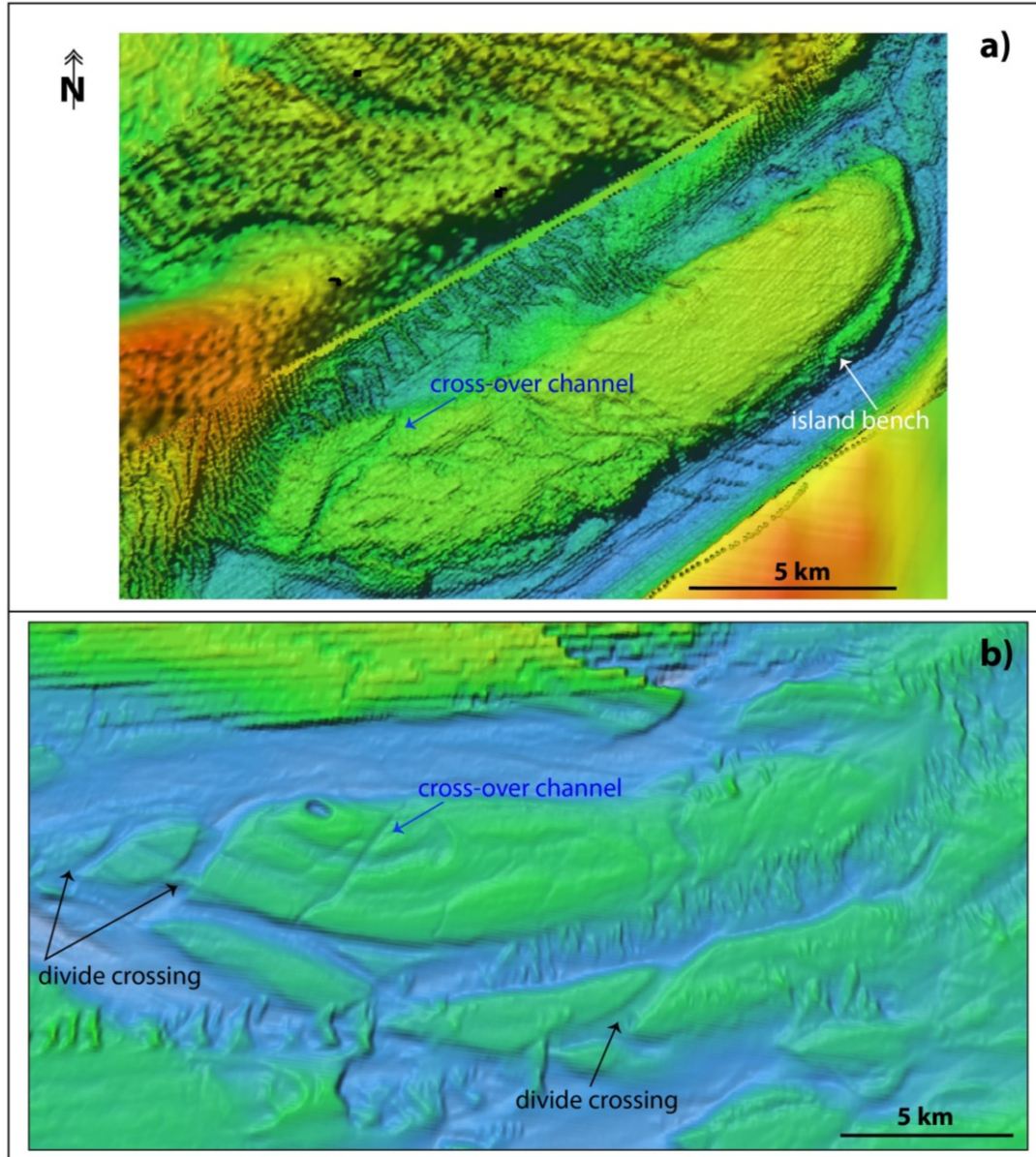


Figure 6.19: Image of the islands of set A. Box a shows the biggest island of the set carved into the Upper Chalk formation, showing cross over channel cutting the top of the island and a series of parallel divide crossings. Colour-scale is the same as in Fig. 6.20.

The island also shows the presence of a cross-over channel cutting its surface with depth of 1 to 2 m and orientation of 60 to 70° respect to the main axis of the island. The smaller islands located on the west (Fig. 6.19b) appear to have previously been part of single larger island that has successively been eroded into smaller islands, divided by divide crossings (Bretz,

1923; Baker, 1978b; Komar, 1983). These channels are originated by flood water overspill from the main channel: in contrast to dendritic drainage channels described on the flood channel benches, these channels are usually linear or nearly linear (Baker, 1978b). Divide crossings represent water marks as they are carved by water overspilling from the main flood channels (Patton and Baker, 1978b). Despite having different dimensions, the smaller and bigger islands, from which the smaller features are derived, seem to have the same shape and l/w value. This is a further evidence of l/w ratio of the islands being scale-independent. In addition, the observed cluster of islands, separated by divide crossings and showing cross-over channels on their surfaces, display strong similarity in geometry to sets groups of islands present in the Cheney-Palouse tract of the Scablands (Bretz, 1923; Baker, 1978b). Importantly, as shown by Fig. 6.19b, the divide crossings that cut the big island into smaller features show parallel orientation to each other and to the cross-over channel on top of the island surface, as for the case of the Cheney-Palouse islands (Baker, 1978b) indicating that the palaeo-direction of the flows that carved these features remained constant.

Set B

Set B is composed of 7 islands, showing the typical airfoil geometry and a less elongated geometry (lower l/w values) respect to the islands of set A. The amplitude of the remnant ranges from 6 to 15 m from the seabed and the slope varies from 1° to 2°, as for set A. As these islands are carved into the softer Palaeogene rocks outcropping in the Hampshire basin, this set of islands has undergone greater erosion with respect to the sets carved into more competent rocks, as shown by the severe erosion undergone by the islands of set B. In addition, the number of islands present in this set is much lower respect to sets A and C. This difference in the number of observed islands could be explained by the higher erodibility of the Palaeogene rocks, causing the preservation of a smaller number of features respect to those carved into Cretaceous rocks. In particular, all the islands observed in this set display almost perfect teardrop geometry and less variability in their morphology. Conversely, sets A and C display more elongated geometry of the islands rather than teardrop-shape. This suggests that only the islands that reached a perfect equilibrium form, corresponding to the teardrop shaped geometry and to l/w values of 3 to 4, survived the erosion by later events into the soft Palaeogene rocks. Despite the high erodibility, the islands that reached their equilibrium geometry were preserved as the erosion rate decreases considerably (Komar, 1983) as resistance to flow is minimized.

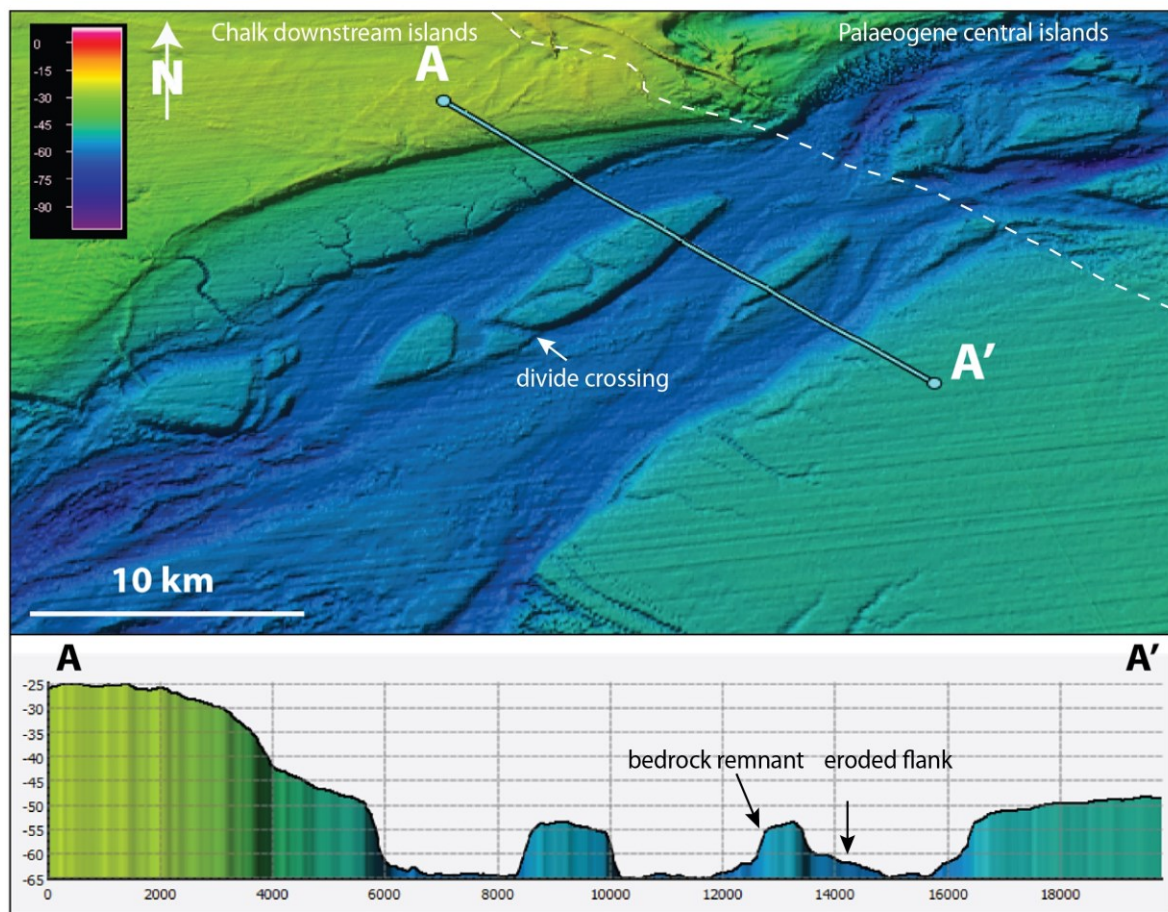


Figure 6.20: Bathymetric profile across two streamlined islands showing the two bedrock remnants with 10 m amplitude at the channel floor. The western islands also show eroded flanks gently degrading to the channel floor. Note that the elevation of the islands is the same as the channel inner bench, indicating that the islands were carved into the surface representing the channel bench (S5). The basin boundary (Wight Fault represented by the dashed line) marks the change in geometry and acoustic texture between Chalk and Palaeogene islands.

Set C

Set C is composed of 16 islands with amplitude of 6 to 18 m above the seabed. These features are narrower and more elongated in shape compared to the features described in sets A and B. Particularly, the slope of their flanks is much steeper, with values up to 11° for the narrowest island: this is likely to be due to the different morphology of the flood channel in this area of the palaeovalley, displaying 25 m deep inner channels and narrow geometry. The variation in the geometry of the channel, controlled by lithology, is reflected on the morphology of the streamlined islands displaying well-defined geometry and acting as flow obstacle, separating the palaeovalley inner channels. Benches are visible on the flanks of the islands: one of the islands shows strongly eroded flanks with a narrow and more elevated central part that

represents what is left of the former island eroded by the high energy flows (Fig. 6.20). Both eroded flanks of the remnant have been symmetrically eroded and form a bench degrading at a very gentle slope (0.1° to 0.7°). The profile in Fig. 6.20 also shows that the islands and the channel bench, interpreted as surface S5 in Section 6.2 both display the same elevation and presence of post-flood sub-aerial erosion drainage channels f3.

The majority of the islands of set B, and in particular the ones displayed in Fig. 6.20, are carved into Upper and Middle Chalk formations, as indicated by their sharp and defined geometry due to the lithological characteristic of the Chalk formations. As the rocks are more competent, islands are well-preserved, showing a wider range of morphologies, from teardrop shaped to more elongated islands (Fig. 6.20). This is consistent with the higher number of islands observed in this set respect to sets A and B: more islands were preserved from erosion of later flows, including those islands that did not reach the perfect equilibrium geometry. This also indicates that, given the high resistance to erosion of the bedrock rocks in this area, the observed streamlined islands and inner channels present at the seabed were carved by high-energy flows capable of eroding such features into Chalk. Streamlined remnants are known to be carved by megaflood flows into a variety of materials such as sediments (Meinsen et al., 2011), loess and basalt (Baker, 1973a) that require different levels of flow energy for their formation, depending on the level of resistance to erosion of the material into which they are carved. The presence of streamlined islands in the Channel Basin area is therefore consistent with the hypothesis of extremely high energy flows to have carved the islands.

6.3.3 Seismic acoustic character of the islands

Observations

The available Central Channel seismic lines, crossing the Central Channel area, were used for the interpretation of the internal seismic facies and structure of the islands, in order to investigate whether the observed features are composed of bedrock or quaternary seabed sediments. As the islands are present in areas with different seabed lithology, as shown in Fig. 6.18, it is possible to observe from seismic data how bedrock geology plays an important role in the definition of the island acoustic facies and geometry. Fig. 6.21 shows three examples of seismic images of islands carved into Palaeogene Bracklesham, Wealden and Upper Chalk formations. A strong change in the acoustic facies of the islands is visible: while

the latter displays a smooth surface and transparent acoustic facies due to the absence of stratification in the Upper Chalk Formation, the islands carved into the Wealden and Palaeogene (Bracklesham and Barton formations) rocks display parallel reflections given by the highly stratified character of these formations. The contrast in the acoustic facies of the islands is consistent with the different geometry observed from bathymetry in the islands carved into different lithologies. This evidence also confirms that the islands carved into the Chalk Group display a higher resistance to erosion than those carved into the softer Palaeogene islands. Fig. 6.21 shows how the surface of the Palaeogene island displays an irregular geometry with evidence of multiple incisions (Fig. 6.21) indicated by the presence of sediments in the upper part of the islands, deposited into small scale incisions (up to 8 mm deep) carved into the bedrock surface.

Interpretation

The streamlined islands present on the English Channel seabed features have been previously interpreted by other authors as sand bars (James et al., 2007): it is nevertheless clear from the seismic data that the islands are composed of bedrock. This is indicated by the acoustic character of the islands observed from seismic data (Fig. 6.21): the islands display different acoustic facies in relation to the variation in lithology of the rocks that outcrop at the seabed. In order to support this hypothesis, the distribution of Quaternary seabed and channel-fill sediments in the area was mapped from the seismic data (Fig. 6.22) and compared to the location of the streamlined islands. This enables to identify the areas of the seabed dominated by channel-fill sediments or surficial sediments and distinguish them from the areas where bedrock outcrops at the seabed.

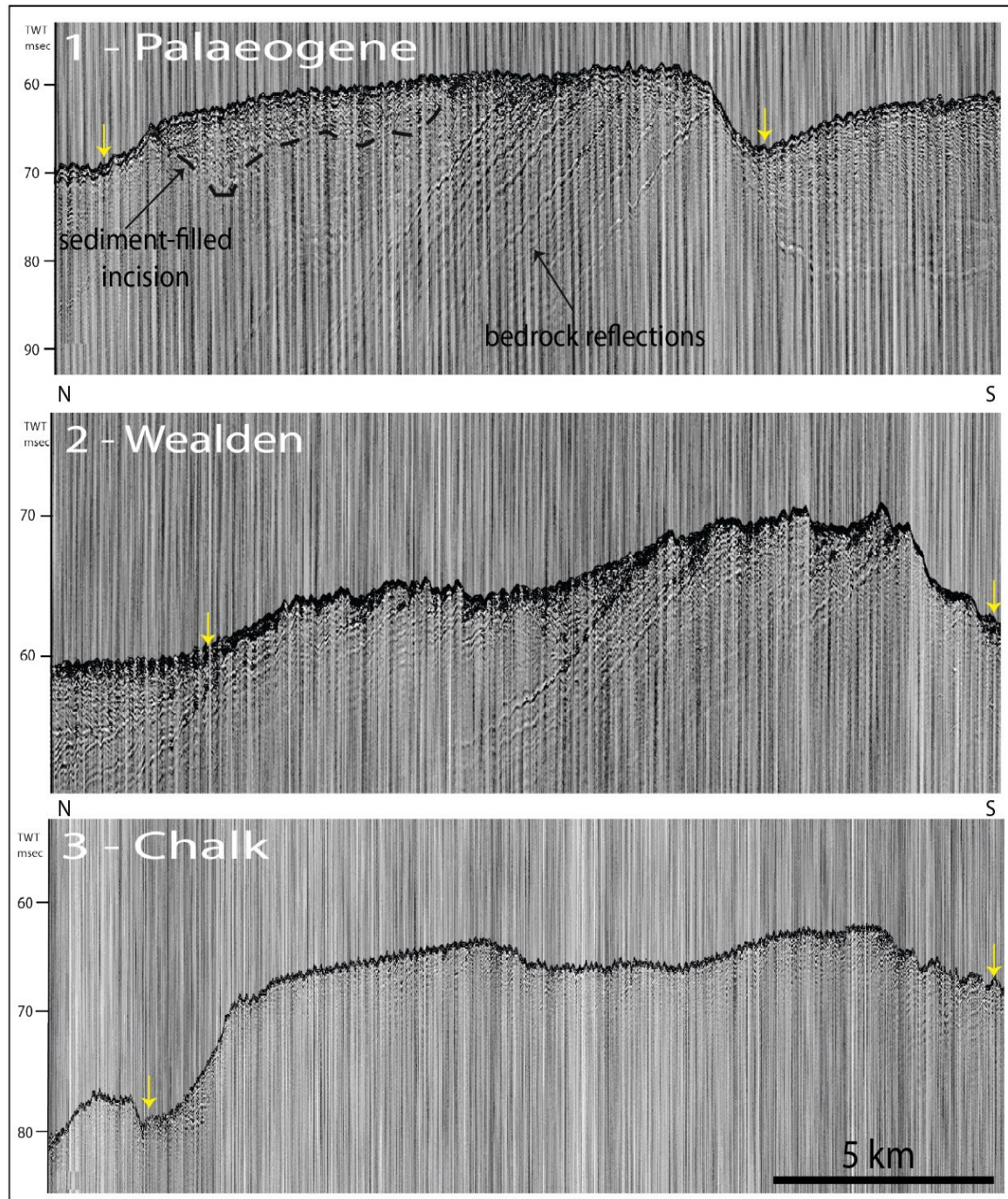


Figure 6.21: Seismic lines showing the seismic character of the islands changing with bedrock lithology. A strong contrast is visible between the competent, smooth surface of the island carved into the Upper Chalk Formation and the eroded surface of the Palaeogene island that shows sediment filled-incisions carved into its surface. Yellow arrows indicate the edges of the islands. Line location in Fig. 6.22.

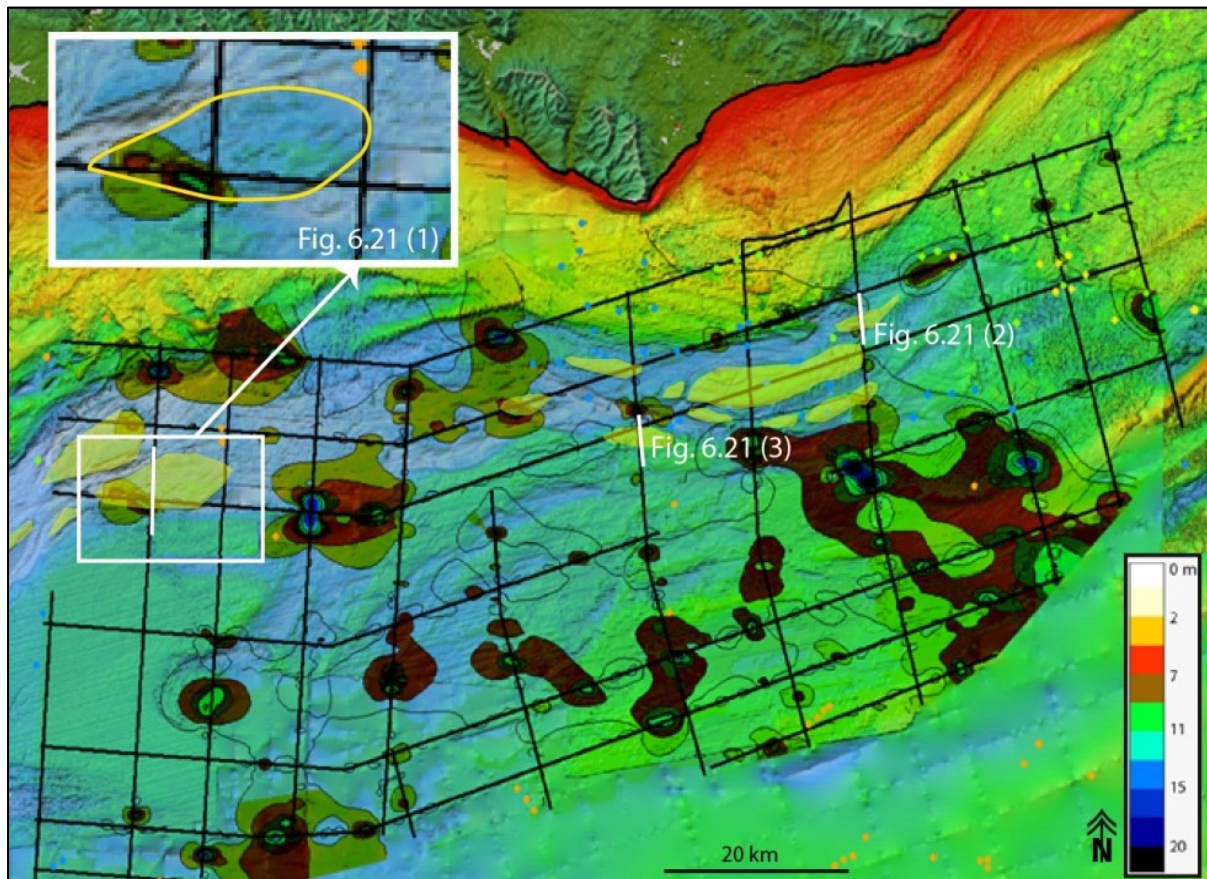


Figure 6.22: Isopach map of the seabed sediments filling and masking the seafloor that is prevalently sediment-free in this area. The bathymetry is displayed in the background, showing how the streamlined islands (yellow areas) are entirely made of bedrock with the exception of small areas where sediments fill the eroded incisions in the island surface (see Fig. 6.21). The coloured dots in the figure represent the available cores (for details see Chapter 4).

Fig. 6.22 shows that the portion of the seabed where the islands are located is dominated by bedrock and no significant accumulation of modern sediments is present, with the only exception of some island showing small scale incisions carved into their surface and filled by the mapped seabed sediments.

6.3.4 Morphometric analyses

The aim of the morphometric analyses is the quantitative characterization of the islands, with the aim of correlating the changes in the geometry in relation to the island location, geology and dimensions. Secondly, the quantitative characterization of the islands allowed for a comparison of their geometry to other islands mapped in Earth megaflood terrains.

To do so, the following geomorphic parameters were calculated:

- length/width ratio (**l/w**): gives indication of the elongation of the island.
- **k**: used for the quantification of the roundness of the islands, with $k=1$ representing a circular shape. k is calculated using the formula:

$$k = l^2 * \pi / 4A$$

where **l** is length and **A** is the area. This parameter was introduced by Baker and Kochel (1978) for the analyses of streamlined islands.

- Maximum width factor (**ASpl**): first used for the analysis of drumlin planar shape (Spagnolo et al. 2010), gives indication of the grade of symmetry of the island geometry (Fig. 6.23).

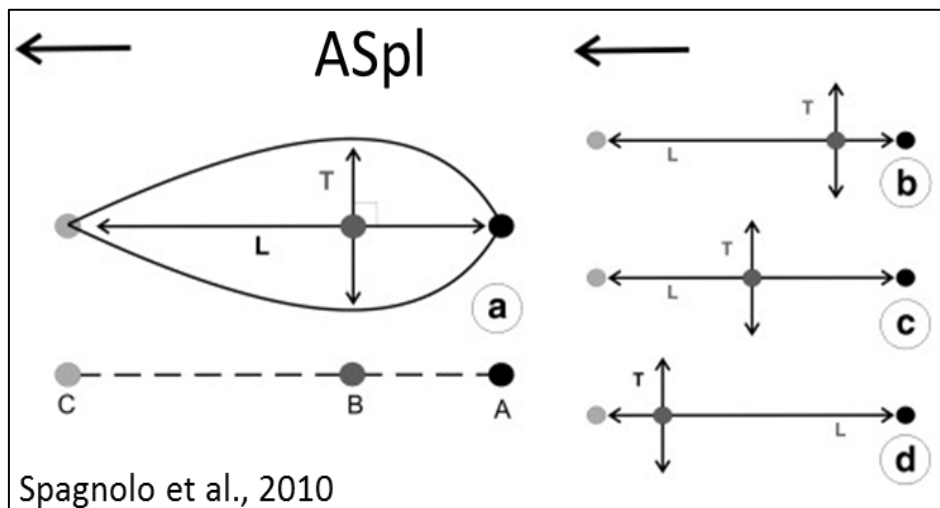


Figure 6.23: ASpl value is calculated by AB/AC . The values for typical shape, considering the direction of the flow (indicated by the arrows), would be 0.2 for shape b, 0.5 for c which corresponds to a symmetrical shape and 0.8 for d that corresponds to a reversed feature.

Table 6.2 displays the results of the geomorphic calculations performed on the island population, displaying the calculated value for the entire population and for each set. The variation of values between the three sets give further indication on the influence of the geology on the morphology on the islands and allows for a quantitative interpretation of the observations already made on bathymetry and seismic data.

Reference	Set of islands	l/w	k	AS_{pl}	A (Km ²)
This study	Entire population	4.04	4.69	0.44	11.04
	Set A	4.45	5.2	0.48	10.06
	Set B	2.58	3.26	0.37	12.93
	Set C	3.74	4.32	0.40	11.75
Literature	Channeled Scablands (Baker and Kochel, 1978)	1.3-5.8 avg. 3.15	2.9-4.3	0.3-0.4	0.10-100
	Airfoil geometry (Komar, 1983, 1984)	3-4	3-4	0.30-0.35	---
	Drumlins, (Spagnolo et al., 2010)	---	---	0.46	---
	Glacial Lake spillway (Kehew and Lord, 1986)	2.8-55 avg. 3.4	---	---	0.6
	NW Germany (Meinsen et al., 2011)	1.3-8.1 avg. 3.3	---	---	---
	British Columbia (McClenagan, 2013)	2.4	---	---	23.8

Table 6.2: Summary of the calculated geomorphic parameters for the entire population and for the single sets, compared to the typical values of megaflood streamlined islands, drumlins and airfoil geometry. Values represent either a range or average value (where one single value is indicated).

The statistical analyses made on the entire population (see Appendix C for plots) show a very good correlation between l/w and k , as both values are related to the type of flow that carved the islands: Baker (1978b) suggested, in its analyses on the Cheney-Palouse Scabland tract islands, that k values tend to increase with higher Reynolds number (most of Scabland tracts had Reynolds number value between 2×10^8 and 2×10^9).

A comparison between the values of the three different sets indicates that values of set B are remarkably different from the sets A and C, as indicated by Table 6.2 and Fig. 6.24. While the more elongated upstream and downstream islands display a mean l/w value of respectively 4.5 and 3.7, the mean value of the central set is 2.58. The same difference is observed in the k values, similar for the upstream and downstream sets and much lower for the central set. In general, lower k values correspond to a more rounded and teardrop-shaped geometry while higher values indicate more elongated shape. This is consistent with the more rounded shape of set B respect to the more elongated shape of the sets A and C. As shown by

bathymetric data, islands of set B underwent stronger erosion as they are carved into softer rocks. As a consequence, a smaller number of islands are observed in set B and, importantly, only the islands that reached an equilibrium geometry (indicated by a teardrop shape and l/w value of 3 to 4) were preserved. The k values of the islands of set B support this hypothesis: islands from flume experiments (Komar, 1983) show k values of 3 to 4 for islands that have reached the equilibrium geometry and set B has mean value of 3.26. Conversely, sets A and C display a higher number of islands that were preserved from erosion as they are carved into harder rocks: as a consequence, a wider range of island geometries is observed, including islands characterised by higher l/w and k values (up to 5.2 for set A and 4.3 for set C). It is also important to observe that both k and l/w are independent from the dimensions of the islands, as the same values are observed, in the same sets, for small and bigger islands as demonstrated for set C that displays islands with a wide range of dimensions.

ASpl values were also calculated for the entire population (Tab. 6.2): this parameter was originally created for the geomorphic analysis of drumlins by Spagnolo et al. (2010). It indicates the position of the island maximum width on its main axis and quantifies the planform symmetry of the island shape. The typical airfoil shape displays ASpl values of 0.30-0.35 with Earth streamlined islands showing values of symmetry that range from 0.3 to 0.4 (Komar, 1983). The English Channel islands display mean value of 0.44 with variations between the sets A and C (values of 0.48 and 0.44) and set B (0.37). This is also consistent with the interpretation of islands of set B being the closest to the equilibrium geometry as they display very similar ASpl values to those obtained in Komar's experiments.

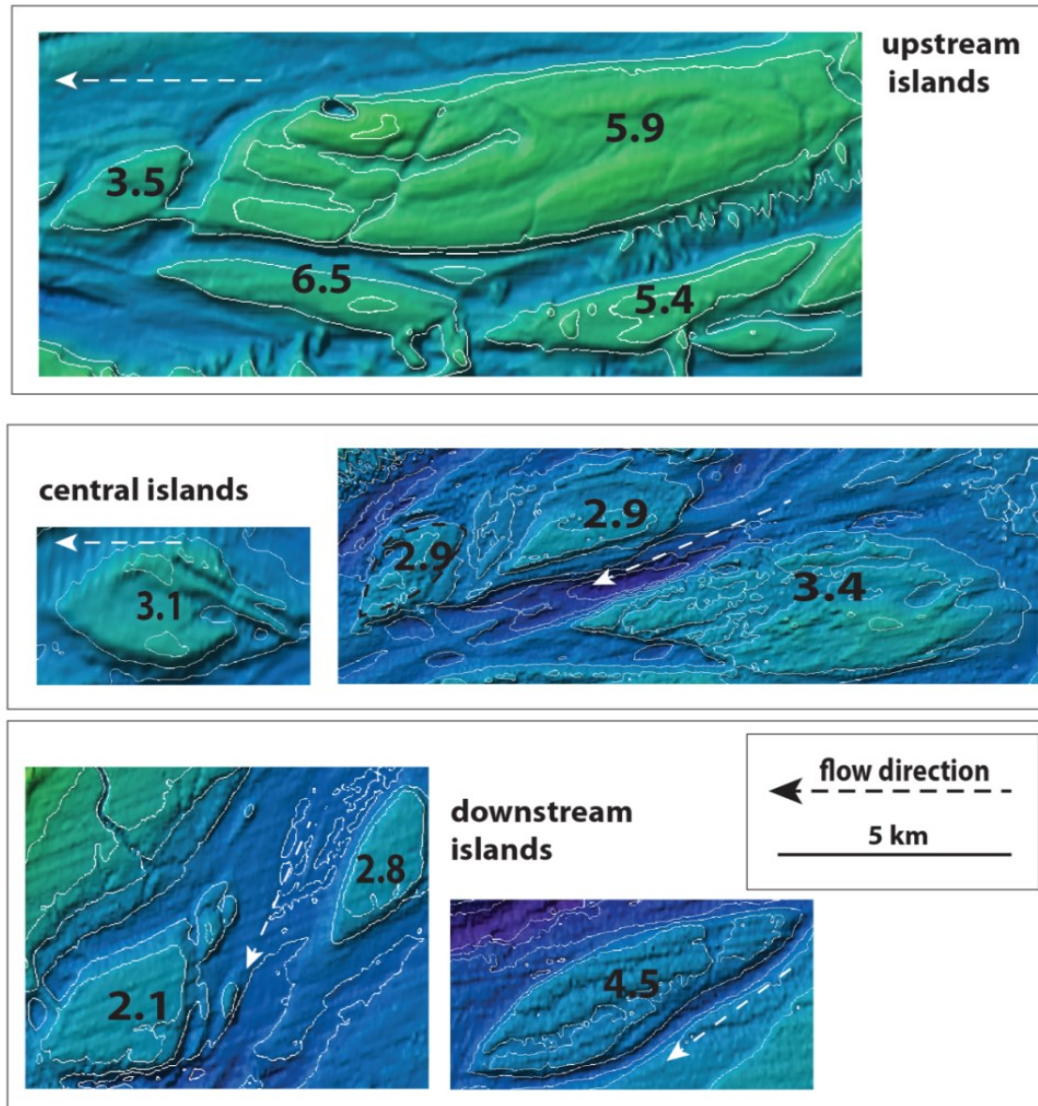


Figure 6.24: Variation of k number with the shape of the islands: central islands show typical teardrop shape which represents the equilibrium geometry. More elongated islands in the central area have been eroded as they did not reach the equilibrium and were therefore not preserved.

6.3.5 Comparison with streamlined islands in other megaflood terrains

Streamlined islands are one of the primary forms of evidence for large floods that overwhelm existing drainage networks (Kehew and Lord, 1986) and represent one of the most typical erosional remnants present in megaflood terrains. The analyses performed on the streamlined islands of the English Channel indicate how these streamlined islands present very similar character and morphology to other streamlined islands observed on Earth (Baker, 1973a,b, 1978a; Komar, 1984; Elfstrom and Rossbacher, 1985; Kehew and Lord, 1986; Meinsen et al., 2011; McClanegan, 2013) and Mars (Baker and Kochel, 1979; Komar, 1983; Burr, 2005). Despite different scales, due to different magnitude of the flows that carved these features

(with Martian islands up to 50 km long), and to different lithology into which they were carved, the English Channel islands display the same morphometric values of the islands interpreted in literature (Tab. 6.2 and Figs. 6.25-26). In particular, the similarity in scale and geometry of the interpreted islands indicates that these features were formed under similar conditions to the islands identified in other megaflood terrains. The islands are interpreted as the result of erosion by extremely high-magnitude water flows during megaflood events. As demonstrated by seismic data, English Channel islands are carved into bedrock, suggesting these features to have been eroded by exceptionally high-energy water flows.

With regard to the character of the flows that carved the islands, Baker (1978a) suggested that greater elongation of the islands could indicate higher Reynolds number and, therefore, higher flow velocities and/or depths. Baker and Kochel (1978) also demonstrated that elongation increases with higher Reynolds number (in the range of 10^8 to 10^9), based on analyses performed on the Scablands islands. This could indicate that islands that display higher elongation values (such as the islands on Mars interpreted in 1978 in Baker and Kochel paper), formed in deep, high velocity water flows with higher Reynolds numbers. Conversely, more rounded islands would have formed in slower and/or shallower flows. This indication can nevertheless only be used as a crude palaeo-hydraulic estimation as Baker and Kochel regression coefficients for these empirical relationships was low (0.79). In the case of the English Channel islands, the variation in the elongation of the islands between the three sets was interpreted with the variation in the degree of erosion of the islands, with the more elongated islands only preserved where more competent rocks outcrop. Nevertheless, comparing the entire population to islands of other megaflood terrains, the interpreted islands show values very similar to those of the Channeled Scablands and other megaflood terrains (with the majority of islands ranging from 3 to 5, as shown in Table 6.2). This, following Baker and Kochel palaeo-hydraulic estimation, would indicate that the islands were carved by flows with similar character and magnitude of Reynolds number to the flows that carved the Channeled Scablands islands, as also indicated by the similarity in the scale of the islands (Fig. 6.25).

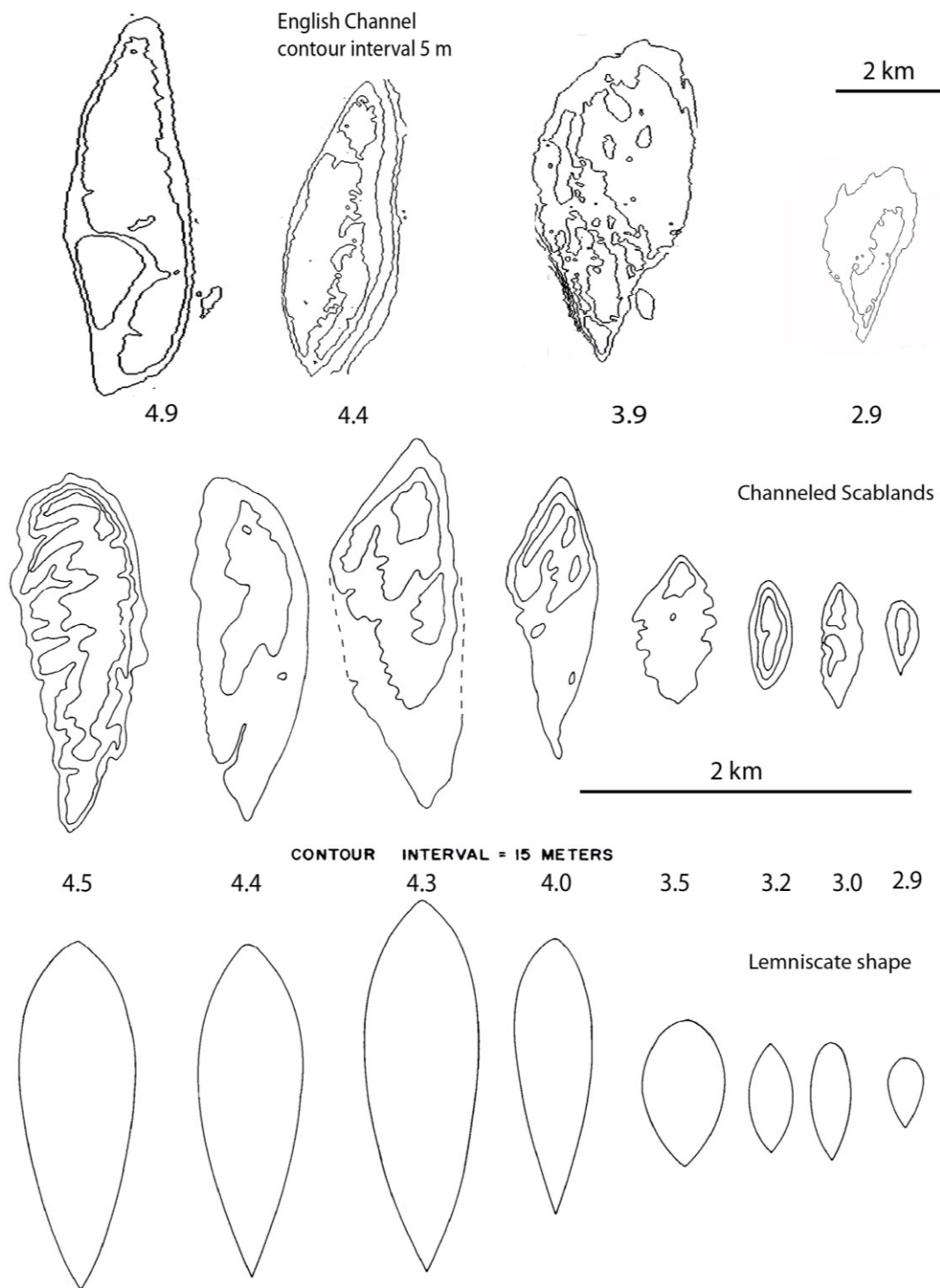


Figure 6.25: Comparison between the shape of airfoil geometry, Channeled Scablands islands and English Channel islands, in relation to their k value. Despite the difference in scale between the Channeled Scablands and the English Channel, the shape of the island is almost identical between the two sets for all the k values. Channeled Scabland islands contours from Baker and Kochel, 1978.

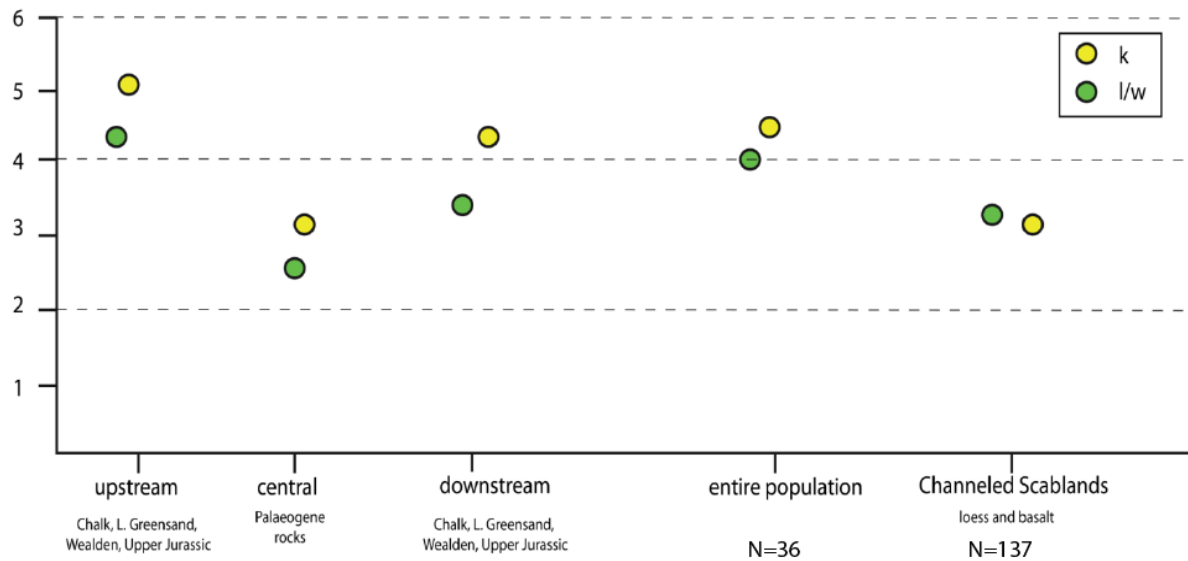


Figure 6.26: Graphic comparison of geomorphic values for the islands showing how k and l/w are strictly related. Note the variation in values for different lithology in which the islands are carved.

6.3.6 Summary of streamlined island interpretation

The following conclusions can be drawn from the analyses and interpretation of the geometry and spatial distribution of the streamlined islands in the Northern Palaeovalley:

- The geometry and spatial distribution of the islands present in the study area is strictly controlled by seabed geology and character of the rocks outcropping at the seabed. In particular, sets A and C, carved in more competent rocks display more elongated islands, a bigger population and wider range of geometries if compared to set B. This is due to lower degree of preservation of islands of set B, carved into softer Palaeogene rocks.
- The islands are similar in distribution, geometry and scale to streamlined islands identified on other megaflood terrains. This is indicated by very similar morphometric values of the islands mapped in the Northern Palaeovalley and those described in literature. This similarity suggests that the islands observed in the study area and carved into bedrock were formed by high-energy flows of water, with similar hydraulic conditions to the flows that produced the streamlined islands observed in literature.

6.4 Longitudinal lineations

Multiple sets of lineations carved into the bedrock were identified at the Dover Strait and carved into the Lobourg Channel floor (see Section 6.2, Fig. 6.5). The lineations are composed of sets of highly parallel ridges and grooves eroded into Middle Chalk bedrock on the channel floor and displaying amplitudes up to 1.2 m (Fig. 6.27). They are parallel and continuous for hundreds of meters to kilometres (Fig. 6.28), each set composed of a variable number of lineations (21 to 220 lineations for the smallest and biggest sets). The six sets (Fig. 6.29), distributed across the Strait, are located at different depths and on three different surfaces: the Lobourg Channel floor, (green sets), the channel bench and a bedrock remnant located on the Lobourg Channel floor (yellow set).

The lineations were grouped in sets according to their geometry/morphology and to their location. Although sets A,B,D and E are all located on the Lobourg Channel floor, they were grouped in different sets as they have different geometrical patterns. See Fig. 6.29 for details.

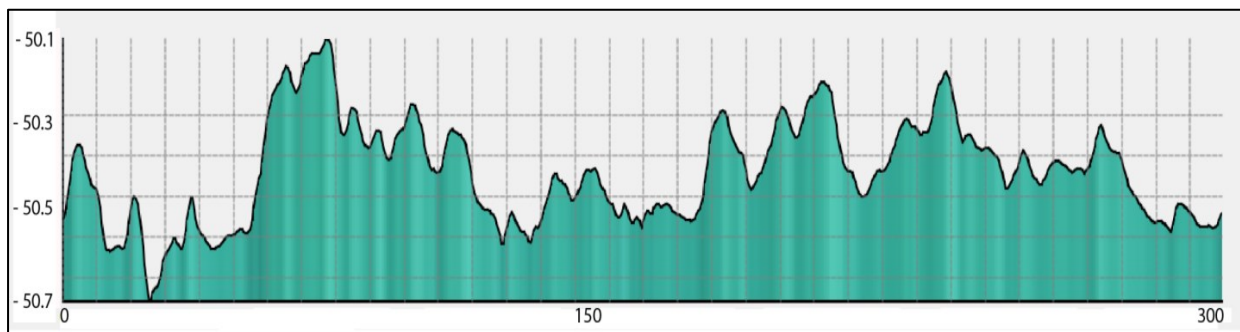


Figure 6.27: Typical profile across longitudinal lineations, showing mean amplitude of 0.2 m and regular spacing of the ridge crests of 30 to 50 m. The vertical and horizontal scales are in m.

6.4.1 Observations

The following section describes the geometry and orientation of each set of lineations, summarised in Table 6.3. A detailed analysis of the geometry of the lineations was performed for each set, mapping the features with positive amplitude at the seabed (ridges). The dimensions of the ridges were measured systematically analysing length, width, elongation ratio (length:width ratio) and crest to crest spacing (distance between adjacent ridges) of each lineation and calculating the mean values for each group.

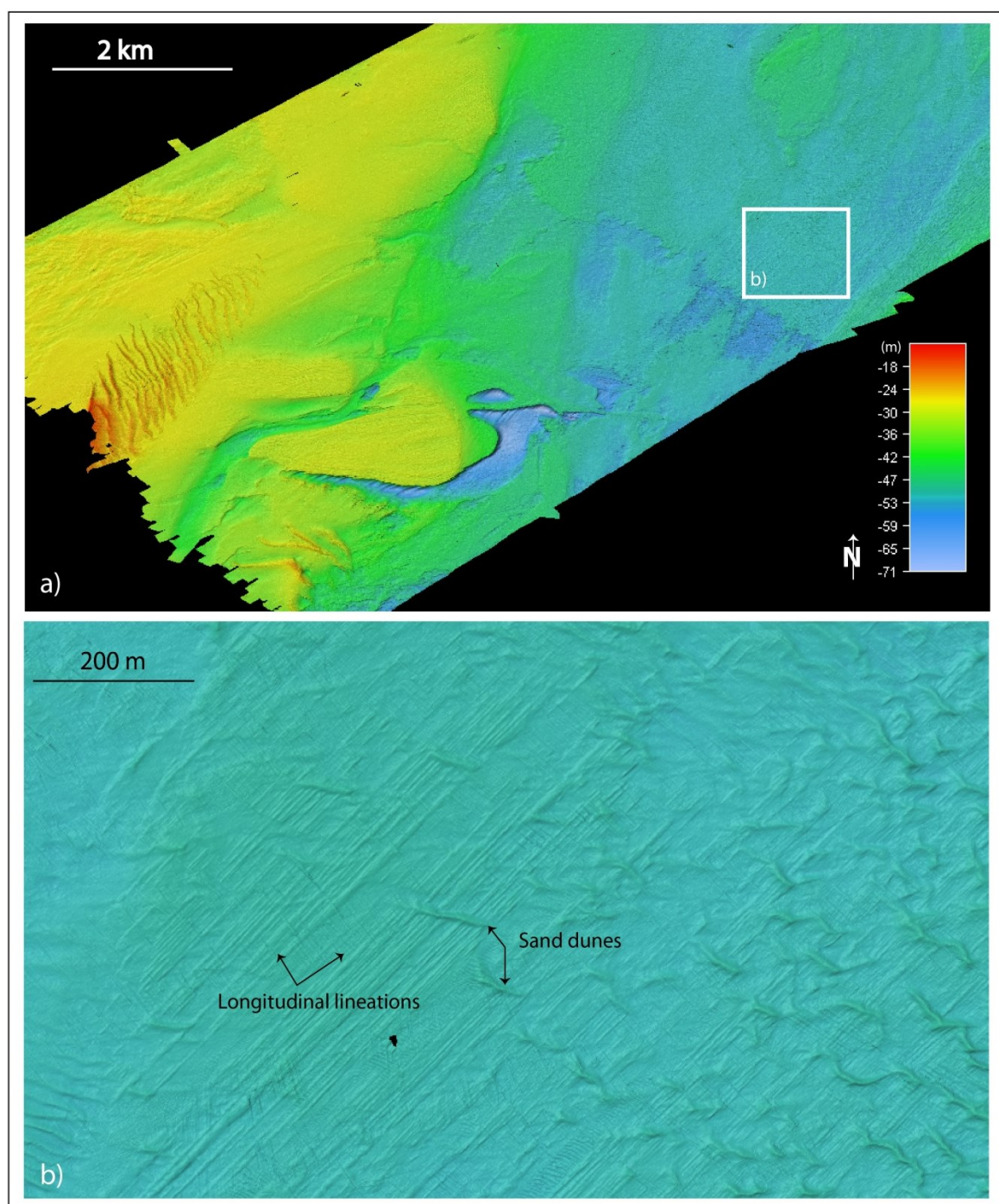


Figure 6.28: Bathymetric image of the portion of the Lobourg Channel floor (a) where the lineations are eroded into the Middle Chalk bedrock, just upstream of the Middle-Lower Chalk boundary visible at the seafloor. The 1.5 m bin size DTM in box b) shows how the lineations are continuous and parallel for hundreds of meters, locally masked by the presence of sand dunes deposited perpendicularly to the lineation main axis.

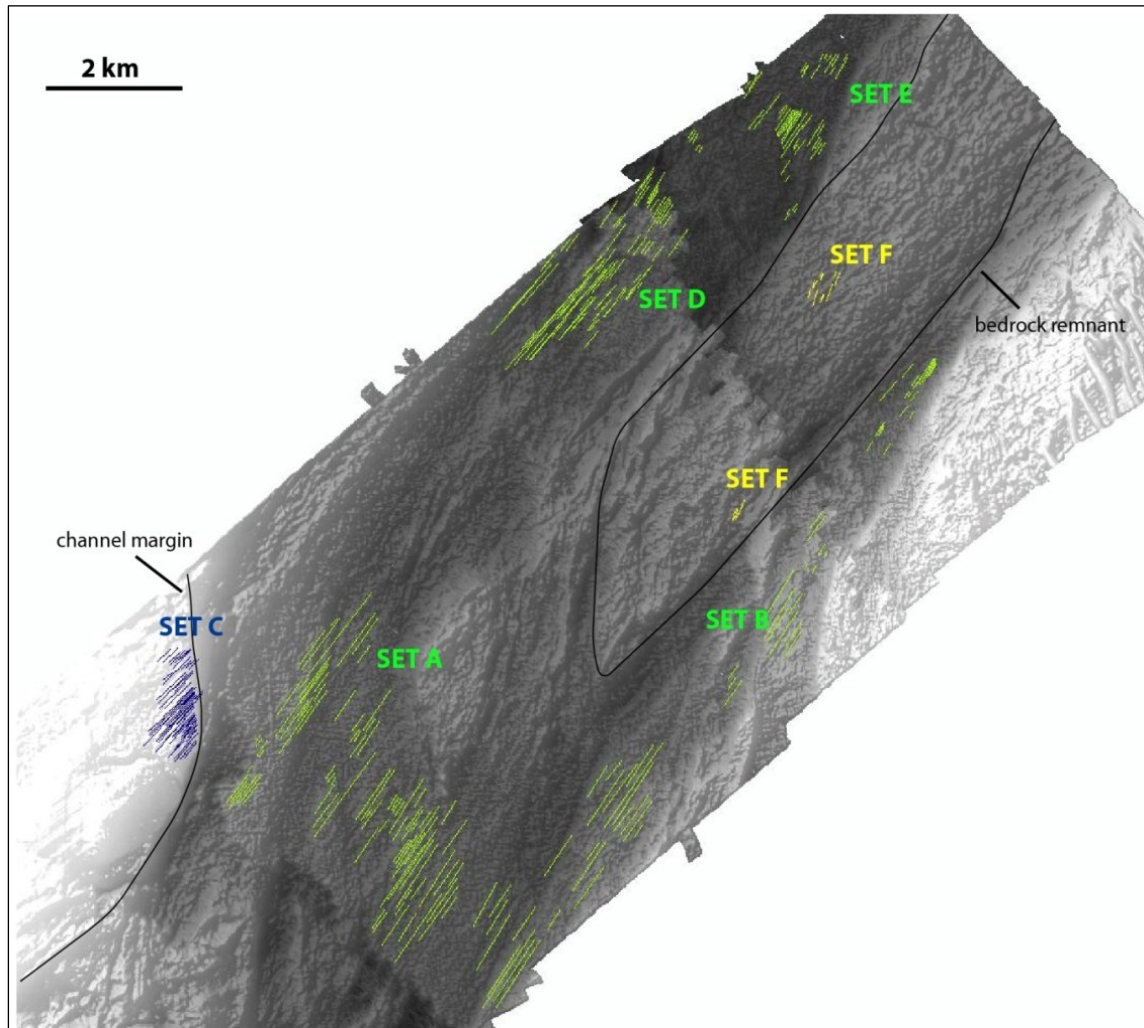


Figure 6.29: Map showing location of different sets of lineations. Each colour corresponds to a different surface on which the lineations are located. The green sets are carved into the Lobourg Channel floor (surface S6), the blue set is located on the channel bench (surface S1) and the yellow set is located on a bedrock remnant. Lineations of set A are all located in proximity of the Middle-Lower Chal boundary. Lineations of set B are located on the E of the bedrock remnant on which set F is carved. Set D and E are located on the W of the remnant. Set E is heavily masked by sand waves and its mapping is strongly affected as only small areas are sediment free.

Set A

This set of lineations is the largest in the dataset, located on the Lobourg Channel floor at depth that ranges from 47 m in the shallower, eastern part to 52 m in the western part. The bedforms show a ENE alignment (060°) and are present over an area of 3 km^2 . The lineations in the southern part of the set are overlain by sand dunes with main axis orientated NW-SE, almost perpendicular to the ridges and grooves that are locally masked by the sediments (Fig. 6.28b). The sand dunes have a crest to crest spacing of 80 to 150 m and measure 0.2 to 1 m in amplitude, 120 m in length and 60 m in width. The lineations are between 500 and 5000 m

long and 5 to 30 m wide, most being ~15 m. The mean amplitude measured on the grooves is 0.5 m, with values up to ~1.2 m.

The length: width ratio ranges from 25:1 for the smaller grooves up to 80:1 for the major observed features. Crest to crest spacing varies from ~30 to 50 m.

Set B

This set is located in the central and deepest (~52 m) portion of the Lobourg Channel floor. The grooves are present over an area of 3 km². The lineations display the same azimuth as set A (ENE 060°) and are overlain by sand dunes showing NW-SE orientation of their main axis. The grooves measure 0.2 to 0.5 m in amplitude (Fig. 11) with a width that varies from 5 to 10 m and length of ~200 to 2000 m. The length: width ratio is 40:1 up to 60:1 and the crest to crest spacing is 50 m.

Set C

A third set of grooves, displaying smaller features with respect to sets A and B, is located on the bench that bounds the NW flank of the Lobourg Channel at 32 m depth, lying 20 m above the Lobourg Channel floor. The lineations are present over an area that measures ~1.8 km². The lineations display a slightly different orientation respect to sets A and B, being aligned ENE (040°). The grooves are ~400 to 600 m long and ~5 to 15 m wide with amplitude of ~0.4 m. The length:width ratio is 50:1 up to 80:1 and the crest to crest spacing 8 to 10 m. By contrast to sets A and B, there is no evidence of sediments masking the seabed and the bedrock lineations in this area.

Sets D and E

Both sets D and E are similar in character to sets A and B, described above in this section. They display the same ENE 060° azimuth although they are located further upstream on the channel floor over an area of 24 km² and show a slightly smaller population compared to sets A and B. The erosional grooves of set D lie at a depth of ~50 m and are characterised by a length that varies from 120 to 2000 m, width of 5 to 10 m and amplitude up to 0.8 m. Grooves of set E are located at ~45 m of depth and display a length that varies from 90 to 650 m, width of 5 m and amplitude up to 0.8 m. Crest to crest spacing and amplitude of the grooves is respectively 15 to 30 m and 0.5 m and the length:width ratio ranges between 25:1 and 60:1. The lineations of sets D and E are partially masked by seabed sediments (Fig. 6.30) as, moving towards the NE into the channel floor, the sediment cover becomes gradually

ticker and more continuous. In particular, the area where set E is located, displays sand waves with amplitude up to 2 m at the seabed, crest to crest spacing of 30 m and groups of sand waves that are laterally continuous for up to 600 m.

As shown by Fig. 6.30, the two sets probably represent a single large set that is masked in its central area by the thick sediment cover.

Set F

The lineations interpreted as set F are located on a bedrock remnant at 56 m of depth, elevated 5 m above the Lobourg Channel floor. This set is composed of a very small number of grooves (21) as only a small part of the erosional surface on which the grooves were carved has been preserved from erosion. The grooves display an ENE $\sim 080^\circ$ azimuth with a length and width that are remarkably lower with respect to the lineations described above. This is probably due to a preservation issue as the lineations are located over a small area (600 m^2) represented by a partially eroded bedrock remnant. The length of the lineations varies from 65 m to 260 m and width 5 to 6 m.

Lineation Set	Crest to crest spacing	Elongation ratio	Length	Amplitude
A, B, D, E	50 m	25:1-80:1	500-4000 m	0.5-1.2 m
C	10-15 m	40:1-60:1	200-600 m	0.2-0.5
F	10-20 m	50:1-80:1	400-600 m	0.2-0.5

Table 6.3: Summary of the measured parameters for each set of lineations.

The aspect ratio is $\sim 15:1$ to $50:1$ while the crest-to-crest spacing and amplitude of the grooves are respectively 15 m, and 0.1 to 0.2 m. Since the amplitude of the grooves is lower than the other sets, only the features that clearly showed an orientation different from linear artefacts in bathymetry data were included in the set in order not to introduce artefacts in the population. There is no evidence of seabed sediments masking the lineations on the bedrock remnant.

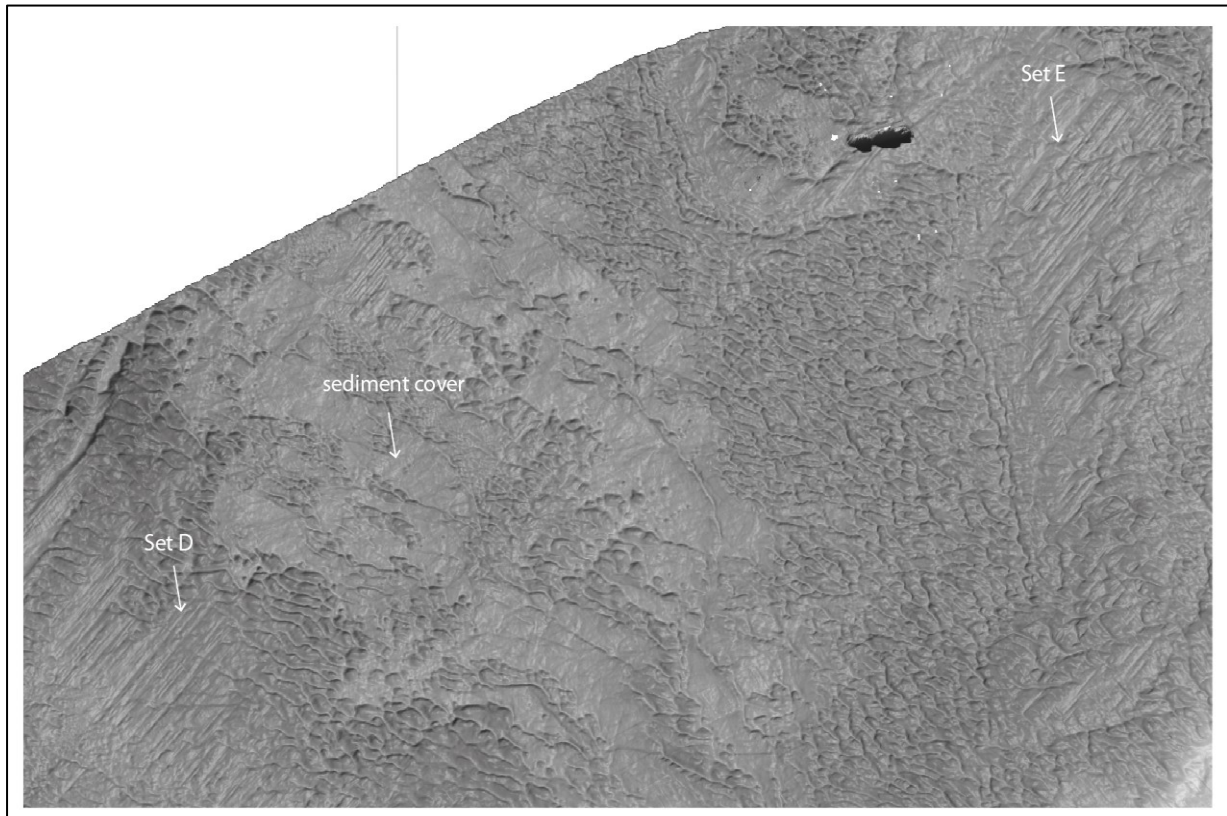


Figure 6.30: Image of sets D and E locally masked by groups of sand waves. The two sets are separated by the sediment cover that becomes almost continuous on the channel floor in the central part of the figure. Depth varies between 40 and 70 m.

6.4.2 Interpretation

Orientation of the lineations

A rose diagram showing the orientation of the different sets of lineations is displayed in Fig. 6.31, using the same colour as in Fig. 6.29, in order to discriminate the location of the sets carved into different erosional surfaces. Figure 6.31 shows how the four sets A, B, D, E, all located on the Lobourg Channel floor (S6), display the same orientation. Sets C and F, respectively located on the bench (~20 m elevated from the channel floor) and on the bedrock remnant (~5 above the channel floor), display different orientations. Since longitudinal lineations are used in geomorphology as palaeo-flow indicators, as their elongation axis is parallel to the direction of the palaeo-flow that carved the lineations, the rose diagram was produced in order to analyse the orientation of the features and therefore of the flows that eroded them. The rose diagram indicates three main azimuths for the interpreted lineations, suggesting that the flows that carved the lineations into the bedrock had more than one

direction. In particular, the diagram shows three preferential orientations of the lineation azimuth, one for each surface on which the sets of lineations are located.

This could indicate that different flows with different directions eroded the lineations located on the three surfaces, possibly at different times as the erosional surfaces on which they are located are not synchronous.

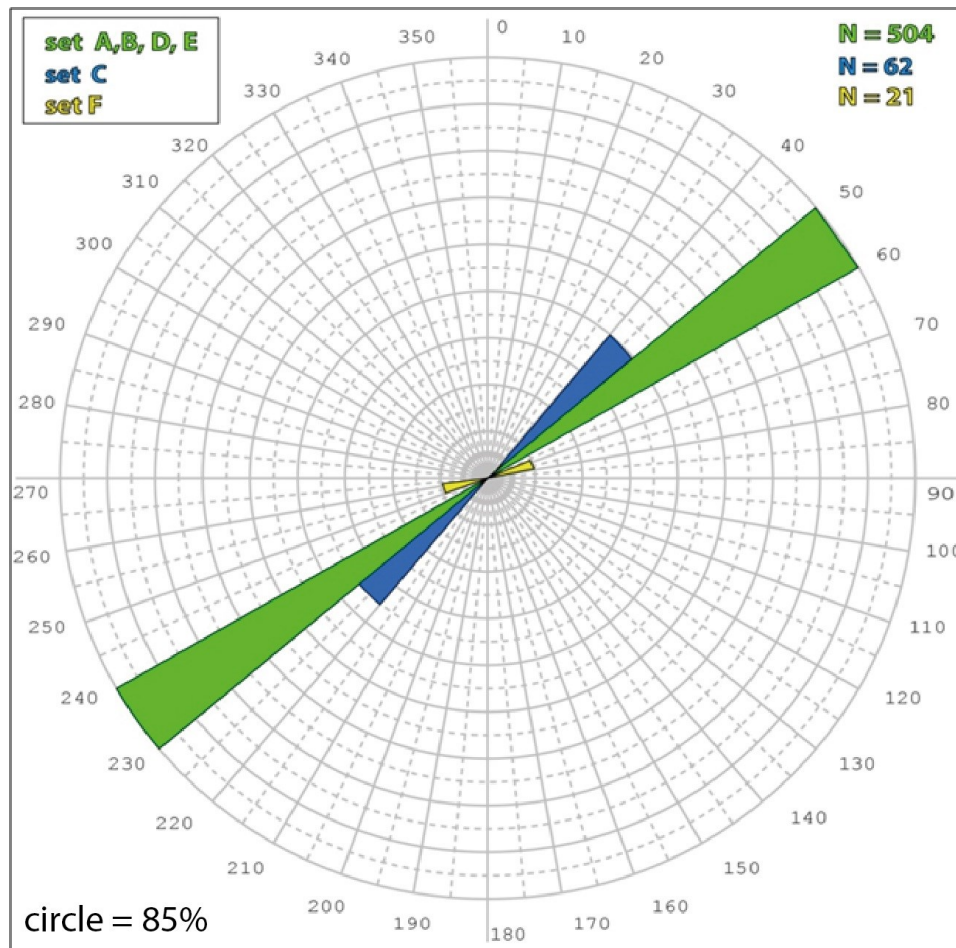


Figure 6.31: Rose diagram showing orientation of the interpreted sets of lineations, represented with the same colour as in Fig. 6.29.

Origin of the longitudinal lineations

With regard to the interpretation of the process that carved the grooves, a tidal origin has been ruled out as sand waves, in which deposition is mainly controlled by tides, are superimposed on the lineations and show an orientation of their main axis perpendicular to the orientation of the lineations. This suggests that the lineations would have the same azimuth of the sand waves if their formation was controlled by tidal processes. Also, lineations are only present on the channel floor and carved into the Chalk bedrock, suggesting

that their erosion is connected to the formation of the channel and of the erosional surfaces on which they are located and were only successively overlain by sand dunes. In addition, while longitudinal lineations are abruptly truncated at the boundary between Middle and Lower Chalk, sand waves are persistent throughout the area maintaining the same orientation and draping the eroded Lower Chalk bedrock of the Lobourg Channel floor.

The interpreted lineations show striking similarity to mega-scale glacial lineations (MSGSL) that are found as relict expression of fast flow on palaeo-ice stream beds and were. These were first described by Clark (1993) who observed kilometre-scale longitudinal lineations carved by the former North American ice-sheet in Canada. MSGSL are usually distinguished from other glacial lineations like bedrock grooves depending on their scale: Clark (1993) defines as MSGSL those features with crest-to-crest spacing of 300 to 5 km and length that ranges between 8 and 70 km in the case of the lineations interpreted in Canada. Other authors (e.g. Heroy and Anderson, 2005) identify MSGSL depending on the lineation elongation ratio (typically $> 80:1$), other than their crest-to-crest spacing (200 to 600 m). Finally, in opposition to bedrock grooves, MSGSL are found on thick sequences of sediments unconstrained from bedrock (Wellner et al., 2001; Heroy and Anderson, 2005) and display higher uniformity in their morphology throughout wide areas, as for the case of the MSGSL found in Antarctica. In many cases (e.g. Clark, 1993; Stokes and Clark, 1999; Wellner et al., 2001; Graham et al 2009, 2010), MSGSL are typically found in association with drumlins, flutes and other glacial bed-forms at different scales, showing cross cutting relationship between different sets of MSGSL and with other glacial landforms of different scale such as drumlins and flutes with which they are associated (Clark, 1993). When multiple sets are present, their cross-cutting relationships and orientation are used as ice sheet palaeo-flow indicators as MSGSL are parallel to the direction of the flow, providing evidence of changes in the palaeo-flow direction. Also, the geometry of the MSGSL termination gives indication of their age: wedge front termination of MSGSL and good preservation in soft sedimentary strata indicates they formed during recent recession (e.g. Lowe and Anderson, 2002; Graham et al., 2010) rather than older lineations that were overridden by younger features (e.g. Ó Cofaigh et al., 2005).

Since the discovery of MSGSL numerous studies have been conducted in different areas where former and modern ice-sheet are located, with particular interest in the Antarctic area (e.g. Ship et al., 1999; Anderson et al., 2002; Heroy and Anderson, 2005; Evans et al.,

2005; King et al., 2009; Graham et al., 2010) where MSGSL are widely observed. The complexity in the direct observation of actively forming MSGSL makes it necessary to base the understanding of the processes that generate these features on the study of ancient features, formed sub-glacially, that are nowadays exposed sub-aerially. As these features tend to form in groups or fields, rather than co-exist with lineations of different scale, scale-dependent processes are thus implied in their formation (Clark, 19993). Two main mechanisms are suggested for the formation of MSGSL: the first theory suggesting that the motion of the ice stream is a combination of till deformation and basal sliding (e.g. Livingstone et al., 2012), producing MSGSL (e.g. Clark, 1993; Stokes and Clark, 2001; Ó Cofaigh et al., 2002; Lowe and Anderson, 2002; King et al., 2009; Graham et al., 2010). This theory indicates that ice streaming is dependent on a soft, water saturated bed deformable bed with minimal frictional resistance (King et al., 2009) and is supported by sedimentological evidence, indicating that MSGSL are typically found in association with sub-glacial deformation tills (Lowe and Anderson, 2002), giving indication of palaeo-ice stream activity. In particular, cores from sub-glacially deformed sediments (Anderson, 1999; Domack et al., 1999; Evans and Pudsey, 2002) show three characteristic facies (diamicton facies at the base, pebbly and sandy mud intermediate facies and siliceous mud with terrigenous silt and clay at the top) with thickness that varies from decimetres to 1000 m (Heroy and Anderson, 2005). Other models for the formation of MGLS through fast-flowing ice sheet imply either groove-ploughing through keel of ice over bedrock (Clark et al., 2004) or transverse flow generating spiral mechanisms in the basal ice-sheet (Schoof and Clarke, 2008; Fowler, 2010). The second suggested mechanism for the formation of MSGSL consists of turbulent flows originating from catastrophic discharge of ice-sheet melt water (Shaw, 1983, 1994; Shaw et al., 2008), also responsible for the formation of associated features such as drumlins and flutes. Shaw (2010) suggests that melt-water megalineations generated by catastrophic events are found both in deformed silt (e.g. Shaw et al., 2008) and carved into bedrock (Kor et al., 1991; Bradwell, 2005; Gupta et al., 2007), considering all features under the name of megalineations, independently from their glacial or flood-related origin.

In the case of the longitudinal lineations observed at the Dover Strait, their length (up to 4 km) and crest to crest spacing (10 to 50 m) is one order of magnitude smaller than MSGSL interpreted in literature (Table 6.4), although they display similar values of elongation ratio (ranging between 25:1 and 80:1) to Antarctic MSGSL (Ó Cofaigh et al., 2005; Heroy and Anderson, 2005; Evans et al., 2005; King et al., 2009; Graham et al., 2010). As

shown by the rose diagram, the observed lineations display multiple azimuths but there is no evidence of cross-cutting relationship between the interpreted sets as they are located in different portions of the seabed and they do not override one another.

Location	Reference	Interpretation given by author	Formation mechanism	Crest to crest spacing	Elongation ratio	Length	Amplitude
Canada	Clark, 1993	MSGL		300-5000m	na	8-70 km	na
Antarctic Peninsula	Heroy and Anderson, 2005	bedrock grooves	streaming ice-sheet	100-3000 m	20:1/	5 – 10 km	na
		MSGL	deformation till under streaming ice-sheet	200-600 m	>80:1/	>20 km	10-20 m
Ross Sea	Ship et al., 1999	MSGL	subglacial flutes and furrows	300-650 m	na	8 - >20 km	na
		grooves	iceberg furrows	na	na	10 km	5 - 30 m
Keewatin, Canada	Greenwood and Kleman, 2010	till MSGL	subglacial	na	5:1 -31:1	9-35 km	10 -20 m
NW Scotland	Bradwell, 2005	bedrock megagrooves	Subglacial meltwater	na	na	0.3-4.3 km	5 - 20 m
Eastern Svalbard	Hogan et al., 2010	MSGL	Glacial lineation	na	> 65:	3-10 km	4 -8 m
Dover Strait	This work	Erosional groove Set A,B,D,E	Megaflood grooves	50 m	25:1-80:1	0.5-4 km	0.5-1.2 m
		Erosional groove Set C		10-15 m	40:1-60:1	0.2-0.6 km	0.2-0.5
		Erosional groove Set D		10-20 m	50:1-80:1	0.4-0.6 km	0.2-0.5

Table 6.4: Comparison between mega-scale longitudinal lineations interpreted in literature and the lineations observed at the Dover Strait.

Furthermore, while MSGL are typically associated to other glacial bedforms like flutes and drumlins and often display transitional drumlin-to-lineation features (Graham et al., 2010), there is no evidence of flutes or transitional features in the area where the longitudinal lineations are observed at the Dover Strait. Furthermore, following the fast flowing ice stream interpretation for the formation of MSGL, these features are found in deformed glacial sediments showing typical facies succession; there is no evidence of deformed tills in the area where the lineations are observed at the Dover Strait as they seem to be entirely carved into Chalk bedrock. This is indicated by high-resolution seismic data, showing that the lineations located just upstream of the cataract are carved into bedrock, represented by Middle Chalk rocks and showing typical acoustic texture on bathymetric data. The same acoustic texture of the seabed is present further upstream, indicating that all the mapped sets of lineations are carved into bedrock, although seismic data is not available in the upper portion of the Lobourg Channel. Finally, while the formation of MSGL is associated to the presence of a palaeo-ice stream, previous studies on the location of the British Ice Sheet margin during Pleistocene glaciations (See Section 2.4) indicate that ice sheet was located further north (e.g. Clark et al., 2004; Gibbard and Clark, 2011), suggesting that the Dover Strait area has never been occupied by a palaeo-ice sheet. This, taken together with the absence of other glacial bedforms associated to the mapped lineations indicates that, based on the available data, the features observed at the Dover Strait cannot be classified as MSGL.

If a glacial interpretation for the formation of the Dover Strait longitudinal lineation cannot be proposed as it is not consistent with the observations made on the data, another group of kilometre-scale lineations carved into bedrock and not related to glacial erosion is represented by megaflood erosional grooves. These features were first interpreted by Baker and Nummedal (1978) in the Channeled Scabland megaflood terrain (Fig. 6.33). These are parallel grooves and ridges, continuous for kilometres, measuring up to 10 km in length, amplitude of 5 to 25 m and spacing of 30 to 500 m. The erosional grooves interpreted by Baker and Nummedal (1978) in the Scabland terrain are carved into basalt and located at the head of the Dry Falls cataracts. The formation of such features has been explained as the result of erosion by macro-turbulence in the form of longitudinal vortices, developed in high-magnitude flood flows (Baker and Nummedal, 1978; Baker, 1979).

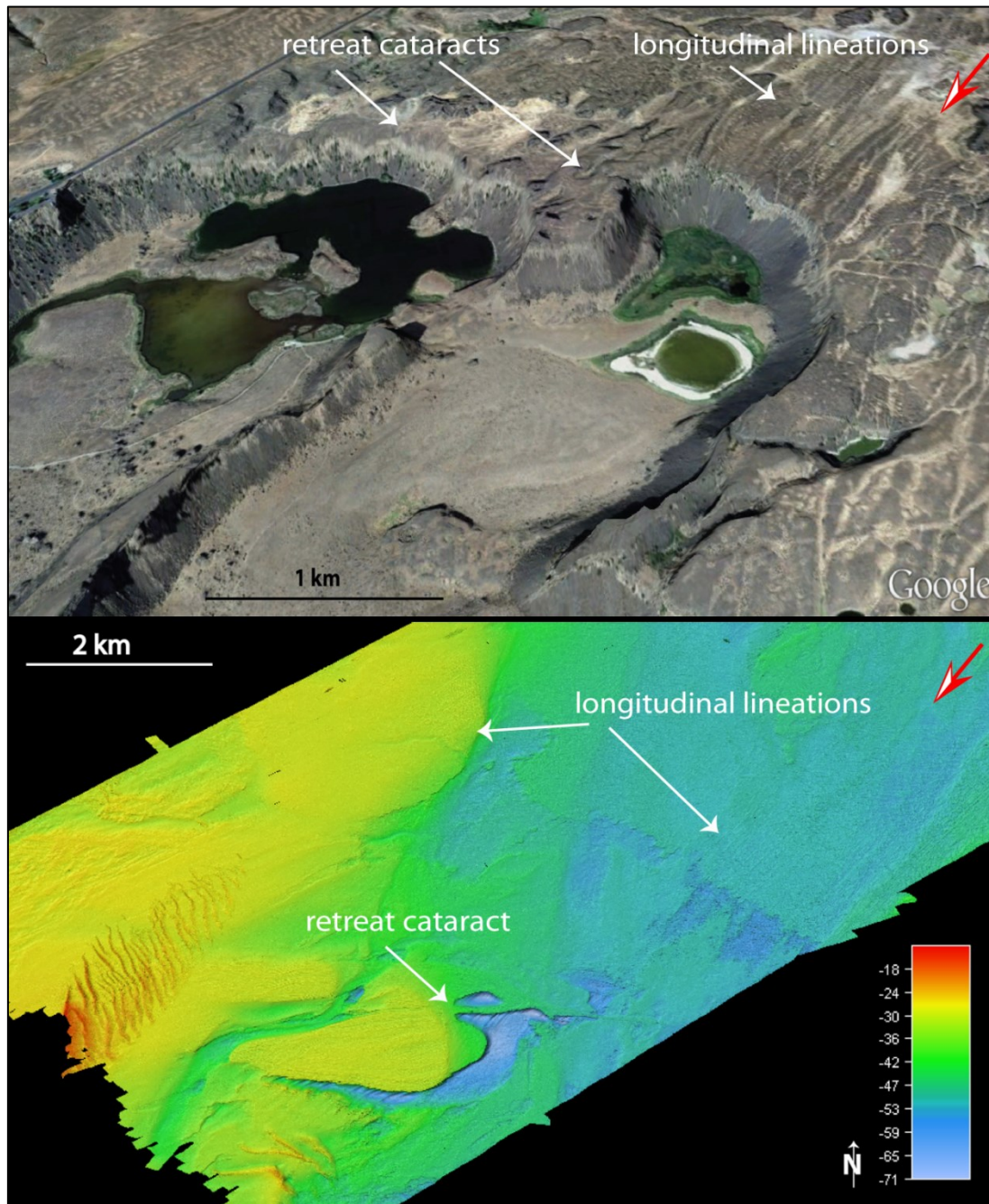


Figure 6.32: Comparison between the longitudinal lineations located at the head of the Dry Falls cataract complex in the Channeled Scabland (source Google Earth) and the longitudinal lineations at the Dover Strait, also located less than 2 km upstream of the head of the retreat cataract on the Lobourg Channel floor. The red arrows represent the suggested direction of the palaeo-flow.

Similar features has been observed by several authors in other megaflood terrains on Earth and Mars (e.g. Baker and Milton, 1974; Baker and Kochel, 1979; Komatsu et al., 2009; Warner et al., 2009). From laboratory experiments on straight streams, performed by Einstein and Li (1958) and Shepherd (1972), these elongate grooves are interpreted to be the product

of a helical array of vortices with alternating sense of rotation and longitudinal axes oriented in stream-wise direction. In natural channels, the roughness and irregularities present on the channel floor create additional vorticity and, where longitudinal bedforms are created, their presence further enhances the flow field, leading to stronger longitudinal erosional (Baker, 1979) as feedback effect. Whenever a defect occurs in the channel bed, longitudinal vorticity might be broken, producing transversal bedforms such as cataracts and flow steps that take the place of lineations.

If the longitudinal lineations observed at the Dover Strait are interpreted as longitudinal bedforms created by vortices, the feedback mechanism explained by Baker and Nummedal (1978) would be consistent with the observed distribution of the bedforms, abruptly truncated at the boundary between Middle and Lower Chalk. No lineations are in fact observed in the area where the Lower Chalk outcrops at the seabed, while the portion of the Lobourg Channel floor dominated by Middle Chalk shows a high density of lineations. The boundary between Middle and Lower Chalk marks, as witnessed by the 5 m scarp at the seabed, a change in the resistance to erosion and in seabed texture between the harder Middle Chalk. The presence of lineations being restricted to Middle Chalk bedrock is therefore likely to be due to a change in the lithological properties of the seabed. As the flow encountered a change in lithology, varying from the smooth and homogeneous substrate of the Middle Chalk to the more irregular and heterogeneous Lower Chalk, a change in the erosional style of the flow occurred. This is indicated, other than by the abrupt termination of longitudinal lineations, by a patchy alternation of scoured areas showing rough and irregular seabed acoustic texture alternated to bedrock remnants with up to 5 m relief at the seafloor. The Middle to Lower Chalk boundary would therefore represent the boundary between the formation of longitudinal bedforms carved into the smooth substrate of the Chalk to the production of irregular, scabland-like topography (Baker and Nummedal, 1978) into the more erodible and heterogeneous Lower Chalk.

6.5 Palaeo-hydraulic calculations

6.5.1 Introduction and aims

The main distinctive characteristic of megafloods is their high magnitude and erosional power: a flow is identified as megaflood when its discharge is comparable to ocean currents (Baker, 2009c). It is therefore crucial to estimate the volume of water involved in the formation of the Northern Palaeovalley channel network, in order to estimate the magnitude of the flood and to compare it to other megaflood flows on Earth.

In particular, the following questions were addressed: a) is the calculated discharge consistent with a catastrophic origin of the bedforms observed on the English Channel floor? b) what is the estimated volume of water involved in the process and what are the global scale implications of such event? c) is the discharge calculated for the English Channel megaflood consistent with the typical megaflood discharges calculated from other megaflood terrains on Earth?

In order to estimate the flow magnitude and compare it to the discharge of other Terrestrial catastrophic flows, it is necessary to perform a palaeo-hydraulic reconstruction of the flood peak discharge (See Section 2.2 for background information). Following the individuation and interpretation of the main flood channels, flood surfaces and relative timing of their formation, it was possible to measure the channel geometry parameters required for the palaeo-hydraulic calculations. In particular, two main objectives were pursued through the calculation of the peak discharge:

- To estimate the magnitude of the flows that carved the interpreted channels in the Dover Strait area, in order to test if the peak discharge of the flow is consistent with the catastrophic breaching of the Strait. To do so, several profiles were calculated across the Dover Strait and the Lobourg Channel, considering two distinct erosional events that carved the Strait, as interpreted from the erosional surface analysis detailed in Section 6.2.

- Calculate the peak discharge of the flood throughout the entire English Channel palaeovalley network, in order to understand how the erosional power of the flow changed from the upstream to downstream area. To do so, multiple profiles were calculated across the Northern Palaeovalley distinguishing, where possible, two erosional events and measuring two distinct channel sections (one for the first and one for the second event) for each profile.

6.5.2 Description of channel parameters

The channel parameters needed for the calculation of the peak discharge were measured drawing perpendicular profiles across the interpreted flood channels, obtaining a cross-channel section from which the main channel parameters were calculated. In particular, the required parameters were channel bankfull width and depth, for the cross sectional area, and slope of the channel. Depth and width were measured considering the bankfull surface of the channel (see Fig. 6.34 and Section 2.2.5). Bankfull depth represents the average vertical distance between the channel bed and the estimated water surface elevation required to completely fill the channel. Bankfull width is the measurement of the lateral extent of the water surface elevation perpendicular to the channel at bankfull depth. Where two events were recognised, width and depth were measured using channel benches as upper limits for the second event.

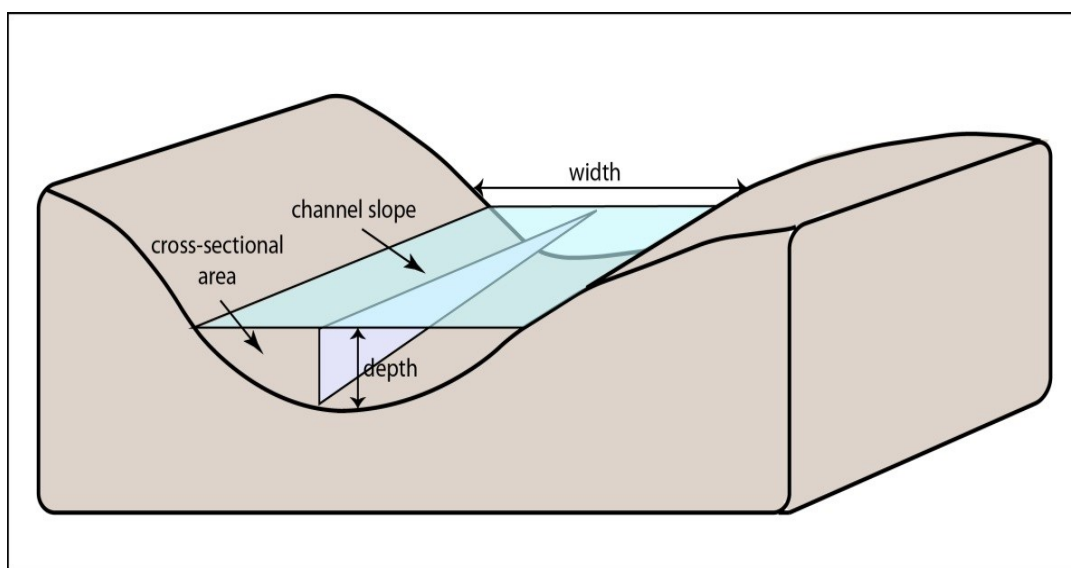


Figure 6.33: Cartoon of channel cross section and the parameters used for the peak discharge calculation.

With regard to the slope measurement, a mean value was considered for each channel profile. The mean value was identified through the measurement of the slope on multiple transects, drawn perpendicular to the channel profile, and defining a mean value. Where two erosional events and therefore two channel sections were identified, the slope value of the first event was measured on the bench, while the slope for the second event was measured on the channel floor (Fig. 6.34). Two main issues affect the measurement of the slope: the local variation of the slope and the variation of the slope with respect to the time of the flood.

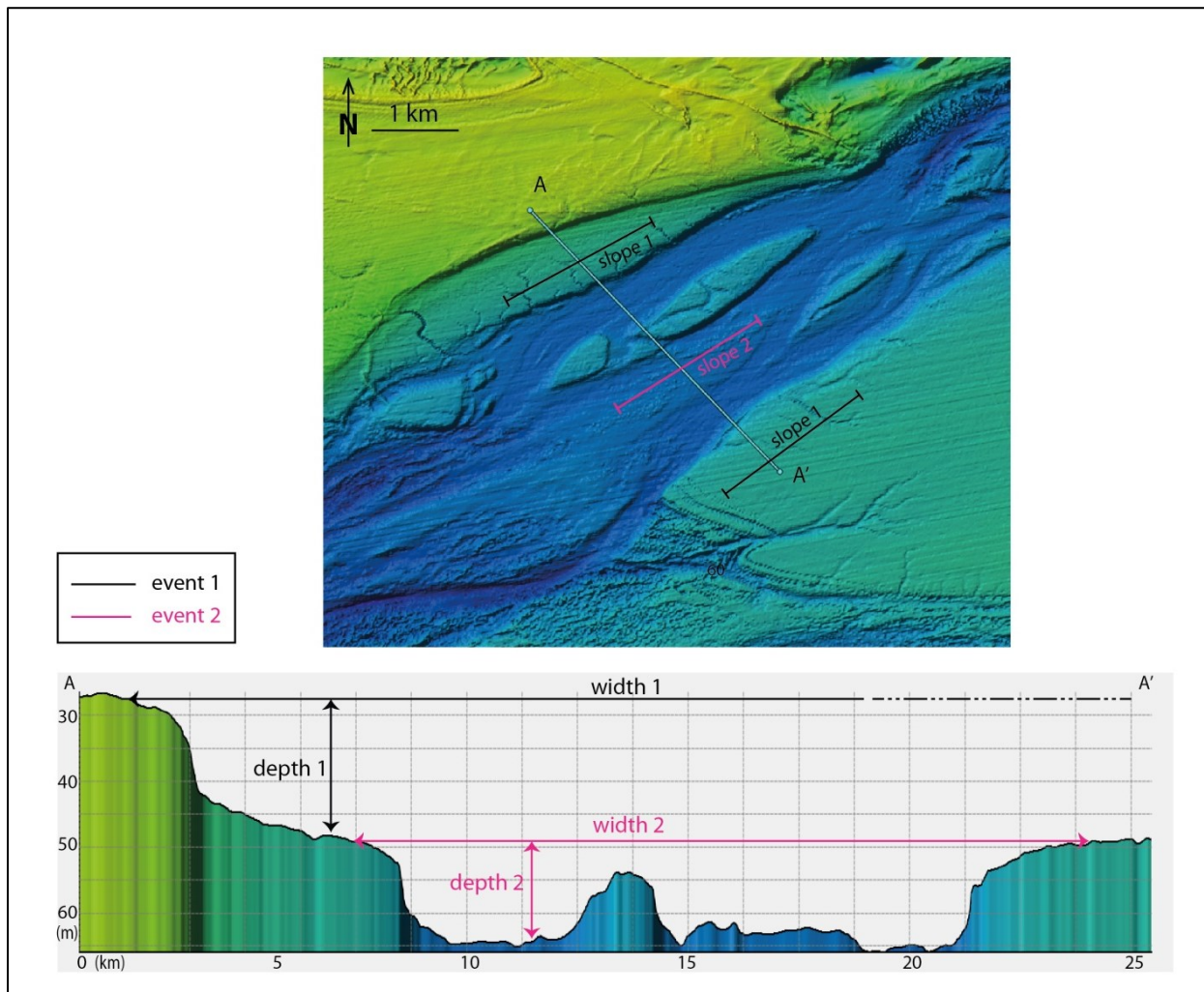


Figure 6.34: Example of channel section from which the channel parameters were calculated. The black lines represent the measurement relative to event 1, the pink lines represent slope, width and depth of the channel for event 2. Colour-scale as in Fig. 6.20.

The first problem is mainly related to the resolution of the data, as low resolution does not allow for measurement of local variations in the slope. This can lead to the measurement of a

value that is not representative of the channel section. In order to overcome this issue, multiple measurements were performed, calculating a mean value for each channel profile. Where it was not possible to define a mean value, a series of calculations were made using different slope values: transects with slope value between 0.009 and 0.0001 were considered. This problem particularly affected the downstream area as the resolution of the dataset is lower and it was not possible to identify a precise value for the slope of the channels for each point. The second issue is related to the change in the slope value throughout time, usually due to uplifting or tectonic movements. This problem particularly affected the Lobourg Channel area, as the Dover Strait underwent isostatic uplifting after glacial times (e.g. Lambeck, 1995). This probably explains the negative slope values (channel floor locally dipping towards the North Sea) that are currently measured on the floor of the Lobourg Channel. Where this problem occurred, the value of the slope was measured on adjacent transects on the same surface, assuming a constant value for the channel slope in the same area and testing a range of values rather than a single slope measurement.

6.5.3 Peak discharge calculation

Manning and Darcy-Weisbach formulas were used to calculate the peak discharge as channel roughness parameters used on similar flood channels were available from previous works in either the English Channel (Gupta et al., 2007) or the Channeled Scablands (Baker and Nummedal, 1978). In order to check the reliability of the results obtained with the two different formulas, results were compared and validated only when results with the same order of magnitude were obtained.

As high level of error is expected in palaeo-hydraulic calculations, the aim of this reconstruction was to estimate the order of magnitude of the discharge obtained from the measured channel parameters, rather than obtaining precise values. In addition, as accuracy and reliability of results are strictly related to chosen values of roughness coefficient and channel parameters, a different range of values was tested for the two formulas.

In this section, a brief introduction to the formula is given, followed by analyses on accuracy and sensitivity of the parameters in the formula.

Manning's equation

$$Q = (R \cdot W / n) \cdot R^{2/3} \cdot s^{1/2}$$

with **s** = slope in degrees, **R** = hydraulic radius for which depth (*d*) of the channel is used as proxy (only valid for channels with width >> depth) and **W** = channel width. **n** = Manning roughness coefficient. This value was selected following **n** value used by Gupta et al. (2007) for the English Channel and by Baker and Nummedal (1978) for the Channeled Scabland, assuming that the **n** value of the English Channel Chalk is equal or lower than the **n** value of channels carved into basalt. In particular, Baker and Nummedal (1978) adopted a value of 0.04 from Chow's (1959) empirical tables for channel floors and 0.05 for channel walls. For the modelling of Lake Bonneville Flood, which also flowed into basalts, O'Connor (1993) used **n** = 0.03 for the main channel and **n** = 0.05 for the valley walls.

Darcy - Weisbach equation

$$Q = R \cdot W \cdot [(8 \cdot g \cdot R \cdot s) / f_c]^{1/2}$$

with **s** = slope, **R** = hydraulic radius -> depth (*d*), **W** = channel width and **f_c** = Darcy-Weisbach friction factor. A value of **f_c** = 0.2, was used, as suggested in Clarke et al. (1984) for the Channeled Scablands.

Roughness coefficient

For Manning's formula, a range of **n values** was tested, varying from 0.03 to 0.05 and obtaining variations of 0.1 Sv (Table 6.5). The tested values were selected following values used by Gupta et al. (2007) for the English Channel and by Baker and Nummedal (1978) for the Channeled Scabland. It was assumed that the **n** value of the English Channel Chalk is equal or lower than the **n** value of channels carved into basalt. In particular, Baker and Nummedal (1978) adopted a value of 0.04 from Chow's (1959) empirical tables for channel floors and 0.05 for channel walls.

For the modelling of Lake Bonneville Flood, which also flowed into basalts, O'Connor (1993) used $n = 0.03$ for the main channel and $n = 0.05$ for the valley walls. It was decided to apply $n = 0.04$ in the formula for consistency with calculations by Gupta et al. (2007) to which the obtained results were compared.

For Darcy-Weisbach formula a value of $f_c = 0.2$ was used, as suggested in Clarke et al. (1984) for the Channeled Scablands.

Channel profile

With regards to channel geometry, detailed analyses on two different erosional surfaces, and therefore calculation for two separate events, were only performed in those areas where high-resolution data were available. As described in Chapter 3 of this work, accuracy of MBES data ranges from 1 to 10% with a resolution better than 0.5 m. This allowed for the detailed identification of channel benches and channel floor, allowing for the calculation of channel bankfull geometry. Dimension of average flood channels in the study area are two to three orders of magnitude higher than data resolution (>100 - 1000 m of width, >10 m of depth).

The channel parameters needed for the calculation of the peak discharge were measured by drawing perpendicular profiles across the interpreted flood channels, obtaining a cross-channel section from which the main channel parameters were calculated. In particular, the required parameters were channel bankfull width and depth, for the cross sectional area, and slope of the channel. Depth and width were measured considering the bankfull surface of the channel (see Fig. 6.34 and Section 2.2.5). Bankfull depth represents the average vertical distance between the channel bed and the estimated water surface elevation required to completely fill the channel. Bankfull width is the measurement of the lateral extent of the water surface elevation perpendicular to the channel at bankfull depth. Where two events were recognised, width and depth were measured using channel benches as upper limits for the second event.

Different values of channel parameters were tested on channel Profile a, located across the Lobourg Channel (Fig. 6.35) in the area where high-resolution data is available. Variations in channel w (7000 ± 500 m) and d (20 ± 1 m) consistent with the resolution of the data (0.5-1 m) would result in a variation of discharge of less than 0.05 Sv (Table 6.5) .

Channel slope

Measurement of channel slope is another factor that has to be taken into account when analysing the accuracy palaeo-hydraulic calculations. In the case of the study area, measurement of channel slope was also affected by post-flood glacio-isostatic movements. These are related to the presence of the ice-sheet in the North Sea and to its retreat during deglaciation as the study area underwent post-glacial subsidence as it was located in the elevated ice-sheet forebulge area. The extent of glacio-isostatic movements have been quantified as 5-10 m after last Pleniglacial, assuming similar behaviour for Saalian and Anglian glaciations (see Chapter 2 for further details). These crustal movements have certainly had an effect on the slope of the channels, especially in the case of the upstream area (Lobourg Channel area), located in proximity of the ice-sheet margin during Pleistocene glaciations. This is confirmed by measurements of negative slope across the Lobourg Channel in some areas.

Another factor that affects measurement of the slope is resolution of data. Slope can vary significantly along the channel profile, within a very limited distance. If this distance is not resolved by low-resolution data, changes in slope value will not be taken into account. As a result, only one profile was calculated in those areas where resolution was lower as average resolution of 5-10 m and identification of two different surfaces would have therefore been not accurate and within error range, assuming channel depth of 10 m.

Where two erosional events and therefore two channel sections were identified, the slope value of the first event was measured on the bench, while the slope for the second event was measured on the channel floor (Fig. 6.34). Two main issues affect the measurement of the slope: the local variation of the slope and the variation of the slope with respect to the time of the flood.

A mean value was used in both formulas for each channel profile. The mean value was identified through the measurement of the slope on multiple transects (10 to 20 transect for each profile, depending on the width of the channel floor/ bench on which they were measured), drawn perpendicular to the channel profile on the channel floor, and defining a mean value. In order to obtain a representative value, outliers were not included in the measurements: only slope value between 0.009 and 0.0001 were considered.

Different slope values were tested on Profile a, where slope values ranging from 0.0001 to 0.0007 were measured (mean value is 0.0003). These were tested using fixed channel width and depth and n value, obtaining discharge in the range of 0.2 to 0.7 Sv (Table 6.5). This result shows that same order of magnitude of the calculated discharge was observed when applying values within the range of measured slopes.

Tested parameter	Tested value	Ref. value	% Error	Result	Ref. result	% difference
n	0.04	0.04	0.0	0.43	0.43	0.0
n	0.03	0.04	25.0	0.58	0.43	34.9
n	0.05	0.04	25.0	0.35	0.43	18.6
w	7500	7000	7.1	0.48	0.43	11.6
d	21	20	5.0	0.44	0.43	2.3
s	0.0001	0.0003	66.7	0.25	0.43	41.9
s	0.0003	0.0003	0.0	0.43	0.43	0.0
s	0.0005	0.0003	66.7	0.56	0.43	30.2
s	0.0007	0.0003	133.3	0.66	0.43	53.5

Table 6.5: Summary of the test performed on peak discharge variations using different parameters. Reference values, which are the values that were employed in the calculations, are in grey in the table.

Peak discharge at the Dover Strait

Assuming that the Lobourg Channel is interpreted to be the product of the erosion of a second megaflood event that followed the formation of the bedrock platform in which it is carved, the peak discharge was calculated using a single-event channel geometry. The elevation of the bedrock platform was used as the elevation of the channel western margin, while the eastern margin of the channel is not well preserved. The channel margin is in fact represented by bedrock remnants that are preserved in some tracts of the channel eastern margin: the profiles were drawn in selected areas where it was possible to measure the elevation of the remnants and, therefore, the depth of the channel. This was measured from the channel floor to its upper margin, represented by the platform and the remnants, whose elevation is assumed as the channel high-water level.

Table 6.6 shows the results of the peak discharge calculations performed on six profiles across the Lobourg Channel, whose location is shown in Fig. 6.36. The profiles were selected as the peak discharge calculated with Manning's and Darcy-Weisbach formulae have the same value. As the measurement of the slope (s) value was not possible in some areas, a range of values was tested for all the profiles using a series of values measured on the adjacent platform, with the most common value on the Lobourg Channel floor being 0.0003. Table 6.6 shows an example for Profile a, where three values of s measured along the channel section were used, obtaining the same order of magnitude for the calculated peak discharge that ranges from 0.4 to 0.8 Sv.

Profile	s	d/R (m)	W (m)	Q (Sv)
(a)	0.0003	20	7000	0.4
	0.0001			0.2
	0.0007			0.7
(b)	0.0003	18	10000	0.5
(c)	0.0003	22	9000	0.7
(d)	0.0003	25	8500	0.8
(e)	0.0003	25	5500	0.5
(f)	0.0003	20	13000	0.8

Table 6.6: Summary of the peak discharge obtained from the palaeo-hydraulic calculations.

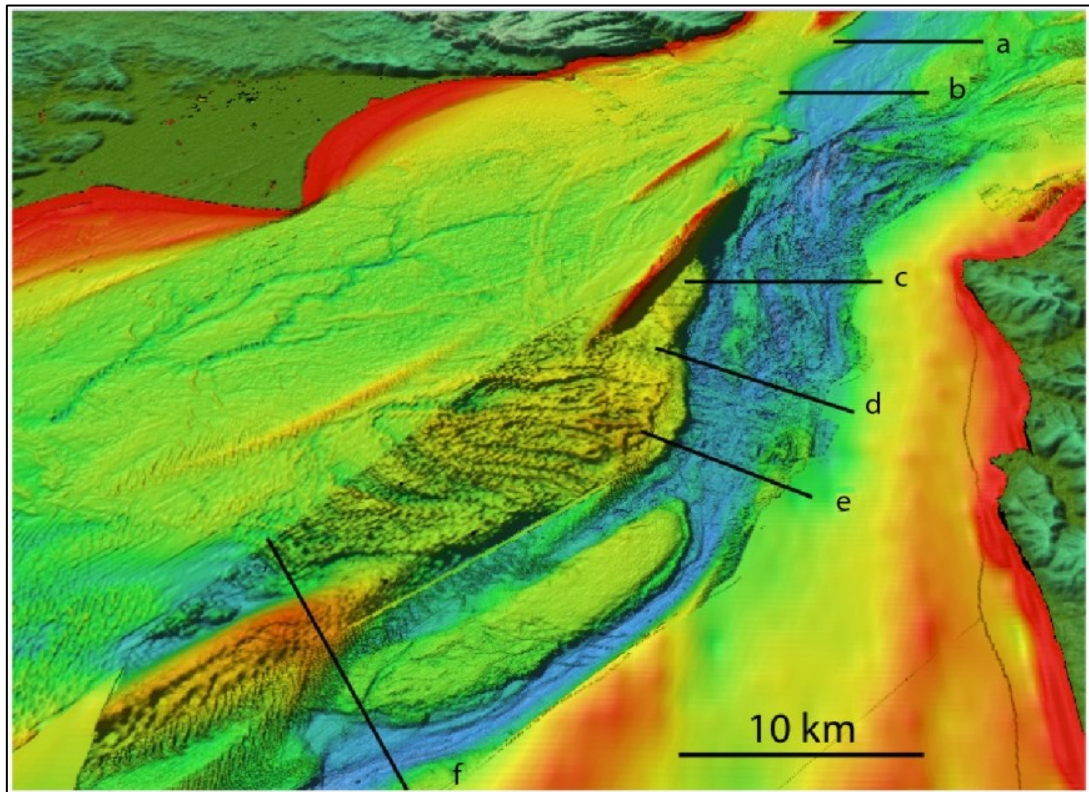


Figure 6.35: Location of cross-sectional profiles drawn across the Lobourg Channel. Colour-scale as in Fig. 6.20.

Peak discharge at the Northern Palaeovalley

The palaeo-hydraulic reconstructions across the palaeovalley were performed through the identification of two distinct erosional events, as indicated by the cross-cutting relationship of surfaces S5 and S6 interpreted in section 6.2.

It was therefore necessary to identify those channel profiles on which it was possible to identify two distinct surfaces for the two events; identifying two channel sections using the channel floor and the channel inner and outer benches (see Fig. 6.34). A set of profiles was drawn across the palaeovalley (Figs. 6.35-36): as for the previous sections, Table 6.7 shows those profile on which the calculated peak discharge have the same value using Manning's and Darcy equations.

For the first erosional event it was possible to calculate the discharge from profiles A and B, whose location is shown by the red lines in Fig. 6.36. The depth of the channel was measured using the inner bench S5 (as base of the channel) and the outer bench S2 (as channel margin). The obtained peak discharge ranges between 0.8 to 1 Sv, confirming that

the event that first carved the palaeovalley is two or more orders of magnitude bigger than typical fluvial flows. This calculations are also consistent with the calculated discharge in Gupta et al. (2007) that indicate a discharge of 1 Sv on a profile drawn in the same area using $s = 0.0002$, $n = 0.040$, $W = 20000$ and $d = 20$.

For the second event, five profiles, whose location is indicated in Fig. 6.36 by the black lines, were drawn using the palaeovalley floor (S6) as the channel base and the channel inner bench (S5) as the channel flank. The peak discharge value obtained with both formulas is shown in Table 6.8 and ranges from 0.3 to 0.7 Sv. Assumed that the second event identified in the palaeovalley corresponds to the second event at the Dover Strait, responsible for the formation of the Lobourg Channel, the comparison between the peak discharge obtained at the Dover Strait (0.4 to 0.8 Sv) is consistent with the 0.3 to 0.7 Sv obtained for the second event further downstream.

The obtained results are consistent with previous calculations made by Gupta et al. (2007) in the area of profiles 3 and 4 for which a discharge of 0.2 to 0.4 Sv was obtained for the second event, using values of $s = 0.0002$, $n = 0.040$, $W = 7000$ -12000 and $d = 15$.

Profile	s	d/R (m)	W (m)	n	Q (Sv)
A	0.0003	18	20000	0.040	1
B	0.0003	16	18000	0.040	0.8

Table 6.7: Result of the peak discharge calculations made on the two profiles where it was possible to identify two different events.

Profile	S	d/R (m)	W (m)	n	Q (Sv)
1	0.001	20	8000	0.045	0.7
2	0.0003	12	13000	0.040	0.3
3	0.0005	12	8500	0.040	0.3
4	0.0003	8	7000	0.040	0.3
5	0.0003	25	5500	0.040	0.3

Table 6.8: Summary of the peak discharge calculations for the second event.

Table 6.9 indicates the average peak discharge of some of the major rivers on Earth: the calculated discharge for the Northern Palaeovalley is at least two orders of magnitude bigger. This comparison represents a further indication of the exceptional high-magnitude flows that carved the Northern Palaeovalley: fluvial processes could not be responsible for the formation of a network of anastomosed channels carved into the bedrock for hundreds of kilometres and characterised by palaeo-flow peak discharge of the order of $10^6 \text{ m}^3/\text{s}$.

River	Average Discharge
Amazon	0.1 Sv
Ganges-Brahmaputra	0.04 Sv
Congo	0.04 Sv
Mississippi	0.01 Sv
Danube	0.007 Sv

Table 6.9: Peak discharge of some of Earth major modern rivers (source Gupta, 2007)

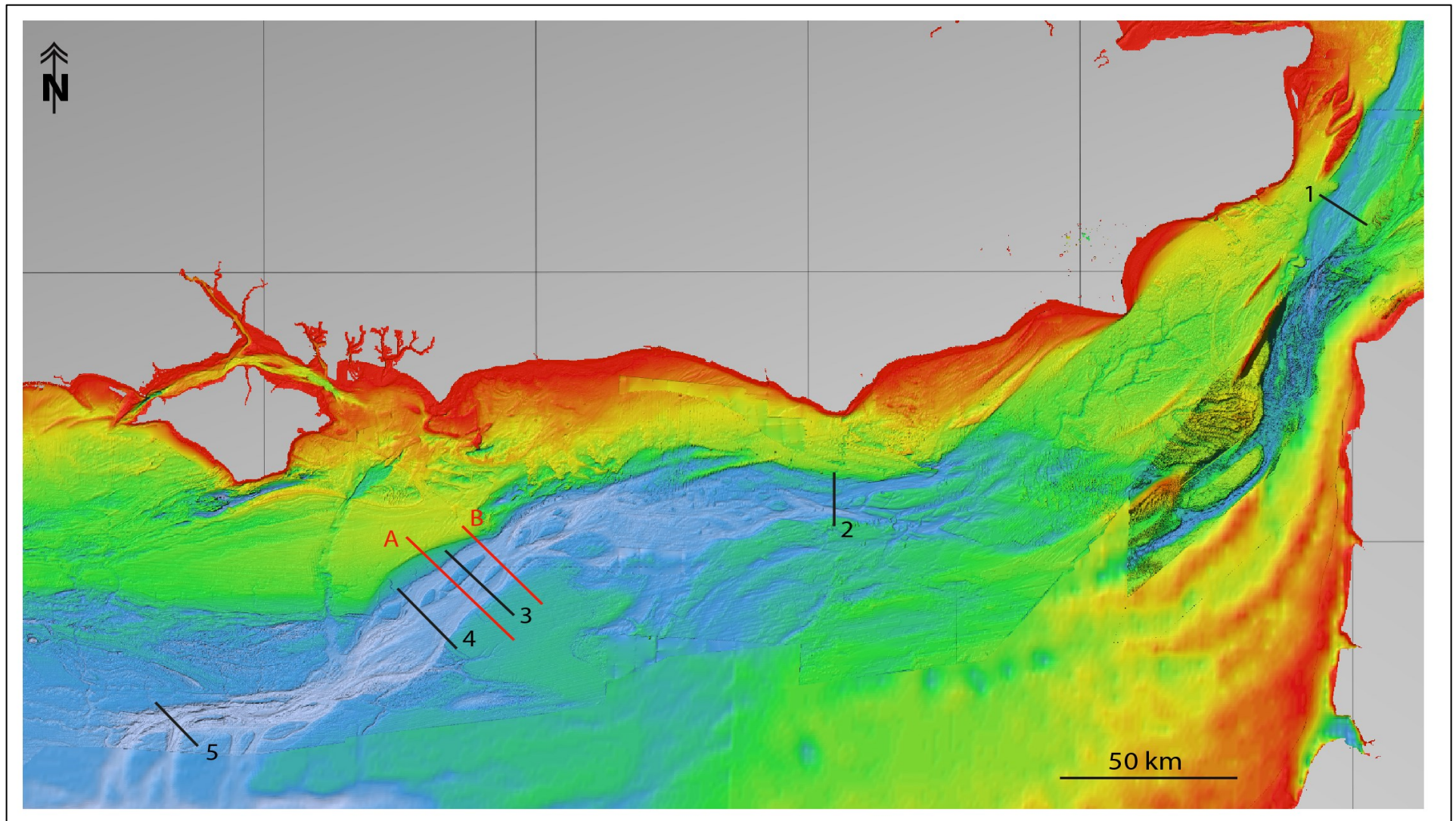


Figure 6.36: Location of cross-sectional profiles used for the calculation of the peak discharge across the Northern Palaeovalley. Colour-scale as in Fig. 6.20

6.6 Discussion

6.6.1 Association of flood features and comparison to other megaflood terrains

The mapping of the erosional features present in the Northern Palaeovalley channel network and described in this chapter indicates that the English Channel seabed shows the presence of a number of typical erosional flood-bedforms, displaying great similarity in geometry and dimensions to megaflood features interpreted in literature. These are represented by kilometre-scale channels (Carling et al., 2009b) displaying inner benches and hanging valleys (Baker and Nummedal, 1978) on their flanks. On the channel floor, streamlined islands (Bretz, 1923; Baker, 1978; Komar, 1984; Rudoy, 2002) and longitudinal lineations (Baker, 1978a) are eroded into bedrock, together with cataracts (Bretz, 1925; Baker and Nummedal, 1978; Lamb et al., 2008) and plunge pools (Rudoy, 2002; Lamb et al., 2010; Meinsen et al., 2011).

Table 6.10 shows a comparison between the typical flood features interpreted in the major megaflood terrains on Earth and the features present on the English Channel seabed. Firstly, the table shows that the interpreted features form an association of bedforms that shows great similarity to those interpreted in other megaflood terrains on Earth, providing further evidence of the flood-related origin of these features. In particular, anastomosed and bifurcating geometry is observed in the kilometre-scale flood channels carved into bedrock. Bifurcation is rarely observed in bedrock channels (Bridge, 2003) that usually only display anastomosis and bifurcation for short distances (Kleinhans et al., 2008), generally controlled by tectonic structures (Carling, 2009; Meshkova and Carling, 2012). In the case of the megaflood channels, anastomosis (Bretz, 1923) and channel bifurcation of bedrock flood channels can be observed for tens of kilometres (Baker, 2009b, 2009c), as in the case of the Northern Palaeovalley where clusters of streamlined islands and bedrock remnants divide channel branches carved into Cretaceous and Palaeogene rocks. Anastomosis is interpreted to occur as the result of immense flood discharge: the volume of water involved in these processes cannot be accommodated by pre-flood channels through which the flood flows from the source to the Ocean. As a result, the pre-flood river network is eroded by the flood into flood channels by divide breaching, forming a complex network of dividing and re-joining channel-ways (Baker, 1978b).

Table 6.10 also shows that the bedforms that compose the flood-feature association observed in the palaeovalley channel network display dimensions of the same order of

magnitude of typical flood features associated in other megaflood terrains. In addition, other than showing remarkably similar dimensions, the interpreted flood-feature association displays same spatial distribution of features and geometrical pattern as -other terrestrial megaflood terrains. An example is represented by the association of longitudinal lineations, cataracts and plunge pools located in the Lobourg Channel floor that follows the typical spatial distribution observed in the Channeled Scablands (Baker, 2009a) and the Box Canyon (Lamb et al., 2008) where lineations are carved at the head of cataracts and plunge pools are present at its base (see Section 6.4). In addition, streamlined islands carved into the palaeovalley floor are observed in typical clusters of islands, divided by inner channels and divide crossings, as recognised for the Channeled Scablands streamlined islands (Baker, 1978b, 2009c).

	English Channel (this work)	Channeled Scabland (Baker, 1978b, 2009a)	Box Canyon (Lamb et al., 2008)	Altai Mountains (Rudoy, 2002)	NW Germany (Meinsen et al., 2011)
Flood channels (width/length/depth)	20/170/30	8/80/250	0.1/2/35	na/10/50	4/na/30
Hanging valleys (elevation)	5-15	50-100	7	--	--
Gravel bars (amplitude)	--	20	--	120	--
Streamlined islands (length/amplitude)	10/50	9/15	--	--	4/na
Longitudinal lineations (length/amplitude)	4/1	10/50	--	--	--
Retreat cataracts (length/amplitude)	6/120	4/20	na/35	--	--
Plunge pools (width/depth)	4/80	1/35	0.1/2	1.2/70	0.8/35

Table 6.10: Summary of the typical megaflood feature association observed and interpreted in this work and in some of the major megaflood terrains on Earth.

A further example of the observed features displaying typical geometrical distribution is represented by the cross-cutting relationship displayed by the main flood channels (represented by the Lobourg Channel upstream and by the palaeovalley in the Central and

Channel Basin area) cutting hanging valleys (represented by channel B and by the Palaeo-Solent) that lie tens of metres above the channel floor, as observed by Baker (1978b, 29909a) and Lamb et al. (2008) for the Channeled Scabland and Box Canyon floods respectively.

Finally, in opposition to the typical flood-feature associations detailed in literature, where flood gravel and boulder bars, current mega-ripples and meter-sized boulders are observed (e.g. Baker, 1978a; Rudoy, 2002; Herget, 2005; Lamb et al., 2008; Carling et al., 2009b) no depositional flood features have been identified in the English Channel terrain. The only flood-related sedimentary feature observed in the study area consists of the bedrock incision infill (see Chapter 5); all the analysed flood-features present in the study area are erosional bedforms carved into bedrock, including the streamlined islands, as shown by both bathymetry and seismic data. The absence of depositional flood features does not appear to be controlled by geology as the whole study area shows lack of depositional flood-bedforms. This could be due to either very limited or no depositional bedforms to have formed during the erosion of the palaeovalley network, or to later remobilisation of sediments. Although the first hypothesis would be explained by reduced sediment availability and supply during the flood (e.g. Magilligan et al., 1998), a later remobilisation of sediments seems more likely in the case of the English Channel terrain. As detailed in Sections 2.3 and 4.2 of this work, the study area is dominated by strong tidal currents and seabed shear stress that could have the potential to mobilize and transport the sedimentary bedforms formed during the flood. In addition to these factors, stripping of sediments during transgression may have eroded and mobilised flood depositional features. Lag deposit (Hamblin, 1992) is present throughout the study area.

Megaflood peak discharge

The calculation of the estimated peak discharge (Section 6.5) for the two erosional events associated with the erosion of surfaces S5 and S3 (first event) and S6 (second event), resulted in a peak discharge of respectively 0.8-1 and 0.3-0.8 Sv. These results are consistent with typical megaflood peak discharge values calculated for major megaflood events on Earth, as illustrated in Table 6.11. In particular, the calculated peak discharge for the first event shows the same order of magnitude as the Lake Agassiz (Clarke et al., 2004) and NW Germany (Meinsen et al., 2011) megafloods, while the calculated discharge for the second event has the same order of magnitude as Lake Bonneville flood (O'Connor, 1993).

	English Channel	Channeled Scabland	Lake Bonneville	Lake Agassiz	Big Lost River	Altai Mountains	NW Germany
Peak discharge (Sv)	1	13	0.6	5	0.06	18/9	2
Reference	This work	O'Connor and Baker, 1992	O'Connor, 1993	Clarke et al., 2004	Rathbourn, 1993	Baker et al., 1993/ Carling et al., 2010	Meinsen et al., 2011

Table 6.11: Summary of the calculated peak discharge for the major megaflood terrains on Earth, showing similar magnitude to the calculated discharge for the English Channel megaflood. Two different values are given for Altai Mountain terrain, according to the two references indicated in the table.

6.6.2 Influence of the bedrock geology on flood bedforms

The change in the acoustic texture of the seabed, together with the morphological expression of the structural elements visible from bathymetry, made it possible to investigate the influence of the bedrock lithology and tectonic structures on the geometry of megaflood features. Morphological analyses of the mapped bedforms suggest a strong control of bedrock geology on the morphology and spatial distribution of flood features. These are in fact strongly influenced by variations in bedrock differential resistance to erosion, due to changes in lithology.

In the case of megaflood channels (Section 6.2), variations in lithology of the seabed have a strong influence on the channel gradient and slope, determining a change in the channel geometry and, consequently, on the hydrology of the flood. In particular, two main changes related to lithology are observed from bathymetry. Firstly, the Lobourg Channel has almost zero relief in the Southern North Sea area dominated by Palaeogene rocks, changing into a 25 m deep channel in the area where it is carved into the Chalk Group. This may be explained either as presence of sediments filling the valley or lower erosional power (or different erosional pattern) of the flows that carved the Lobourg Channel in the most upstream area. A further change in the geometry of the channel is then observed at the transition between Middle and Lower Chalk. Secondly, further downstream, the Northern Palaeovalley displays a marked transition in its geometry in the Hampshire Basin area (see Fig. 6.1 and Fig. 6.12). Here, the palaeovalley changes from low-gradient, wider and shallower channel geometry to deeper, bifurcating geometry where 20 m-deep inner channels are carved into the palaeovalley floor. In addition, the outcropping lithology also seems to control the preservation of channel margin and inner benches, as no bench is preserved in the

tract of Northern Palaeovalley that enters the Central Channel area in the Hampshire Basin. Conversely, well defined benches are observed in the tract of the palaeovalley located in the Channel Basin, where Cretaceous rocks outcrop. In conclusion, observations suggest that the channel geometry and palaeo-flow path are driven by lithology, rather than tectonic features. Structural control has nevertheless a secondary influence on the channel geometry and path, as the position and orientation and of the observed lithological boundaries that drive the channel path is controlled by tectonic features, as in the case of the bedrock platform bounded by the WNW-ESE trending fault.

The analyses performed on streamlined islands (Section 6.3) show a strong contrast in the geometry of the islands carved into the Palaeogene beds, located in the Hampshire Basin (set B), respect to the islands carved into Cretaceous rocks of the Wealden and Channel Basins (sets A and C). Firstly, the geometry of the islands varies with the outcropping geometry, showing highly eroded islands in the Hampshire basin and better preserved islands in the area where more resistant Cretaceous rocks outcrop. Secondly, as Palaeogene rocks display lower resistance to erosion due to their heterogeneous composition, a smaller number of islands is preserved in the Hampshire Basin. Among these, only the islands that reached equilibrium geometry (l/w and k value between 3 and 4) seem to have been preserved from erosion through time. Where more competent rocks outcrop at the seabed, a higher number of islands are observed, showing a wider range of geometries. This could be due to higher resistance to erosion of the outcropping bedrock, allowing for the preservation of those islands with more elongated geometry that have been eroded in the Hampshire Basin.

The erosional longitudinal megalineations (Section 6.4) are another example of how lithology can control the distribution of flood features through changes in erosional pattern. This is indicated by the longitudinal lineations only being present in the area where the Middle Chalk Formation outcrops. In the Lower Chalk Formation, lineations are truncated and replaced by different erosional bedforms just downstream of the escarpment present at the boundary between Middle and Lower Chalk. South of the escarpment, bedrock remnants are patchily carved into Lower Chalk. This sharp change in the erosional style is likely to be due to a change in the character of the seabed in correspondence of the lithological boundary, moving from the smoother Middle Chalk where longitudinal vorticity created erosional grooves by plucking action (Einstein and Li, 1958; Baker and Kochel, 1978) to different

erosional pattern observed in the Lower Chalk, probably dominated by erosion by abrasion rather than plucking.

6.6.3 Relative timing of erosion of the Northern Palaeovalley flood channels

The mapped erosional surfaces show cross-cutting relationships that indicate that the palaeo-channel network was carved by multiple erosional events. The interpretation of the cross-cutting relationships between surfaces suggests the presence of at least two erosional flood events (Figs. 6.37-38). The first event is interpreted as responsible for the formation of the bedrock platform, together with the palaeovalley bench further downstream. A second erosional event would have carved the Lobourg Channel into the platform, on the Eastern sector of the Strait, forming the giant streamlined island located into the channel floor (Fig. 6.37). In the downstream area, the streamlined islands were carved into surface S5 during the second event that carved the modern palaeovalley floor (Fig. 6.38).

The Lobourg Channel is interpreted as the result of a later event respect to the bedrock platform, as indicated by erosional, cross-cutting relationships observed on the data. Firstly, the cross-cutting relationship between the channel and the bedrock platform indicates that the hanging valley Channel B was truncated by the Lobourg Channel entering the Northern Palaeovalley at a later stage with respect to the erosion of the bedrock platform. In addition, distinct sets of longitudinal megalineations were observed on the Lobourg Channel floor (sets A, B, D, E), all displaying the same orientation. A further two sets of lineations with different orientation were mapped on the platform (C) and on a bedrock remnant (F) elevated from the channel floor, displaying different orientation.

The observed lineations are interpreted as the result of vortices in the flow (Einstein and Li, 1958; Baker et al., 1979) that carved the channel. It is not possible to determine if the lineations were formed by a single flow (Baker and Nummedal, 1988), but the almost identical orientation of the lineations present on the Lobourg Channel floor suggests that they were carved by flows with the same orientation, possibly during the same flood event. Their formation could have occurred in more pulses during a very restricted period of time, typically days to weeks (e.g. Baker, 1978a, 2009c). Conversely, the different orientation of the lineations located on the bedrock platform (oriented ENE (040°) against ENE (060°) of the lineations on the channel floor) suggests that they were carved by a flow with different orientation respect to the lineations on the channel floor. This is also supported by the

difference in elevation (25 m) between the channel margin and the channel floor, on which the different sets are located.

Evidence of multiple erosional events to have carved the Northern Palaeovalley are also present in the downstream area (Fig. 6.38) where a second example of hanging valley, other than channel B, is represented by the former channel of the Palaeo-Solent River. This is truncated by two erosional surfaces that are now the Palaeovalley bench and floor, truncating the Palaeo-Solent Channel for a second time. The streamlined islands located downstream also show the same elevation as the channel bench, suggesting that they were carved into the bench, which probably represented the palaeovalley floor after the first event. The presence of two sets of small scale drainage channels f3 and f4 post-date the two flood events.

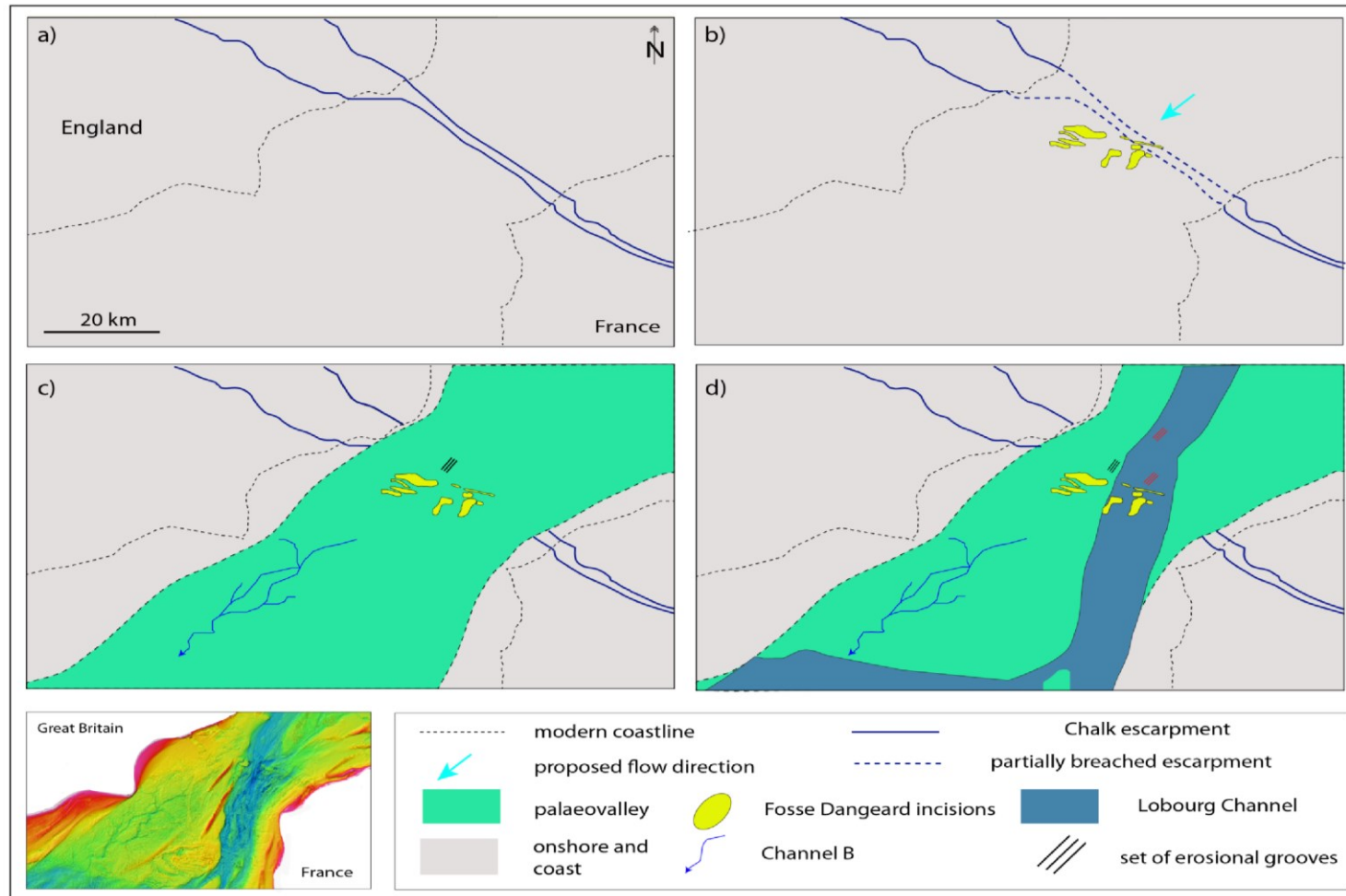


Figure 6.37: Model for the erosion of the Northern Palaeovalley and Lobourg Channel in the Dover Strait area displaying the four main stages of the formation of bedrock and Lobourg Channel. Box a represents the pre-flood Dover Strait scenario, with Chalk escarpment partially breached in box b, together with the formation of the Fosse Dangeard incisions. Box c shows the erosion of the platform (and longitudinal lineations) and post-flood drainage Channel B. Another flood episode is represented by box d with erosion of Lobourg Channel (and sets of longitudinal lineations carved on the channel floor). Channel B is truncated by the flood.

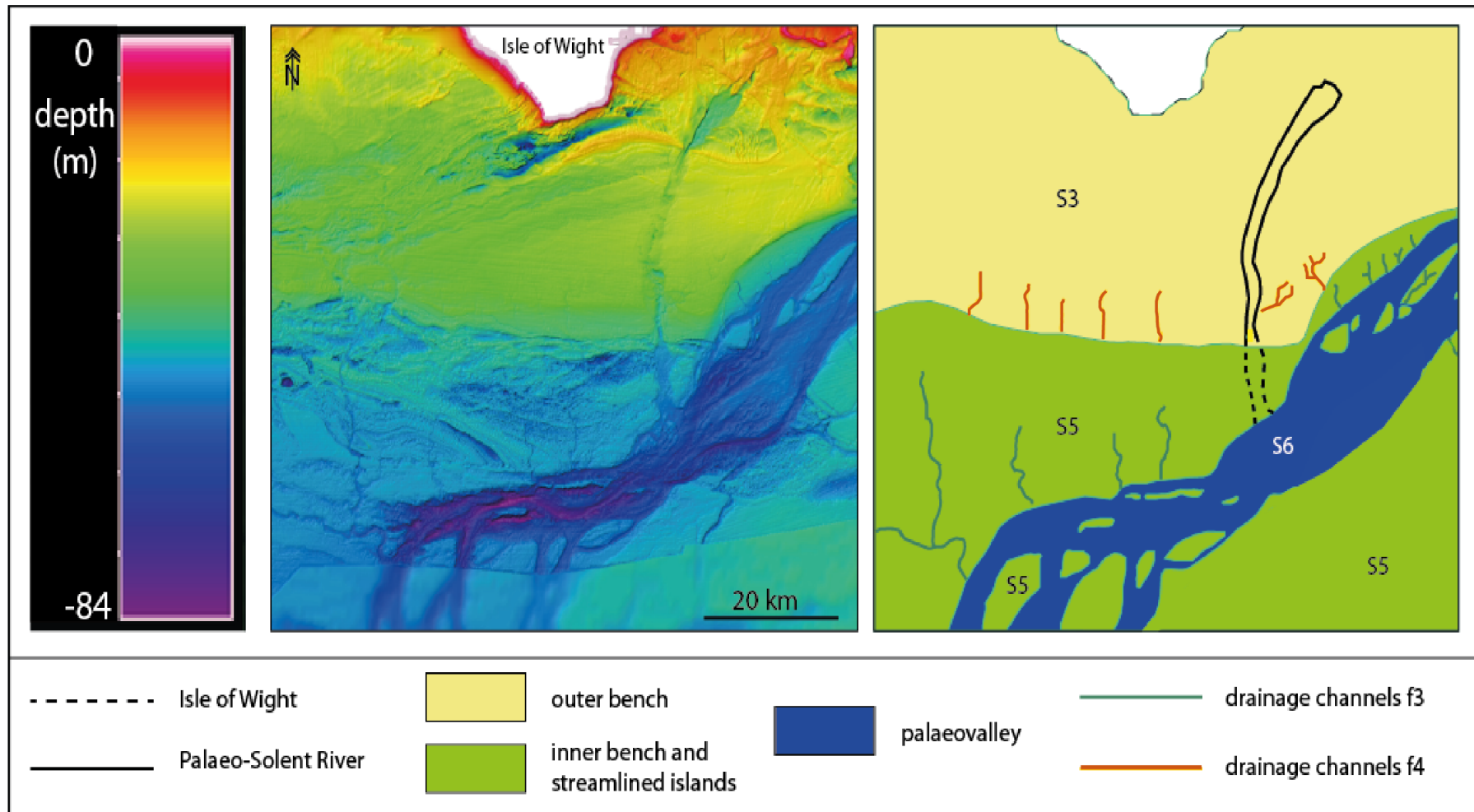


Figure 6.38: Bathymetry (left) and line drawing (right) of the downstream sector of the Northern Palaeovalley. The different colours indicate different erosional surfaces and times of erosion (see text).

6.7 Conclusions

Geomorphologic analysis and interpretation of landforms mapped from bathymetry data in the Northern Palaeovalley enable four main conclusions to be drawn:

- The bedforms mapped in the study area form an assemblage of features typical of flood-eroded landscapes, showing very strong similarity in dimensions and spatial distribution to flood features observed in major terrestrial megaflood terrains, for example the Channeled Scablands (Baker et al., 2009a; Carling et al., 2009b), Box Canyon (Lamb et al., 2008), Altai Mountains (Herget, 2005) or the Munsterland Embayment (Meinsen et al., 2011). Differently from typical flood-landform assemblage observed in literature no depositional flood features has been preserved in the English Channel terrain, probably due to later sediment remobilisation by tidal action.
- The geometry and spatial distribution of the flood features is strongly influenced by the bedrock geology; in particular, changes in lithology of the seabed determine different levels of resistance to erosion of the rocks into which the bedforms are carved. A clear example of the control of seabed geology is observed on the distribution and different levels of preservation of the mapped streamlined islands. The effect of outcropping bedrock geology is also visible from the change in erosional pattern observed at the Middle-Lower Chalk boundary, downstream of which the observed longitudinal lineations are replaced by patchily distributed remnants carved into the seafloor.
- The analyses of the flood erosional surfaces present in the study area and of their cross-cut relationships indicate that at least two erosional events carved the Northern Palaeovalley. The first erosional event is interpreted to be connected to the formation of the bedrock platform and Fosse Dangeard depressions, successively incised by a second event during which the Lobourg Channel and streamlined bedrock remnants were formed and the infilled incisions were truncated.

- The performed palaeo-hydraulic analyses indicate a peak discharge of up to 1 Sv for the first erosional event and up to 0.8 Sv for the second erosional event, associated to the formation of the Lobourg Channel. These values are consistent with typical megaflood peak discharge values, as indicated by the comparison to the main events described in literature.

7 . Final megaflood model and palaeo-geographic considerations

7.1 Introduction

Following the analyses and interpretation of Northern Palaeovalley channel network, a summary model of the flood events that carved the palaeovalley is developed here. This will be described and discussed in this chapter, with the aim of reconciling the interpreted model with the Middle and Late Pleistocene palaeo-geographic history of the study area. In particular, the following points will be presented: 1) the interpreted number and relative timing of the floods, allowing for the reconstruction of the relative chronology of the formation of the interpreted flood features and channel network (Figs. 7.1-2); 2) Palaeo-geographic analysis of the proposed flood model, discussing the formation of the mapped flood-features in relation to regional-scale events that affected the study area during Middle and Late Pleistocene; 3) Estimation of the flood erosion depth according to the interpreted flood model and proposed palaeo-geographic interpretation; 4) Final considerations on the possible triggers of the floods in relation to their source.

7.2 Relative timing of interpreted flood events

Following the performed analyses on seismic and bathymetry data, the interpretations proposed in Chapter 5 and 6 of this work indicate the flood history to be composed of (at least) two main erosional events that carved the palaeovalley. Figs. 7.1 and 7.2 summarise the proposed model for the formation of the Northern Palaeovalley channel network (Fig. 7.1) and Fosse Dangeard (Fig. 7.2), showing the relative timing of the interpreted flood events.

7.2.1 Erosion of the Palaeovalley network

The interpreted erosional events (Fig. 7.1) are indicated by cross-cutting relationships of the mapped erosional surfaces, both in the upstream and downstream sectors of the palaeovalley, as detailed in Chapter 6. For details and location of the interpreted flood features see Chapter 6 and map in Fig. 6.1.

In the upstream area, cross-cutting relationship between the bedrock platform and the Lobourg Channel (both flood surfaces) suggests that the surfaces were carved at two different times. The bedrock platform, which was carved during the first flood event (identified as event 1), possibly extended across the English and French sectors of the Strait after the first erosional event (stage C in Fig. 7.1), during which flood channel B also formed on the platform. The formation of the Lobourg Channel is interpreted to have occurred during a second flood event, identified as event 2 (Stages D1-D3), truncating both the platform and channel B and carving the giant streamlined island present on the Lobourg Channel floor into the existing flood surface. Further downstream the bathymetry also shows evidence of flood surfaces carved at, at least, two different times. Firstly, the inner bench of the palaeovalley is interpreted to have been carved during event 1 (Stage B2), followed by the erosion of the modern palaeovalley floor and of the streamlined islands into the older flood surface (stage E). As for the streamlined island located upstream in the Lobourg Channel floor, the observed teardrop-shaped islands present downstream are erosional remnants of the existing flood surface, successively carved by a second flow during event 2. In addition, the Palaeo-Solent River displays two knick-points: these are present in correspondence of the truncation of the channel by the two flood surfaces. This indicates that the palaeo-river was truncated a first time during the formation of the first flood surface (Stage B2), and at second time during the formation of the palaeovalley floor (Stage E).

Despite the lack of constraints on the absolute age of the erosion, the geometry of the interpreted channels suggests that a correlation between the erosional surfaces in the upstream and downstream area can be made, as shown in Fig. 7.1, where the model of the palaeovalley network erosion is presented. As the two main erosional events were interpreted to have occurred both in the upstream and downstream area, the bedrock platform in the upstream area is interpreted to have formed at the same time as the erosion of the palaeovalley further downstream, during event 1. Equally, for event 2, the Lobourg Channel would have carved the bedrock platform and entered the palaeovalley eroding the surface that is the modern channel floor in the downstream area.

In relation to the duration of the flood events, events 1 and 2 are likely to have had duration of weeks to months, as suggested by typical duration of megaflood flows associated with pro-glacial lake drainage events (see Section 2.2 of this work). Typically, each drainage events is composed of multiple erosional pulses, usually characterised by duration of less

than a week. In the case of the Northern palaeovalley network, as there is no stratigraphic evidence to constrain the absolute timing of the flows, in order to identify multiple pulses to have occurred during events 1 and 2, it was necessary to identify erosional elements and cross-cutting relationships that suggest the erosion by multiple flows during the same flood event. In the upstream area, where high-resolution bathymetry is available, the presence of inner benches in the Lobourg Channel (identified as erosional surface S4 in Chapter 6 and map in Fig. 6.1) may indicate multiple flows to have carved the Lobourg Channel during event 2. This is also supported by the presence of distinct sets of longitudinal megalineations carved on the Lobourg Channel floor that show variable orientations, indicating that the lineations could have been carved by flows with different orientation. As the Lobourg Channel represents the youngest of the flood surfaces in the interpreted relative chronology of flood events, the presence of multiple surfaces and sets of lineations carved on the channel floor would indicate that the observed features were all carved during event 2, but during distinct flood pulses. The hypothesis of multiple flows to have occurred during events 1 and 2 is also supported by erosional contacts identified in the Fosse Dangeard infill and detailed in the section below.

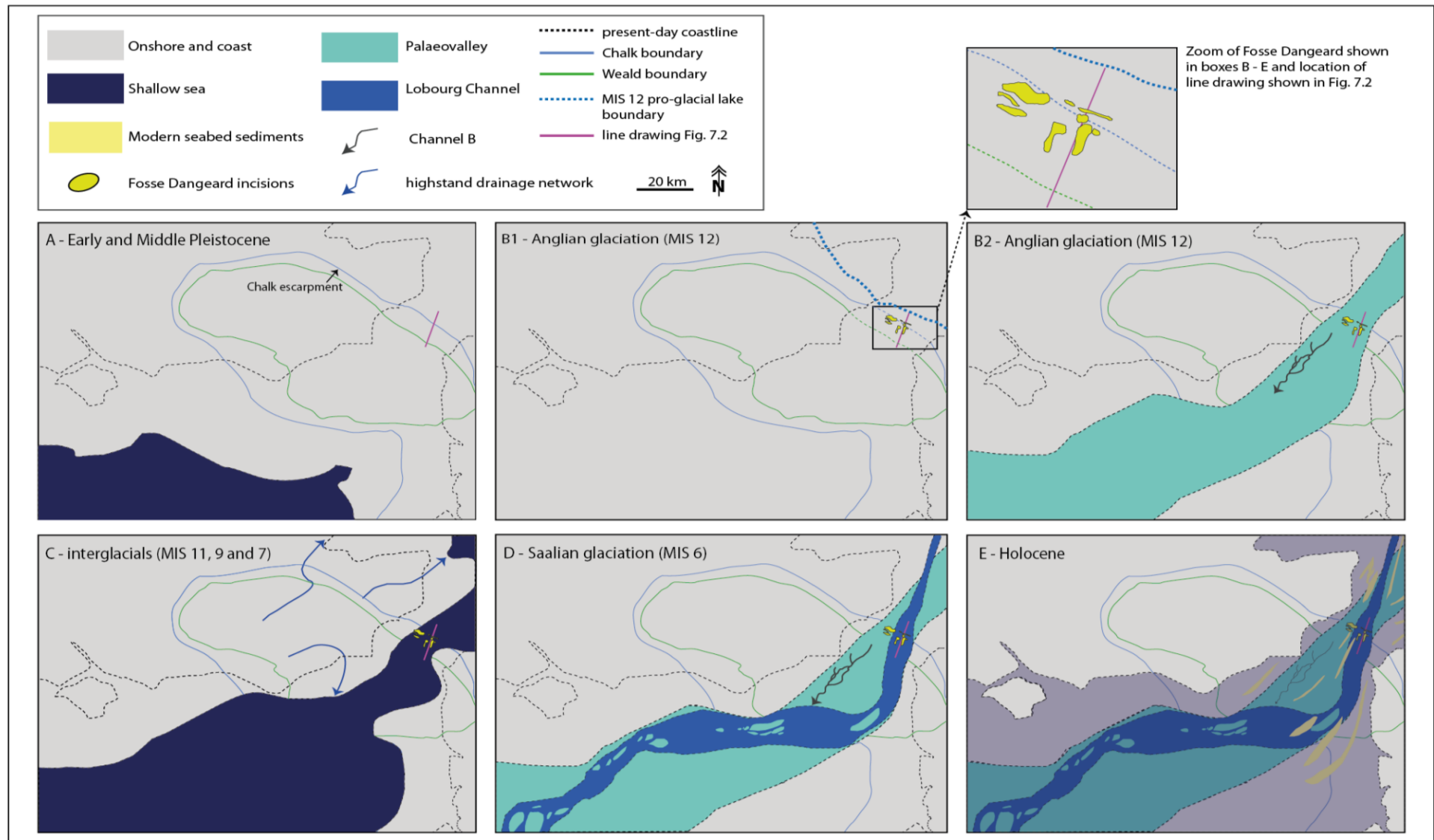


Figure 7.1 (previous page): Interpretative cartoon summarising the palaeo-geographic evolution of the study area during Pleistocene up to Holocene (in box E, where the area occupied by sea is partially transparent in order to show the Northern Palaeovalley and modern seabed sediments). Each box represents a different stage, starting from Early to Middle Pleistocene (A) when the Weald-Artois anticline represented a bedrock bridge, to the two phases of breaching of the Strait at MIS 12 and MIS 6. Blue semi-transparent colour in box E represents present-day submerged English Channel. The evolution of the Fosse Dangeard at the same time intervals is shown in Fig. 7.2 with a line drawing cartoon (whose location is represented by the pink solid line in this figure). Drainage network and palaeo-shorelines in boxes A and C from Gibbard, 1995; Cohen et al., 2005 and Hijma et al., 2012.

7.2.2 Formation of the Fosse Dangeard

The reflection seismic analysis and interpretation of the formation of the bedrock depressions located at the Dover Strait and detailed in Chapter 5 of this work is consistent with the hypothesis of (at least) two flood events to have carved the Northern Palaeovalley. A first high-energy erosional event is interpreted to have eroded the base of the incisions into bedrock, interpreted as Stage B1 in Fig. 7.2. As shown in the figure, the depressions were then carved for a second time during the erosion of the palaeovalley (Stage B2) and successively filled by sediments draping the basal erosional surface and filling the depressions, probably during interglacial times and/or intervals between flood pulses (Stage C). The described flood stages B to C would all have occurred during event 1, possibly representing erosion-deposition pulses of the same flood. Event 1 would therefore represent the flood flow that breached the Chalk escarpment for the first time (stage B1 in Figs. 7.1 and 7.2), eroding the pro-glacial lake bedrock dam and initiating the drainage of the lake. As described in Chapter 5, the presence of the escarpment at the Chalk ridge would have formed an elevation drop: this would explain the exceptionally high erosional power of the flow that breached the Chalk ridge for the first time, carving the depressions at the base of the escarpment.

A second major erosional event is suggested to have carved the sediments deposited into the depressions (stage D1), as indicated by the presence of a second erosional surface carved into the infill and visible in the incisions. As shown by Stages D1-D3 in Fig. 7.2, the depressions would have been carved and sediment-filled for a second time, with deposition of sediments probably occurring during intervals between flood pulses, and erosion truncating the sediments at the time of the formation of the Lobourg Channel (Stage D3).

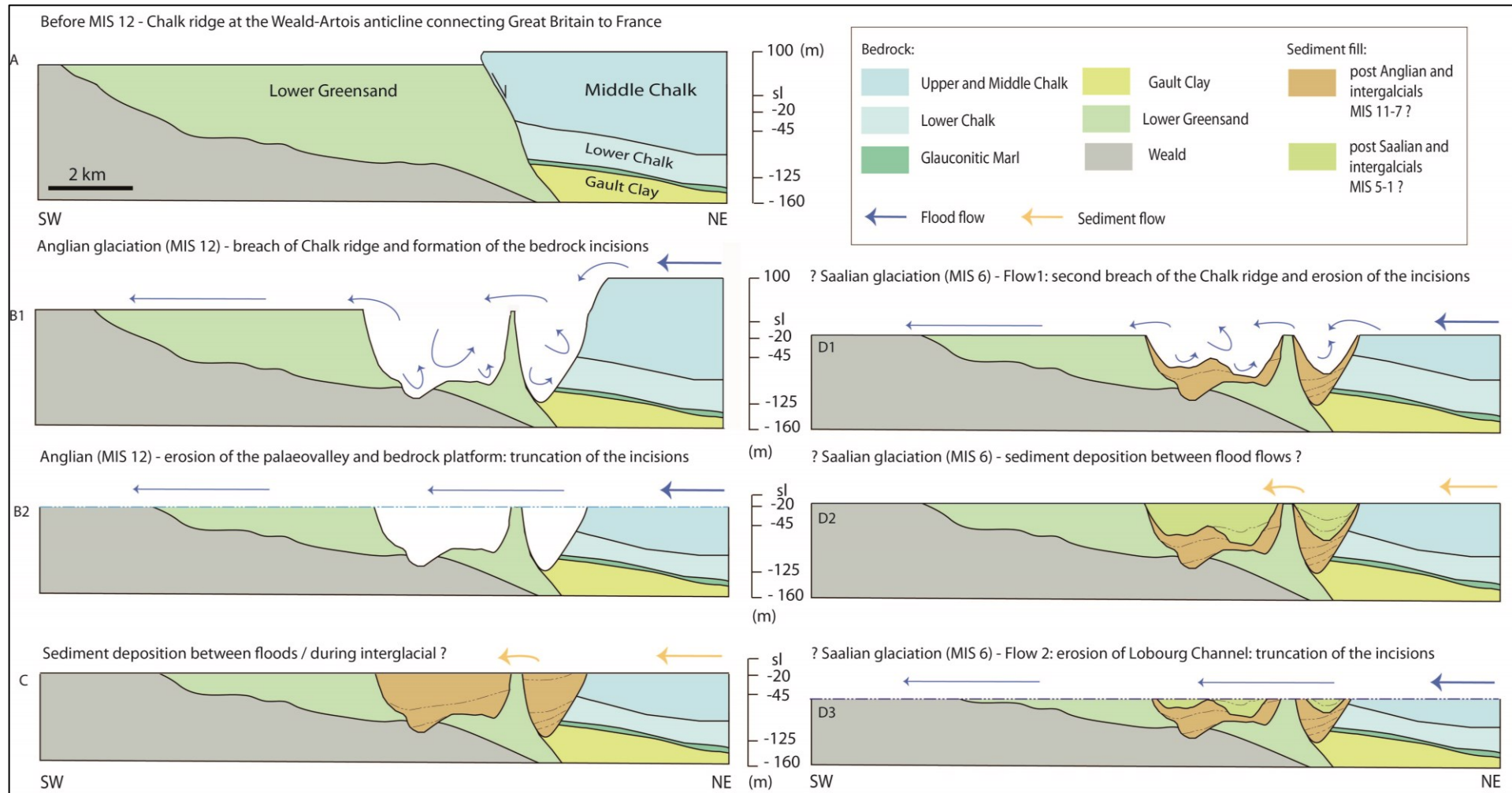


Figure 7.2: Model of the formation of the incisions of the Fosse Dangeard during Anglian and Saalian glaciation times. The line drawing shown in this figure is located in the Fosse Dangeard on the profile shown Fig. 7.1. For details on elevation and depths of the flood surfaces see text. sl represents the present day mean sea level elevation.

This series of erosion and deposition events of flood event 2 (Stages D1 to D3): this event was probably characterised by lower erosional power respect to event 1, as indicated by the presence of incisions carved exclusively into the infill of the existing features and not into bedrock. The lower erosional power of event 2 could be explained by a reduced elevation drop present at the partially eroded Chalk escarpment at the time of the second flood, generating lower acceleration of the flow and resulting in a more limited erosional power respect to event 1. This is also consistent with the source of event 2 being a second proglacial lake bounded by the partially breached Chalk escarpment. In this case, the lake water-depth would have been lower than during the first event, as the lake was bounded by the partially eroded ridge (see Table 7.2 for estimation of surface elevation).

The analyses of the surfaces on which the incisions are located, performed on bathymetry and detailed in Chapter 6, is also consistent with a formation of the depressions during event 1, as suggested by Figs. 7.1-2. This is indicated by the incisions lying on two different erosional surfaces represented by the bedrock platform, formed during the first breaching of the Strait (event 1), and by the Lobourg Channel floor, formed during event 2 into the existing bedrock platform. Although there is no constraint on the absolute timing, the Lobourg Channel appears to truncate the bedrock incisions and their infill, suggesting that the depressions were probably carved and infilled twice before the Lobourg Channel was eroded into the bedrock platform, as shown by Fig. 7.2.

In summary, the proposed model for the formation of the palaeovalley network and bedrock incisions can be described by stages A to E, as also summarised in table 7.1.

Stage A: Presence of a bedrock ridge at the Dover Strait, located on the Weald-Artois anticline which connected Britain to the continent. **Stage B1:** Formation of the Fosse Dangeard during the breaching of the bedrock dam at the Dover Strait, starting the drainage of the lake and initiating flood event 1. **Stage B2:** Erosion of the palaeovalley and formation of the bedrock platform across the Strait, breached for the first time. Formation of flood channel B on the bedrock platform. **Stage C:** Deposition of sediments into the Fosse Dangeard bedrock incisions during interglacial times. Submersion of the eroded palaeovalley and bedrock platform during high-stands, with the partially-breached Dover Strait intermittently connecting the Southern North Sea and the Atlantic Ocean during low-stands.

Stage D1: Second major flood event (event 2) started by the second breaching of the partially-eroded Chalk ridge, during which the Fosse Dangeard incisions were carved for the second time into the existing depressions filled by sediments. **Stage D2-D3:** Lobourg Channel and streamlined islands were carved into the existing flood surface (Stage E3) and into the depression infill deposited at Stage D2. **Stage E:** Submersion of the Northern palaeovalley channel network and tide-controlled deposition of modern seabed sediments (see Chapter 4 of this work for details).

Stage		Age (MIS)	Erosion/Deposition	Dover Strait Bedrock dam	S. North Sea – Eng. Channel seaway connection
A		Before MIS 12	Fluvial drainage network erosion/deposition	not breached	not connected
B1		Event 1 MIS 12	Formation of the Fosse Dangeard	first breaching event	not connected
B2			Formation of palaeovalley	partially breached	connected
C		Interglacials MIS 11-7	Deposition of infill sediments	partially breached	connected during low-stands
D	D1	Event 2 MIS 6	Erosion of the Fosse Dangeard	second breaching event	connected
	D2		Deposition of infill sediments	breached	
	D3		Formation of Lobourg Channel	breached	
E		MIS 5-1 Holocene	deposition of modern seabed sediments	breached	connected

Table 7.1: Summary of the interpreted stages for the formation of the palaeovalley network and Fosse Dangeard incisions from MIS 12 to Holocene.

7.3 Palaeo-geographic reconstruction

The proposed flood model summarised in Figs. 7.1-2 and Table 7.1 is consistent with existing previous studies on palaeo-geography (e.g. Belt, 1834; Ussher, 1913; Stamp, 1927, 1936; Gullentops and Huyghebaert, 1974; Roep et al. 1975; Jones, 1981; Smith, 1985; Ehlers and Rose, 1991; Gibbard, 2007), river drainage configuration (e.g. Gibbard, 1995; Busschers et al., 2008; Toucanne et al., 2009a, 2009b; Gibbard and Clark, 2011) and palaeontology (e.g.

Waitt, 1985; Gamble, 1987; Roebroeks et al., 1992; Roebroeks, 2001; Ashton and Lewis, 2002, 2012; White et al., 2006; Ashton and Hosfield, 2010) of north-western Europe during Middle and Late Pleistocene. This section describes how the proposed flood model fits into the study area palaeo-geographic framework.

7.3.1 Anglian glaciation

The first event proposed to have carved the palaeovalley and identified as event 1 is interpreted as the erosional event that occurred at MIS 12 during Anglian glaciation (~450 kyr before present) and previously interpreted by other authors as a catastrophic flow that eroded the English Channel seabed (Smith, 1985; Gibbard, 2007; Gupta et al., 2007; Toucanne et al., 2009a, 2009b). The megaflood flow would have been triggered by the dam failure of a pro-glacial lake located in the Southern North Sea (Figs. 7.1 and 7.3) and bounded to the north by the British and Scandinavian ice-sheets that merged for the first time during Anglian glaciation (Oele and Schuttenhelm, 1979; Gibbard and Clark, 2011). The southern margin of the lake was bounded by the steep north-eastern flank of the Weald-Artois anticline, composed of Cretaceous Chalk rocks. The presence of the ice sheet would have caused the north-western European river drainage network to divert, with the pro-glacial lake receiving run-off from the re-routed river system (e.g. Gibbard, 1995) that discharged an exceptionally high volume of water into the lake, also fed by ice-sheet melt water. This resulted in a catastrophic drainage of the pro-glacial lake through breaching of the bedrock dam: the Fosse Dangeard incisions represent the only evidence of the breaching of the bedrock dam that bounded the lake before the flood. Following the erosion of the palaeovalley, at the end of event 1, the bedrock platform was the emerged land-bridge that connected Britain to the continent during low-stands (Keen, 1995; Gibbard, 1995; Hijma et al., 2012), and a sea-way during high-stands. This is also supported by mollusca record showing presence of marine organisms on both sides of the Channel and therefore indicating intermittent marine environment in the Channel (Meijer and Preece, 1995; Meijer and Cleveringa, 2009). The intermittency of connection and isolation of Britain during this time is consistent with the hypothesis of a partial breaching of the Chalk. The presence of a land-bridge between Britain and France would probably have not been present if the ridge was completely eroded by the first flood event.

7.3.2 Saalian glaciation

Erosional event 2 has been interpreted to have been the high-magnitude erosional event that occurred in the Strait area (Gupta et al., 2007; Gibbard, 2007; Busschers et al., 2008; Meinsen et al., 2011) at the end of the Elsterian glaciation (~150 kyr before present, MIS 6). At this time, the British and Scandinavian ice-sheets likely coalesced for the second time, bounding another pro-glacial lake located in the Southern North Sea, as shown in Fig. 7.3 (Rappol, 1987; van der Berg and Beets, 1987; Ehlers 1990; Kluiving et al. 1991; Murton and Murton, 2012). Sedimentological data (Cameron et al., 1986; Joon et al., 1990; Laban, 1995; Cohen et al., 2005, 2011; Busschers et al., 2008) give evidence of the presence of the pro-glacial lake, whose southern margin would have been located further north of the location of the ice margin at MIS 12 (Fig. 7.3), as the ice-sheet probably did not advance as south as for the Anglian glaciation (Stokes and Clark, 2001). The pro-glacial lake would have been fed by the melting ice-sheet water and by the diverted north-west European river network (e.g. Cohen et al., 2005; Gibbard, 2007; Busschers et al., 2008; Hijma et al., 2012). The drainage of the lake would have produced a second megaflood flow, corresponding to the interpreted event 2, to breach the partially eroded Dover Strait. An alternative hypothesis for source of the megaflood has been proposed by Meinsen et al. (2011), suggesting that the water coming from the drainage of pro-glacial lake Weser in north-west Germany would have joined the Rhine-Meuse river system and entered the Dover Strait at MIS 6. Geochemical data from sediment cores collected at the mouth of the Channel also indicate the presence of an increased discharge of sediment through the Dover Strait and English Channel, interpreted to be related to the presence of the ice-sheet occupying the North Sea area at that time (Eynaud et al., 2007; Toucanne et al., 2009a, 2009b). Event 2 would therefore represent the definitive breaching of the Strait that, after MIS 6, would have been a seaway between Britain and France. The Chalk escarpment, partially breached by event 1, would have been completely eroded throughout the Strait, with the Lobourg Channel flowing from the Southern North Sea up to the eastern English Channel.

With respect to the palaeontological framework suggested by previous studies (see Section 2.4 of this work), a first decline in human presence in Southern Britain between MIS 11 and 8 (e.g. Ashton and Lewis, 2012) is suggested, possibly related to the isolation of Britain after the breaching of the Strait at MIS 12 and to the successive phases of connection/separation of Britain from the continent after event 1. The following erosional event, which led to the complete opening of the Strait, could be interpreted as the cause of the

disappearance of human presence in southern Britain between MIS 6 and MIS 4 suggested by palaeontological studies (Stuart, 1976; Curren, 1986; Wymer, 1988; Curren and Jacobi, 2001; Ashton, 2002; Ashton and Lewis, 2002).

7.4 Calculation of the flood erosion depth

Fig 7.4 shows two profiles drawn across the megaflood terrain, displaying the elevation of the interpreted flood surfaces and the relative erosion depth of the two flood events that carved the surfaces. Profile 1 (green) was drawn on the flood surfaces produced by event 1, while profile 2 (blue) shows the elevation of the Lobourg Channel and of the bedrock island present in the channel floor and formed during event 2. In order to estimate the erosion depth of events 1 and 2, the elevation of pro-glacial lakes at MIS 12 and 6 (representing the proposed source of the flows) and of the Weald-Artois anticline (representing the lake dam) are also plotted in the diagram, together with the depth of the erosional surfaces interpreted as the Fosse Dangeard incisions.

From the observation of the elevation profiles, other than calculating the erosion depths for events 1 and 2 (detailed in the sections below), some final observations on the main morphological features interpreted in this work can be made. In particular, the comparison between the two profiles shown in Fig. 7.4 confirms that the bedrock platform and the giant streamlined island lay at the same elevation, consistently with the hypothesis that the island was carved into the platform during the formation of the Lobourg Channel. In addition, Fig. 7.4 emphasizes the influence of bedrock geology, already detailed in Chapters 4 and 6, and clearly visible on the profiles that display a sharp change in geometry in correspondence of changes in lithology. In particular, a change is observed at the Middle-Lower Chalk escarpment where both profiles are characterised by a more irregular geometry respect to the smooth profile visible in the area eroded into the Upper and Middle Chalk. This is consistent with the observations made from both seismic and bathymetry data, displaying a marked change in the acoustic character of the seabed and in the geometry of the flood features in correspondence of lithological boundaries. Drainage channel f1 and flood channel B are also visible from on the profiles, stressing the difference in scale between flood-related features and typical fluvial channels.

7.4.1 Erosion depth of event 1

For event 1, the erosion depth can be calculated using the elevation of the lake (or of the Weald Artois anticline) and the depth of the Fosse Dangeard incisions: the lake represents the elevation of the source from which the flood was generated, while the base of the incisions represents the base of the erosion during the bedrock dam breach.

With regard to the position and elevation of the lake margin, the following assumptions were made: 1) the elevation of the lake was assumed to be 30 m, as proposed by Cohen et al. (2005), Busschers et al. (2008) and Hijma et al. (2012), based on gravel terrace deposits interpreted by Roep et al. (1975), although there is no other clear sedimentological data to support this interpretation. Considering a maximum depth of 80 m for the base of the Fosse Dangeard incisions (Fig. 7.3) and 45 m of depth of the seafloor from which the depth is measured, the total depth of the Fosse base is ~125 m below mean sea level (msl). This indicates that the erosion depth for the event at MIS 12, using lake elevation of 30 m above msl, corresponds to ~155 m. 2) Although Hijma et al. (2012) and Murton and Murton (2012) suggest the lake south-eastern margin to be located ~50 km north-east of the Strait (as shown in Fig. 7.3 by the pink line), the interpretation of the formation of the Fosse Dangeard depressions proposed in Chapter 5 of this work suggests the lake margin to have been located further south-east, as shown in Fig. 7.3 by the blue line. If the Fosse Dangeard incisions are interpreted as plunge-pools formed at the base of the palaeo-waterfall formed at the Chalk escarpment, the source of the flood (pro-glacial lake) must have been located immediately upstream of the bedrock ridge, forming the hydraulic jump that would have created the palaeo-waterfall at the base of which the incisions were carved. The location of the lake-margin proposed by Hijma et al. (2012) and Murton and Murton (2012), located further upstream at the boundary between Chalk and Palaeogene rocks, is not entirely consistent with the proposed formation of the incisions, as the Chalk-Palaeogene contact is more than 50 km upstream of the incisions. Assuming that the lake margin was associated to a lithological boundary, the proposed location of the lake margin across the Strait and shown in Fig. 7.3 was associated to either Upper-Middle Chalk or Middle-Lower Chalk boundaries, as both are located just upstream of the incisions and may have represented the lake margin. Nevertheless, this interpretation remains purely speculative as there is no precise sedimentological evidence to suggest the exact position of the lake margin.

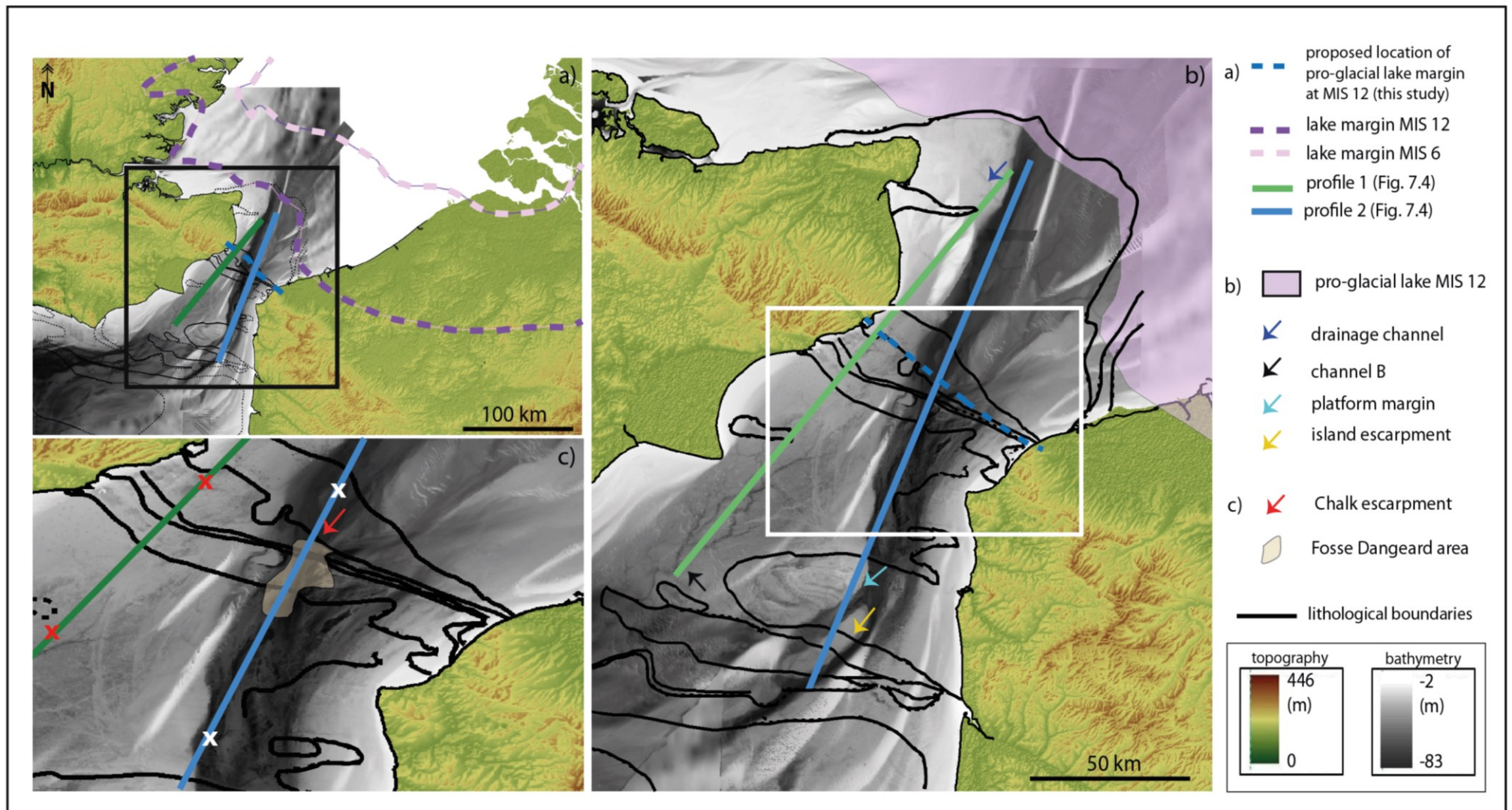
An alternative method for the reconstruction of the flood erosion depth during event 1 is by calculating the elevation of the lake using the elevation of the Chalk escarpment at Weald-Artois anticline, assumed to be the lake bedrock dam during Anglian glaciation. To do so, as there is no direct indication of the elevation of the bedrock dam at that time, the present-day onshore elevation of the Chalk ridge at the anticline can be used as a proxy. As the Chalk ridge shows, at present day, an elevation of ~100 m above mean sea level, as indicated by available STRM data (see Section 3.3 and Chapter 4 for details), and the base of the depressions is located 125 m below mean sea level, a total erosion depth of ~225 m is obtained (calculated using the present-day elevation of the Chalk escarpment), against a depth of ~155 m estimated using the lake elevation proposed by Hijma et al. (2012) and Murton and Murton (2012).

The elevation drop (see table 7.2) present at the Strait during the breaching of the bedrock dam and forming the palaeo-waterfall that carved the incisions can be estimated using the proposed elevation of the lake (or Weald-Artois anticline) and of the top of the incisions at the moment of their formation. The elevation of the bedrock platform (20 m below msl) or present-day mean sea level can be used as a proxy of the top of the incisions, obtaining an elevation drop that varies between 30 m (using the lake elevation and mean sea level) and 120 m (using the present-day Weald-Artois elevation and the bedrock platform). This calculation is nevertheless only a speculative estimation as there is no precise indication of the elevation of the top of the incisions at the time of their formation as they have been successively carved twice (see Fig. 7.2).

7.4.2 Erosion depth for event 2

For the second erosional event, erosion depth can be calculated as the difference in elevation between the bedrock platform and the internal erosional surface observed in the incision infill, interpreted to have been formed during event 2, and obtaining an erosion depth of ~ 65 m. The cross-cut relationship of the incision infill and the Lobourg Channel floor suggests that the erosion of the internal surface of the infill probably occurred at an earlier stage respect to the formation of the Lobourg Channel (see Fig. 7.2), although both events are related to event 2. The elevation drop at the Chalk escarpment during event 2 is interpreted to have amplitude of 25 m. This was calculated using the depth of the bedrock platform (20 m below msl) as the top of the incisions and the elevation of the pro-glacial lake and partially

eroded Weald-Artois anticline at MIS 6 (shown in Fig. 7.4) suggested to be 5 m by Cohen et al. (2005, 2011), Busschers et al. (2008) and Hijma et al. (2012).



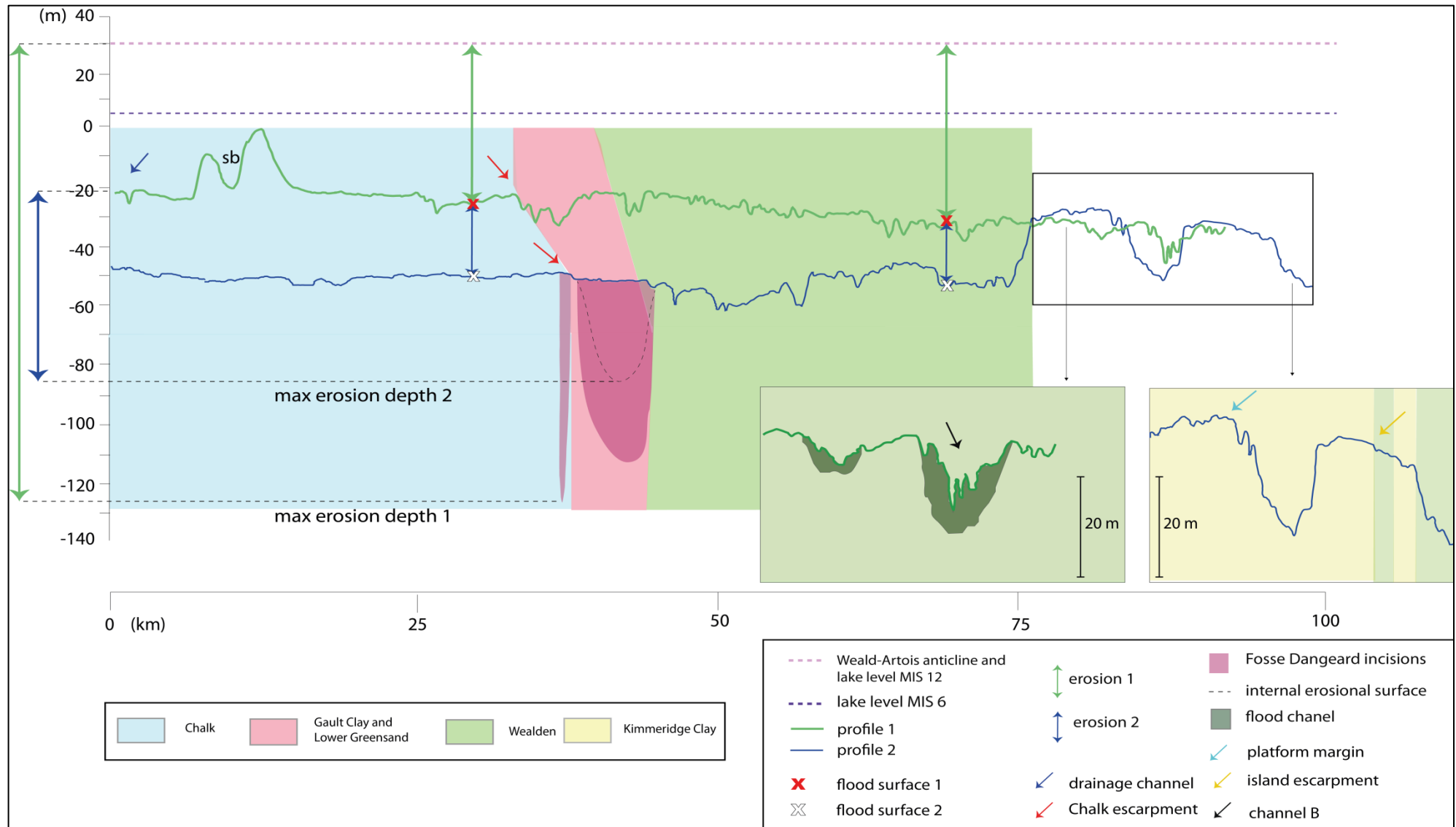


Figure 7.3(p. 287): Box a: cartoon of the location of the pro-glacial lake limit at MIS 6 and MIS 12 and of the two profiles shown in Fig. 7.3. Box b: Zoom of the Dover Strait area, showing the interpreted lithological boundaries and morphological features, indicated by the arrows and visible in the profiles (where the features are indicated by arrows with same colour). In particular, the blue arrow indicate drainage channel f1 cutting surface S1, the light blue arrow points the margin of the bedrock platform bounding the Lobourg Channel, the orange arrow points the streamlined island. Box c shows the location of the Fosse Dangeard incisions, while the red arrow points the chalk escarpment, also visible on both profiles. Lake margins at MIS 12 and 6 from Hijma et al., 2012 and Murton and Murton, 2012.

Figure 7.4(p. 288): Elevation profiles across the channel bench and bedrock platform (profile 1) and Lobourg Channel and streamlined island and platform margin (profile 2) overlapped to bedrock geology. Morphological elements on the bathymetry in Fig. 7.3 are also visible on the profiles. Note in both profiles the difference in the erosional pattern upstream and downstream of the Middle-Lower Chalk escarpment (red arrow). The transition between Kimmeridge and Wealden formation is also visible in profile 2 from a scarp on the surface of the island (yellow arrow). The lower left box represents the infill of palaeo Channel B carved into the platform. The elevation of the platform margin and streamlined island clearly shows how these surfaces lay at the same elevation as the probably represented a single wider flood surface before MIS 6. Lake level at MIS 12 from Cohen et al., 2011 and Hijma et al., 2012. Lake level at MIS 6 from Busschers et al., 2008; Cohen et al., 2011.

Table 7.2 summarises the calculated erosion depth and elevation drops for events 1 and 2, using different surfaces and elevations.

Event	Lake/Dam Elevation (m)	Top erosion (m)	Max erosion depth (m)	Erosion depth (m)	Elevation drop (m)
Event 1	30	0	-125	155	50
	100	-20		225	120
Event 2	5	-20	- 85	65	25

Table 7.2: Summary of the interpreted stages for the formation of the palaeovalley network from MIS 12 to Holocene. See text for details on the surface elevation.

7.5 Possible triggers of the megaflood events

Although the two proposed megaflood flows have been interpreted to have been originated by catastrophic drainage of pro-glacial lakes, as detailed in the section above (Smith, 1985; Cohen et al., 2005; Gibbard, 2007; Gupta et al., 2007; Busschers et al., 2008; Meinsen et al., 2011), the exact location and areal extension of the lakes is only partially constrained. It is nevertheless possible to speculate on the mechanisms that triggered the drainage of the pro-glacial lakes and caused the floods. If the source of both floods is interpreted to have been a pro-glacial lake located in the Southern North Sea (Figs. 7.3-5), the lake dam would have been, both at Anglian and Saalian times, the Weald-Artois anticline. Seismic reflection data interpreted in this work and described in Chapter 4 (see Fig. 4.8) and present elevation of Chalk ridge measured on SRTM data (~100 m) indicate that Chalk rocks are, at present, up to 200 m thick at the location of the proposed lake dam.. The Saalian pro-glacial lake, located to the north-east (Figs. 7.3-5) in the Southern North Sea (e.g. Cohen et al., 2005; Busschers et al., 2008; Hijma et al., 2012), was also bounded by the partially breached Weald-Artois anticline, although there is greater uncertainty on its position and areal extent.

The proposed driving mechanisms of the failure of the Chalk dam present at the south-eastern margin of the lake would have been overspilling of the lake, due to the exceptionally-high volume of water discharged into the lake by the diverted river network and melting ice-sheet. This would have been accompanied by two possible phenomena that triggered the sudden breaching of the Strait. Although there is no evidence of what triggered the flood, two different mechanisms are considered here. The first trigger of the floods would be an “internal mechanism”, represented by potential difference between the mass of water, elevated up to 100 m above the surface on the other side of the ridge that bounded the lake (Fig. 7.5). The difference in potential would have been sufficient to create a breach in the ridge, possibly through the propagation of cracks in the fractured Chalk of which the ridge was composed.

The second possible trigger of the dam breach would have been an “external factor” such as a tectonic event that occurred in the Strait. Earthquakes in the Dover Strait are not unlikely, as demonstrated by historical seismicity records (e.g. Melville et al., 1996; Musson, 2004, 2007, 2012; Baptie, 2007; Sargeant et al., 2008; Ottemoeller et al., 2009). The British earthquake record of the last three hundred years (see Section 2.5 of this work for details)

demonstrates that the Dover Strait and Sothern North Sea areas are the most seismically active locations in southern Britain (Fig. 7. 5), also being the location of two of the strongest earthquakes ever observed in the UK (1382 and 1850 quakes). The average recurrence times from the past thousands year event record (Musson, 1994) indicates that events of up to magnitude 6 M_L (like the 1382 quake) hs the potential to occur every 100 years, but it has only happened once since 1352.

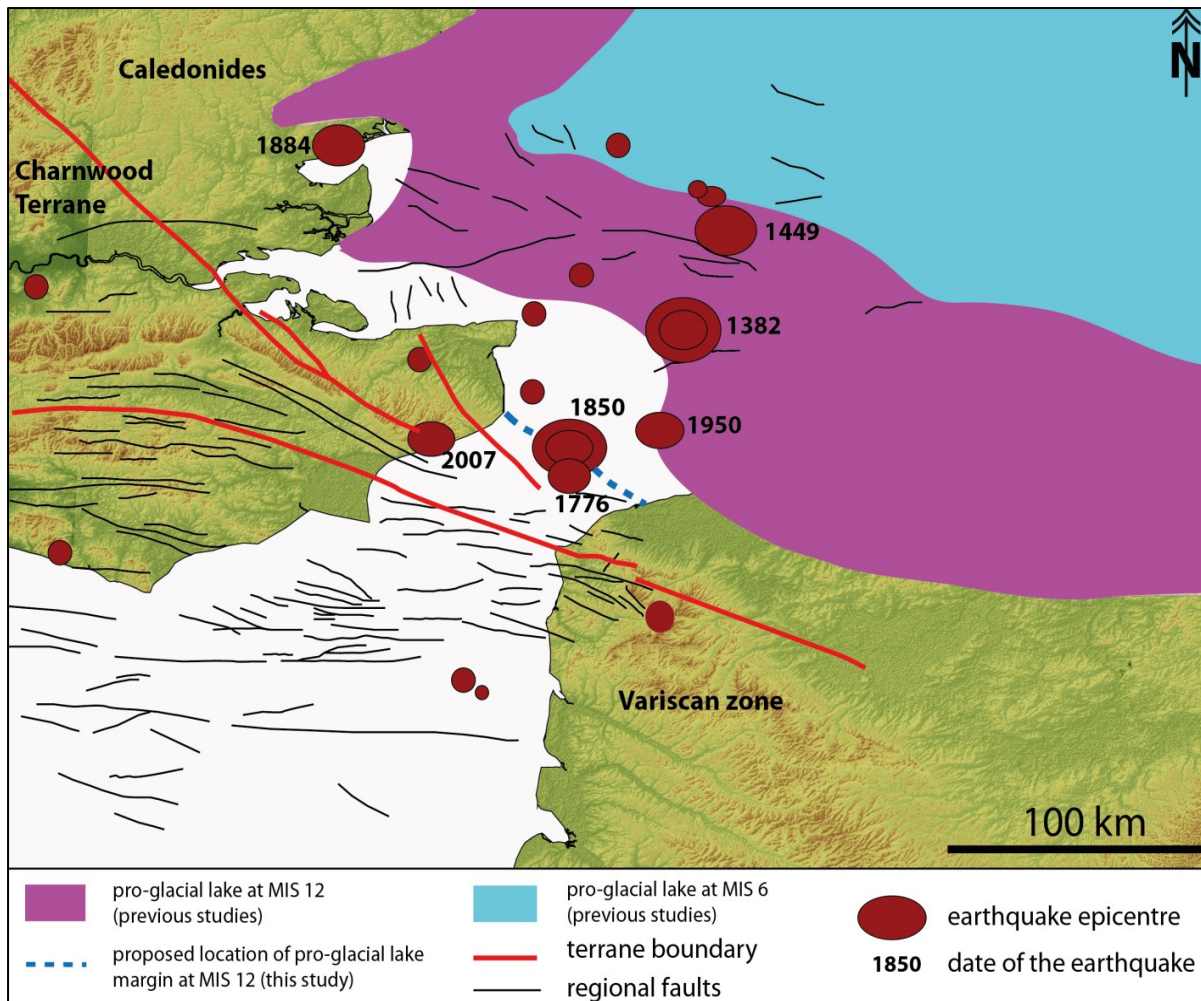


Figure 7.5: Cartoon of the location of the pro-glacial lakes at MIS 12 and MIS 6 plotted on the structural map of the study area and location of the epicentres of the major historical earthquakes. The diameter of the red circles is proportional to the magnitude of the quakes. (adapted from Musson, 2007: sources for lake location are the same as Fig. 7.3).

A tectonic event of similar magnitude may have occurred at MIS 12 and/or MIS 6 in the Southern North Sea area. Figure 7.5 shows that the proposed epicentres of the strongest quakes occurred at the Dover Strait/Southern North Sea area are located in the area occupied

by the Anglian (dotted blue line and pink area) pro-glacial lake and adjacent to the Saalian lake (blue area) boundary.

In addition, recent studies on seismic activity related to deglaciation (Mörner, 1979, 2004; Knight, 1999; Muir-Wood, 2000; Brandes et al., 2011, 2012) demonstrated that the presence of the ice-sheet and related pro-glacial lake have the potential to increase the lithospheric stress in the area adjacent to the ice-sheet front. The produced stress in the upper lithosphere, together with sediment and water loading linked to the ice movements and pro-glacial lake level fluctuations (Brandes et al., 2011) can cause the reactivation of existing tectonic structures present in the ice-sheet and pro-glacial lake area. Analyses on tectonic activity in the area of pro-glacial Lake Weser in north-western Germany (Brandes et al., 2011, 2012) demonstrates that faults with orientation parallel to the ice-sheet margin are more likely to be reactivated during deglaciation seismicity. As shown by Fig. 7.5, multiple faults with WNW-ESE orientation are present in the Variscan and Caldedonides terranes across the Strait. These show parallel orientation to the lithological boundaries present in the area, including the Chalk boundary that may have corresponded to the pro-glacial lake southern margin. The orientation of the regional-scale faults present in the area and parallel to the pro-glacial lake and to the ice-sheet margin would have favoured the fault reactivations during deglaciation times. This may have triggered tectonic activity and the consequent failure of the lake bedrock dam, forced by increasing lake-water levels due to exceptionally high input from melting ice-sheet and river drainage discharge during deglaciation times.

7.6 Summary

Following the reconstruction of the flood model, four main conclusions can be drawn:

- Two main flood events possibly composed of multiple pulses and identified as event 1 and event 2 are interpreted to have carved the Northern Palaeovalley channel network. Event 1 probably occurred at MIS 12 at the end of Anglian glaciation, representing the erosional event previously suggested to have been generated by the catastrophic drainage of a pro-glacial lake located in the Southern North Sea during Anglian (e.g. Smith, 1985; Gupta et al., 2007; Toucanne et al., 2009a, 2009b). This event would have partially breached the Strait for the first time, partially eroding the Chalk escarpment that bounded the lake. The flood would have carved the bedrock platform in the upstream area and the surfaces that are now the palaeovalley benches further downstream. Event 2 may have occurred at MIS 6, generated by the drainage of a pro-glacial lake that formed at the ice-sheet margin during Saalian glaciation. Event 2 was responsible for the erosion of the Lobourg Channel and the palaeovalley, together with the erosional remnants observed carved into the older flood surface. The flood would have breached the partially eroded Chalk ridge for the second time.
- The Fosse Dangeard incisions have also been interpreted to have been carved by events 1 and 2. The first flood event would have eroded the bedrock incisions during the breaching of the lake dam, with the kilometre-scale Dangeard depressions representing evidence of high magnitude flows breaching the Chalk ridge that bounded the Anglian pro-glacial lake. Sediments would have then filled the incisions, probably during interglacial times between MIS 12 and 6, with erosional events carving the internal erosional surfaces observed into the sediments. This would have been followed by sediment deposition and final truncation of the top of the incisions at the time of the formation of the Lobourg Channel, during one of the flows that composed event 2 at MIS 6.
- The calculations of the erosion depth for the two erosional events (in the areas where it was possible to clearly distinguish multiple erosional surfaces) suggest an erosion depth that ranges from 125 to 225 m for event 1 and 65 m for event 2. The elevation

drop present at the Chalk escarpment and formed by the presence of the Chalk ridge would have had amplitude of 30 to 120 m during event 1 and of 25 m during event 2.

- The Chalk dam failure that initiated the flood events, and which was caused by abrupt release of exceptionally high volume of water from the pro-glacial lake, may also have been triggered by a tectonic event in the Southern North Sea- Dover Strait area. Potential difference due to the elevation drop could have been the cause of the dam breach for both events. Alternatively, the occurrence of an earthquake in this seismically active area may have been related to the reactivation of existing tectonic features as the result of an increase in lithospheric stress caused by ice, sediment and lake-water loading during ice advance and retreat phases (e.g. Brandes et al., 2011, 2012).

8 . *Conclusions*

The reconstruction of past megaflood events can be performed through the study of ancient landscapes and preserved flood-generated bedforms that allow for the identification of typical flood landform assemblage. The study presented in this work examines the Northern Palaeovalley channel network and the associated erosional features carved into the English Channel seabed. Morphologic and seismo-stratigraphic analyses enable the mapping, characterisation and interpretation of the bedform geometry, cross-cutting relationships and distribution, thus allowing for the reconstruction of the nature, magnitude and relative timing of the events that carved the channel network. Following the results presented in the previous chapters of this thesis, six main conclusion points were drawn:

- 1) The interpreted bedforms show strong similarity to typical megaflood-related features described in previous studies on several terrestrial megaflood terrains. Remarkable similarity is observed in geometry, scale and distribution of the mapped features throughout the entire palaeovalley network, composing a typical flood-feature assemblage. Differently from most terrestrial megaflood terrains, no depositional flood bedform has been preserved in the English Channel terrain, probably due to later sediment remobilisation by sea-bottom currents or erosion during transgression.
- 2) As the seabed of the study area is characterised by outcropping bedrock, detailed geological mapping of the seafloor was performed from bathymetry, revealing a strong influence of the seabed geology on the geometry and distribution of the analysed flood features. In particular, morphometric analyses indicate that changes in lithology and different levels of resistance to erosion strongly affect the preservation of flood bedforms, as for the case of the mapped streamlined islands and channel benches. Changes in channel geometry are also strictly related to seabed geology, as shown by channel profiles of both Lobourg Channel and Northern Palaeovalley.
- 3) The performed analyses indicate the Dover Strait as the proposed flood spill-point: the Strait would have been breached by the flood flows originated by the catastrophic

drainage of a pro-glacial lake located in the Southern North Sea. The Chalk ridge present at the Dover Strait would have been the bedrock dam of the lake that generated the flood. The Fosse Dangeard incisions are interpreted as sub-circular plunge pools carved by the floods during the breaching of the bedrock dam, evidence of the dam-breaching process that initiated the megaflood through the catastrophic drainage of the lake. The geometry, infill and internal architecture of the incisions indicate at least two main erosional events to have breached the bedrock dam and carved the incisions.

- 4) Analysis of erosional surfaces and their cross-cutting relationships performed on seismic and bathymetry data are consistent with the interpreted megaflood model composed of two main flood events, each of which formed of possibly multiple erosional pulses. The floods would correspond to two distinct drainage events of pro-glacial lakes located in the Southern North Sea that would have occurred at the end of Anglian (MIS 12, ~450 kyr bp) and Saalian (MIS 6, ~150 kyr bp) glaciations, breaching the Chalk bedrock dam located at the Dover Strait. After the first flood, which would have partially breached the Strait and carved the palaeovalley for the first time, the Dover Strait area would have represented a shallow seaway during high-stands and a land-bridge between Britain and the Continent during low-stands. The complete breaching of the Strait would have occurred during the second event which also carved the Lobourg Channel and streamlined-shape remnants into the existing flood surface.
- 5) The performed palaeo-hydraulic calculations suggest flood peak discharge that ranges between 0.3 and 1 Sv for the two main erosional events that carved the Northern Palaeovalley and Lobourg Channel. The obtained values are consistent with typical megaflood peak discharge observed in major terrestrial megaflood terrains and are two to four orders of magnitude bigger than average discharge of Earth major rivers. In addition, the estimated peak discharge is consistent with the observed kilometre-scale erosional features and flood channels carved into bedrock for hundreds of kilometres, indicating erosion by exceptionally high-energy erosional events.

- 6) The proposed flood model is consistent with previous the studies on the palaeo-geographic and palaeontological evolution of north-western Europe during Middle and Late Pleistocene. This study indicates that the interpreted flood events that carved the Strait had a regional-scale effect, regulating the phases of connection/isolation of Britain from the continent, thus strongly affecting human migration and occupation patterns in Southern Britain, stressing the potential global-scale and long-term effect of megaflood events.

8. 1 Future work

In view of the significant impact of megaflood events on Earth landscape and climate, detailed knowledge and understanding of catastrophic floods and their driving mechanisms is needed. The presented results indicate that the Northern Palaeovalley channel network is one of the major megaflood terrains on Earth, illustrating how the English Channel megaflood terrain can provide important constrains on flood-feature characterisation and flood-event modelling and reconstruction. Following the research detailed in this thesis, new investigations can be carried out in order to: 1) confirm the interpretations presented in this work 2) set new constrains on the proposed megaflood model and 3) pose wider-scale key questions related to the research and analysis of Terrestrial megaflood events.

In particular, future investigations would target four main objectives:

- 1) Set new constrains on the absolute timing of the events that carved the Palaeovalley. To do so, cores and sedimentological analyses on the Fosse Dangeard infill sediments are needed. Drilling of the incision infill would provide crucial information on the nature and age of the material that fills the incisions. This would allow for a possible correlation of the infill material with sedimentological and geochemical data from cores located at the Bay of Biscay (Eynaud et al., 2007; Toucanne et al., 2009a, 2009b), in order to correlate the erosional events that breached the Dover Strait to the flows that deposited the sediments at the English Channel mouth further downstream. In addition, dating of the infill sediments would provide a precise constrain on the age of the events that deposited and successively carved the infill material, relating the erosional events presented in this work to the proposed glaciation/deglaciation phases.

Finally, provenance analyses on the drilled infill material could be performed in order to test different hypotheses (e.g. Gibbard, 2007; Gupta et al., 2007; Busschers et al., 2008; Toucanne et al., 2009a, 2009b; Meinsen et al., 2011) on the provenance of the floods and of the material that fills the incisions.

- 2) Build a better constrained palaeo-geographic model for the evolution of north-western Europe during Middle and Late Pleistocene. This would contribute to reconstruct a more precise timing of emersion and submersion stages for the Southern North Sea and English Channel, with particular focus on the Dover Strait area, aiming at the reconstruction of human occupation patterns (e.g. Ashton and Hosfield, 2010). To do so, a more precise reconstruction of pro-glacial lake location and areal extension and shoreline elevation during glacial and interglacial times is required.
- 3) Further investigation of the interpreted erosional longitudinal megalineations is needed, allowing for testing of multiple mechanisms proposed as responsible for the formation of the lineations, including the hypothesis of glacial mechanisms. This would imply the British ice-sheet to have reached the Dover Strait area during at least one of the Quaternary glaciations in Britain. Firstly, as the nature of the processes that formed these features can only be interpreted by analogy with similar bedforms, further morphologic and morphometric analyses are required. Secondly, as the high-resolution data availability is limited to the upstream area of the megaflood terrain, mapping of the spatial distribution of the lineations on a more extensive high-resolution dataset is crucial in order to investigate their possible presence in the downstream area. This can be achieved through the acquisition of a more extensive high-resolution bathymetric dataset throughout the study area. In addition, dedicated high-resolution reflection seismic would allow for the investigation of the presence of multiple generations of longitudinal lineations (e.g. Graham et al., 2007) carved below the seabed. The observation of the lineations on reflection seismic data would allow for a better understanding of the formation of the lineations, revealing what material they are carved into (i.e. Chalk, soft glacial sediments) and how many generations can be observed on a section other than that from bathymetric plan view.

- 4) Palaeo-hydraulic reconstructions of ancient megaflood flows require a detailed knowledge of channel geometry and hydraulic parameters that can only be determined from very high-resolution data. In the case of the Northern Palaeovalley network, availability of high-resolution data is restricted to a portion of the Lobourg Channel, limiting the precision of the performed peak discharge calculations. Collecting a more regional high-resolution bathymetry dataset would allow for the analyses of channel cross-sections from which the identification of two distinct channel sections (for the proposed first and second erosional events), channel inner benches and high-water marks throughout the entire palaeovalley network.

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Appendix

Appendix A. Channel Tunnel investigation works

Since the early nineteenth century, many attempts were made in order to create a connection between Britain and mainland Europe in the Dover Strait area, but the very first modern investigation related to the Channel Tunnel only took place in 1958 (Varley et al., 1996). After this first attempt, several campaigns were undertaken using borehole and geophysical investigations through seismic profiling, gaining a wide knowledge of the geology of the Dover Strait area. The main objective of the investigations was the geologic and the geotechnical characterisation of the different formations present along the proposed tunnel route, through direct (borehole tests and sampling) and indirect (reflection seismic) investigations. The geophysical surveys (Fig. A.1) that were collected for engineering and geotechnical purposes provide a wide database for investigating the geology of the Dover Strait area.

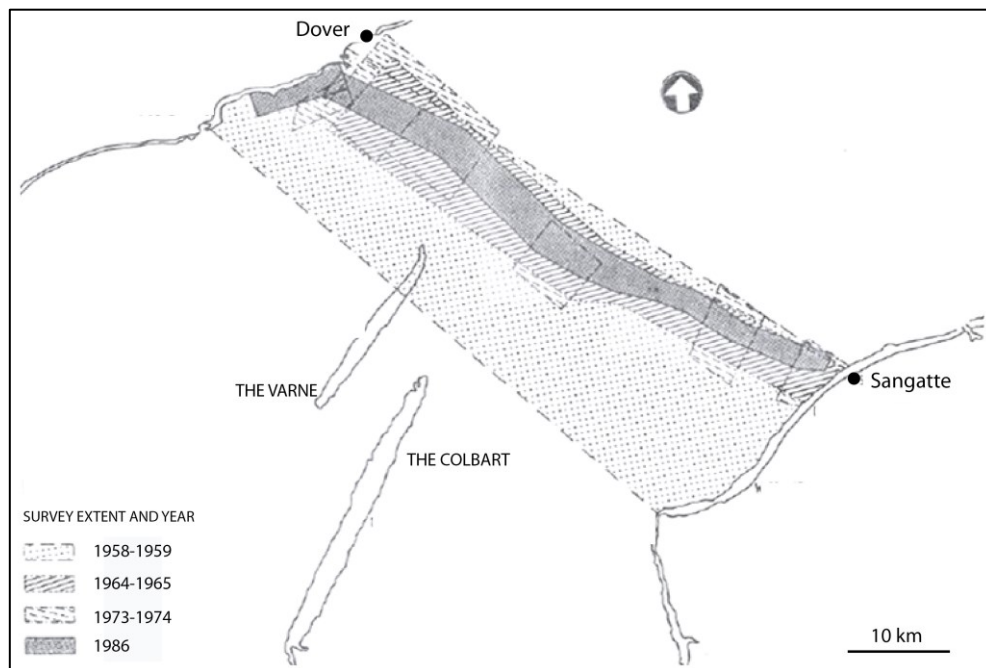


Figure A. 1: Geophysical survey extent of Channel Tunnel investigations (from Varley et al. 1996, modified).

The projected route of the Tunnel changed three times (Fig. A.2) after new information on the geology of the area was gained with new investigation campaigns, particularly after the detection, along the designed path of the Tunnel, of the Fosse Dangeard; this structure was

considered a geologic hazard for the construction of the project. Every campaign had different targets and focus as the investigation went ahead, ranging from wider stratigraphic characterization, to small-scale geotechnical and mechanical analysis of the Lower Chalk Formation where the tunnel was finally drilled in 1986. Next section describes the main purposes and results of the different campaigns conducted between 1958 and 1988 (from Varley et al., 1996).

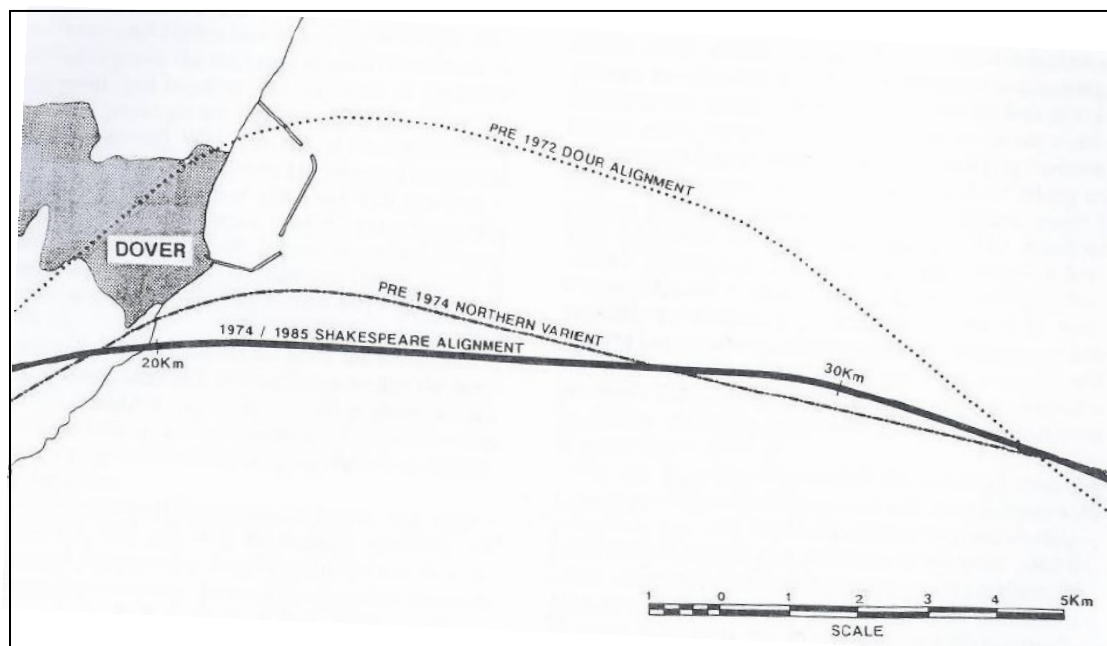


Figure A. 2: Cartoon of the projected Tunnel routes during geophysical investigations (from Varley et al., 1996).

A.1 Geophysical campaign results

1958/1959 Campaign: Two deep marine boreholes were drilled on the two side of the Channel from the Middle Chalk to the Lower Greensand, enabling a detailed bio-stratigraphic zonation of the Chalk. 1600 km of seismic lines were also acquired together with nine boreholes (called the W series), including one land borehole (W13). The results were of limited use from the geological point of view as the main information was the differentiation of the Glauconitic Marl and the definition of the Albian-Cenomanian boundary from the provided submarine crops. Pump tests were also performed in order to test the permeability and a logger was used to test the resistivity and self-potential, other than laboratory tests to measure moisture content, triaxial compression and velocity of longitudinal vibrations.

1964/1965 Campaign: This drilling campaign consisted of boreholes, in situ testing and laboratory tests. Particularly, the boreholes series were called in alphabetical order starting from Series P (8 boreholes including a borehole inside the shelter of Dover Harbour) to Series V (including borehole V50 of the Fosse Dangeard). The results led to two different interpretation of the thickness of the Chalk Marl (Craie Bleue in French) on the two sides of the Channel; this was due to different definitions of the Chalk Marl/Grey Chalk boundary adopted by the geologists of the two Countries. This issue stressed the importance of uniformity in the definitions and correlations on both sides of the Channel for future investigations. The *in situ* tests, which were strongly improved by the oil industry new techniques included velocity measurements of the Chalk, in order to correctly convert and interpret the geophysical data and velocity logs to mark eventual levels of velocity change. Amplitude logs were also performed to measure the degree of fracturing, gamma logs to detect the Glauconitic Marl-Gault Clay boundary and neutron logs related to water content and density, together with resistivity and calliper logs. Finally, the laboratory tests consisted of 1146 samples that allowed the determination of chemical and mechanical properties of the rocks.

1972/1973 Campaign: Additional boreholes to the previous campaign were performed mainly in the area of the English coast as a consequence of the identification of a deep weathering area and to operational constraints identified in the 1964 sparker survey. This set of investigations brought to a detailed bathymetry map of the area and to a broader knowledge of the morphology of the seabed and its approximate depth along the route. An updated geological map of the Channel was realised, identifying faults with throws up to 10 m on the French side and two wide channels that were called Fosse Dangeard. The V50 core revealed these channels to be filled with fine sand and marl of local origin. The geotechnical characterisation of the base of the Lower Chalk and its main lab properties were defined.

1986/1988 Campaign: This campaign was undertaken in two parts: pre construction investigations in 1986/87 (Phase I) and site-specific investigations, particularly addressed to the Fosse Dangeard area (Phase II), in 1988 (Varley et al., 1996; Arthur et al., 1996). Phase I main objective was the determination of the depth of the top of the Gault Clay within 5 m at maximum, through twelve control boreholes and seismic profiling in both sectors of the Tunnel. Deep profiles were also acquired in order to identify possible basement faults potentially connected to earthquakes in the Tunnel area. Phase II, on the other hand, had the

purpose of investigating the Fosse Dangeard area through seven boreholes drilled in the Chalk and Gault Clay, into the Lower Greensand and in the Fosse area (FD1 and FD2), together with seismic profiling that were collected particularly close-spaced, in order to detect small scale features such as shallow faults.

- Phase I results: the interpretation of the main reflectors and of the Fosse Dangeard was the main results of the geophysical data of Phase I (Fig. A.3). A litho-stratigraphic correlation was also made at the base and top of the Chalk Marl and Grey Chalk. On the geotechnical side, more permeability test and engineering properties of the Lower Chalk were tested, marking the anisotropy and the variations in stiffness and strength that characterise the Lower Chalk. In summary, this investigation phase led to a more reliable degree of knowledge and geological characterisation of the area and no further adverse features to the construction of the Tunnel were identified.
- Phase II results: this last phase of the investigations was designed to drill and characterise the infill of the Fosse that represented a potential geological hazard to the Tunnel. The infill, in fact, showed different geotechnical and lithological character respect the bedrock where the scours were carved, being made of unconsolidated material of unknown origin. Not only the main scour was monitored, but also the tributary valley that had been recognised during previous investigations was considered as a threat to the stability of the drilling operations and to the Tunnel itself. More seismic reflectors were identified in the Fosse Dangeard area and its orientation was interpreted as being linked to small-scale faults that affect the Cretaceous strata with a NW-SE trend, parallel to the structural pattern of the area. Through the deep target seismic, deeper faults were recognised; in particular, a major Palaeozoic fault already recognised during Phase I was repositioned and its throw was measured as up to 50 m, probably associated to the Variscan thrust.

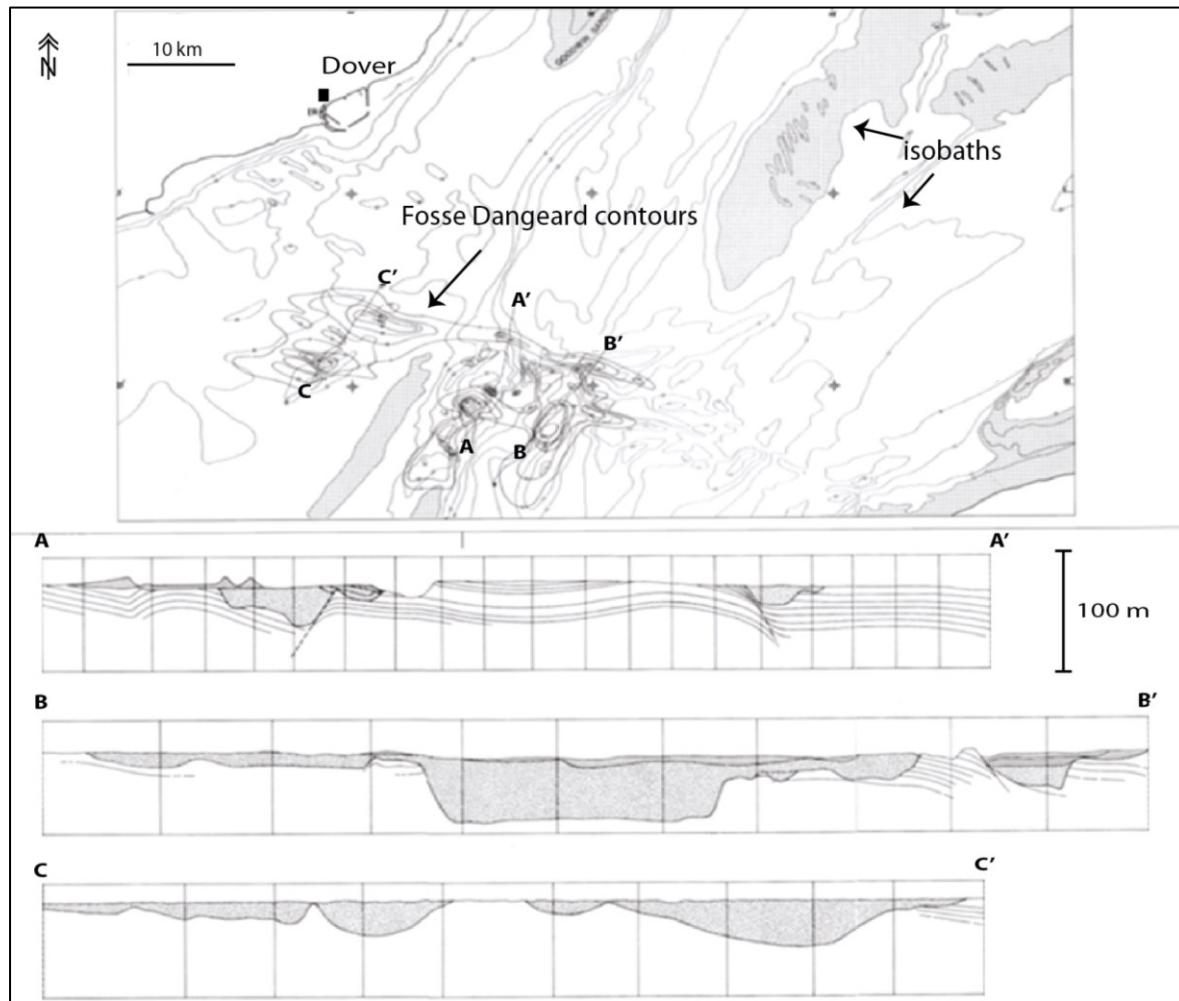


Figure A. 3: Map and along-axis sections of the Fosse Dangeard drawn by Destombes et al. (1975) after the Fosse Dangeard investigations during Phase I.

Appendix B. MCA MBES Tidal Model and Corrections

This section details supplementary information on the tidal corrections applied on MCA MBES data during acquisition and processing phase by UKHO.

B.1 Tidal models

As the surveyed area is strongly affected (up to more than 6 m in Dover) by the effect of tides, the tide correction procedure played a key role in the final quality of the data. Two main tidal models were calculated and applied in different areas, depending on the nearest tide gauges. Three tide gauges were employed: Dover, Beachy Head and North Foreland (Fig. B.1 and Tables B.1 and B.2), although the latter two were only available for part of the survey time.

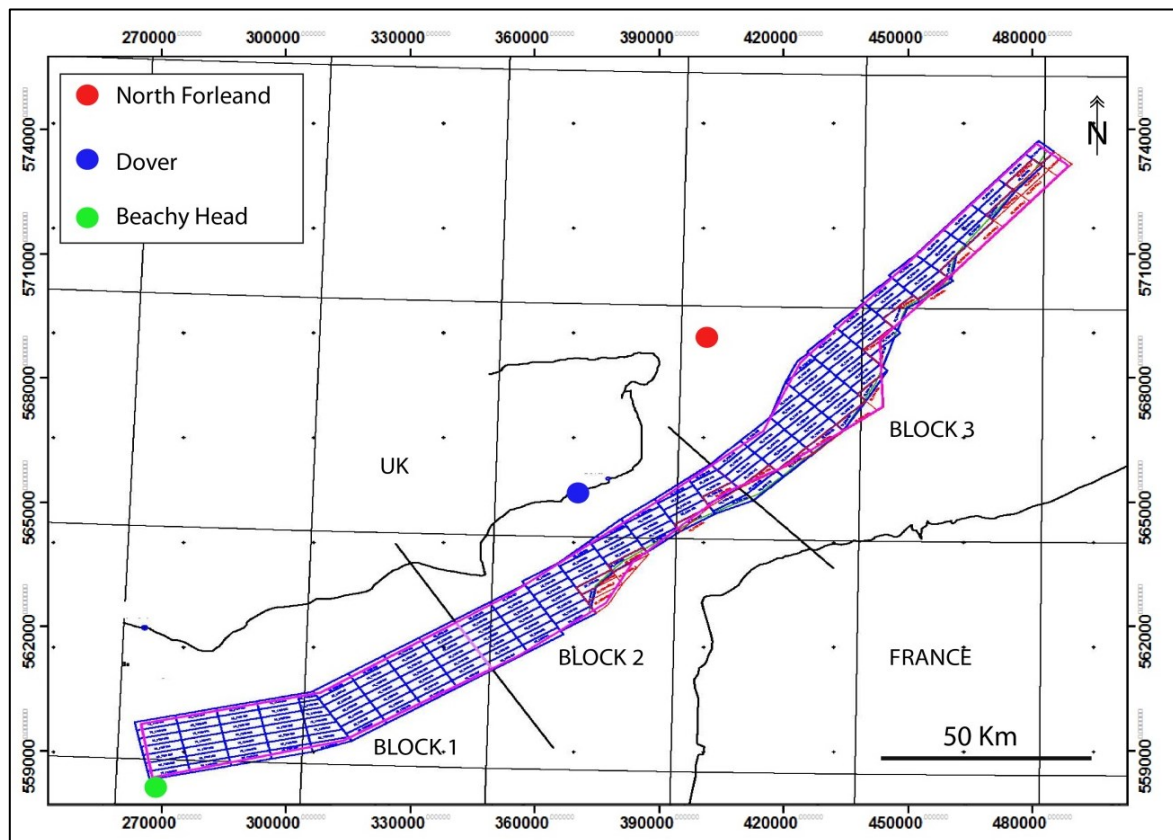


Figure B. 1: Location of the tide gauges employed for the survey (from navigation report received with data).

Despite two tidal models were calculated, their complexity such as high and low water arrivals, could not be supported by the acquisition software, resulting in a simple model that only suited part of the dataset. In particular, during acquisition and pre-processing, GPS derived heights were used instead of the tidal model in order to reduce the height gap that was generated in those areas that were located very far from the tide gauge Dover. The tidal model was successively applied to the data before the final delivery.

In particular, other than the distance of some areas from the main tide gauges the tidal model showed the following problems:

- Several tidal pressure sensors were lost as vessels hit the buoys.
- The available Dover co-tidal charts have different scales and resolutions and do not contain low water levels and the resolution is therefore different for different areas.

A complex model with a significant shift between high and low water times would have been necessary but Caris HIPS cannot apply such a complicated model and no software on the market is able to apply these kinds of model.

Tide gauge	Position		Location	Owning Authority
	Lat	Long		
Beachy Head	50°40.68'N	0° 06.81'E	Offshore	OSAE Temporary Installation
North Foreland	51°43.70' N	1° 57,60'E	Offshore	OSAE Temporary Installation
Dover	51° 06.86' N	1° 19.33'E	Prince of Wales Pier	Dover Harbour Authorities

Table B. 1: Summary of the location and main characteristics of the three tide gauges (from navigation report).

Tidal Model	Block 1	Block 2	Block 3
Dover	not used	091_111	113_124 (note that there is no cell 112)
Dover, Beachy Head, North Foreland	001_063 (entire Block 1)	064_090 (Western Block2)	125_175 (Eastern Block 3)

Table B. 2: Summary of the tidal models applied to cells.

Appendix C. Supplementary diagrams for statistical analyses of streamlined islands

Histogram and regression plots showing the statistical analyses made on the morphometric parameters calculated on the streamlined island population and detailed in Section 6.5 of this work.

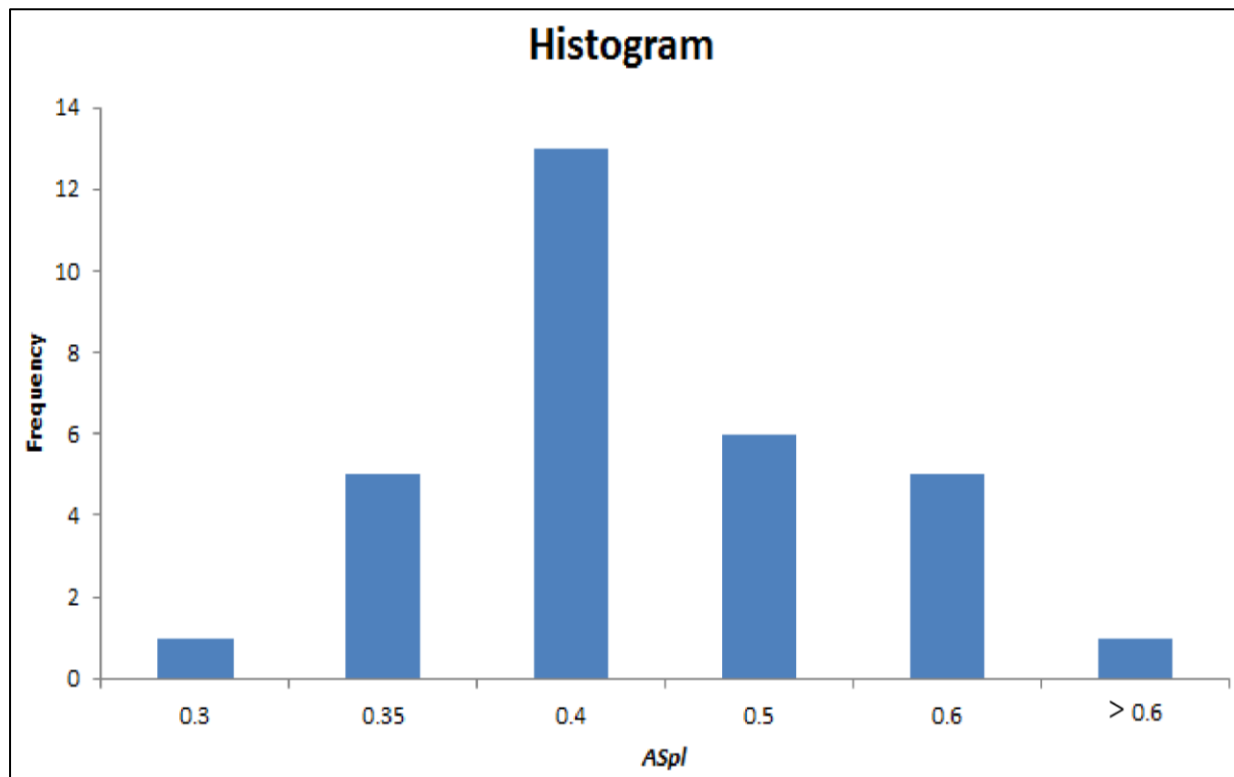


Figure C. 1: Normal distribution of the ASpl values calculated for the streamlined islands.

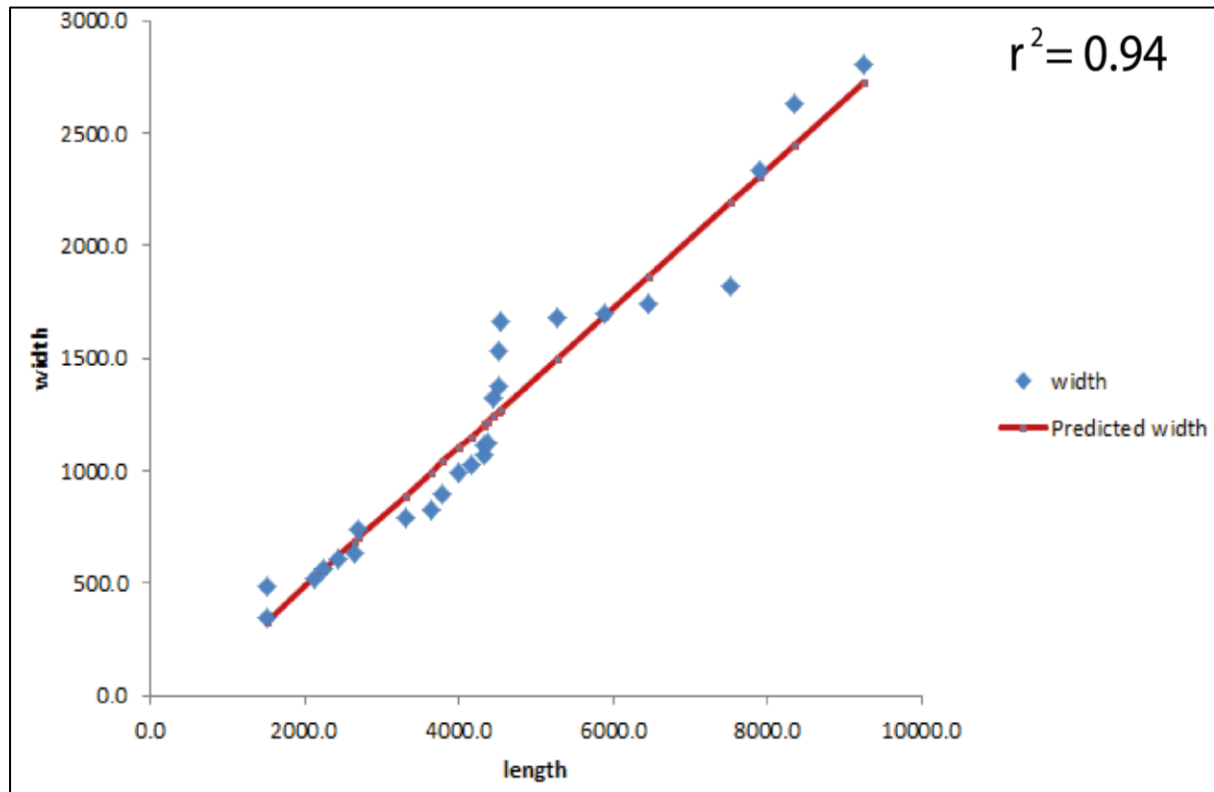


Figure C. 2: Island width versus length plotted for the entire population. Values are in m.

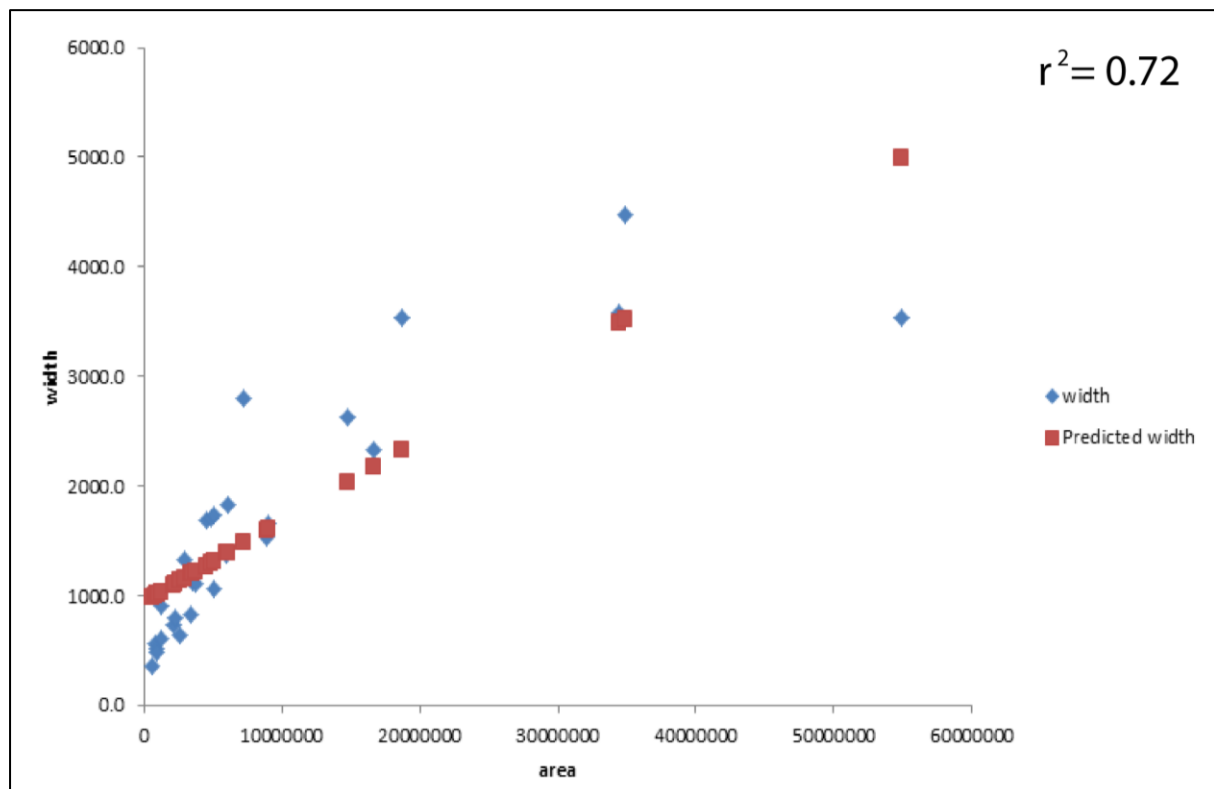


Figure C. 3: Island width versus area plotted for the entire population. Values are in m.

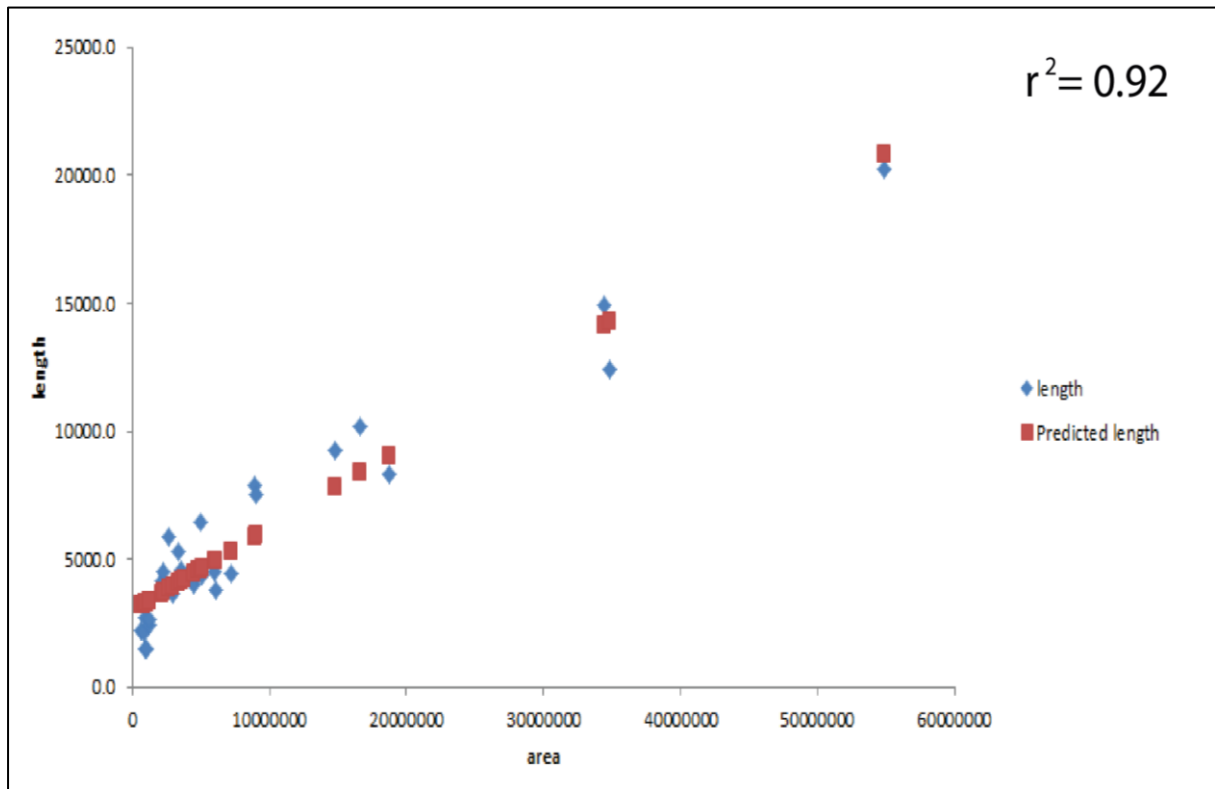


Figure C. 4: Island length versus area plotted for the entire population. Values are in m.

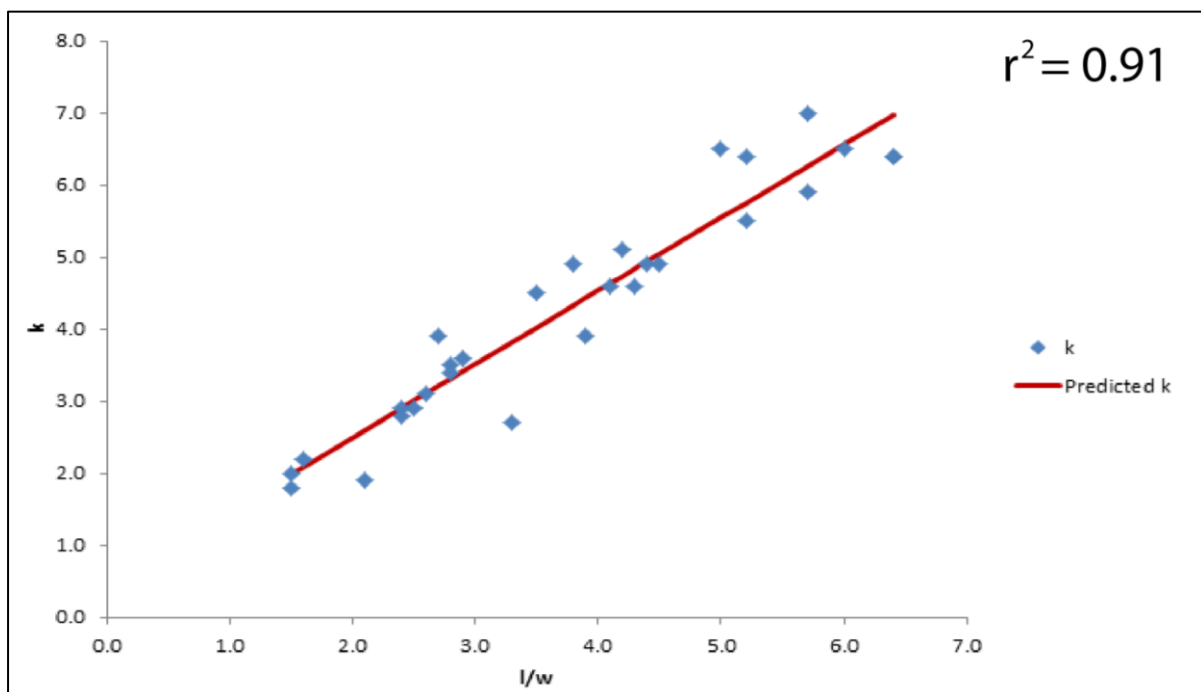


Figure C. 5: Island k values versus l/w ratio plotted for the entire population. k represents shape coefficient of the islands, as explained in Section 6.5 of this work.

