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Drowned landscapes of the Belgian Continental Shelf

implications for northwest European landscape evolution and preservation potential for submerged heritage

Maikel De Clercq

Academic year 2017–2018

Supervisor: Prof. Dr. Marc De Batist

Co-supervisor: Dr. Tine Missiaen

Submitted to the Faculty of Science of Ghent University
in fulfilment of the requirements for the degree of
Doctor of Science: Geology

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Verdronken landschappen van het Belgisch deel van de Noordzee:

implicaties voor noordwest Europese landschapsontwikkeling en bewaringspotentieel van onderwater erfgoed

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RESEARCH HIGHLIGHTS

The expansion and reduction of ice masses across the North Sea Basin has a profound impact on eustatic sea levels and glacio-isostasy across northwest Europe. The complex interaction between eustatic sea level and glacio-isostatic adjustment of the landscape, from the time of ice mass expansion to their deglaciation, exerts a major control on local/regional landforms such as rivers and how fast or slow relative sea level movement occurs on a local/regional scale.

Proglacial lakes in the North Sea Basin may have been a common landform during glacial periods. They were fed by meltwater from the Fennoscandian and British-Irish ice-sheets causing their levels to rise and exert major pressure on their confining barriers. During ice mass disintegration major volumes of meltwater enter proglacial lakes and may cause them to burst out of their confinements resulting in catastrophic glacial lake outburst floods.

A high-resolution depth-converted structure map of the pre-Quaternary surface combined with regional palaeogeographic reconstructions demonstrates that **the late Middle to Late Pleistocene geological evolution of the Belgian Continental Shelf is very complex**. The various regions in which different depositional environments are preserved testify this. The Coastal Plain-Inner Shelf comprise the buried incised-valley system of the palaeo-Scheldt River and is mostly controlled by local/regional-scale processes, whereas the Middle-Outer Shelf landforms are controlled by southern North Sea glacial lake outburst floods from the Penultimate and Late Glacial period.

The formational process interactions of and the interactions between depositional environments control the preservation potential of submerged heritage. Preservation of archaeological artefacts and palaeontological and human bone material is different for each depositional environment. Landforms and local palaeotopography control deposition and erosion in the landscape, which in turn determine if any form of submerged heritage can be preserved or accumulated within a primary, secondary or tertiary context.

The preservation potential for archaeology and palaeontology has been visualised for the first time. Profile type mapping was used as a technique. The maps can be used as a guidance tool for preservation potential assessment at sea. This helps commercial-industrial end-users and policy makers to determine the risk for damage to the submerged heritage. *Vice versa* this pro-active engagement helps them save time and money for offshore activities because unexpected delays are prevented this way.

SUMMARY

The Belgian Continental Shelf is the smallest exclusive economic zone of the North Sea and is bounded by the French, British and Dutch Continental Shelves, however the number of economic and societal claims is very large, comprising fishing, military defence, sand extraction, maintenance dredging, shipping, wind farms, pipelines and cables. On top of that, the scarce economic resource of the Quaternary sediments of the Belgian Continental Shelf is thin and fragmented, which is the main reason why it has been so difficult up to now to produce a coherent reconstruction of the Quaternary evolution of the BCS. Moreover, the limited thickness of the Quaternary cover makes its entire section potentially vulnerable to the economic claims. Marine spatial planning hence exists to accommodate these various economic interests. It is a process to allocate the space available on sea to certain actors within a certain time frame and at the same time to ensure that all ecological, economic and social objectives are achieved. An ecosystem vision and involvement of all stakeholders (science, industry, policy) are hereby of great importance.

Within the Marine Spatial Plan of 2014 it is recognised that next to shipwrecks all traces of human presence, be it cultural, historic or archaeological, belong to the submerged heritage. However, the marine spatial plan does not elucidate on how this submerged heritage is to be protected from damage or destruction that comes forth from these economic claims. Furthermore, it does not elaborate on palaeontological bone material, which is inherently connected to past human activities and use of the land and is therefore a fundamental part of this submerged heritage (from here on, submerged heritage refers to drowned palaeolandscapes, prehistoric archaeology and palaeontological bone material). An efficient policy at sea is therefore imperative. A first step to achieve a more efficient policy was developed by the SeArch project (2013–2016). The project offered solutions to improve the offshore policy by: 1) providing stakeholders with new and improved remote sensing technologies and an efficient survey methodology for the assessment of submerged heritage; 2) to provide stakeholders with a new, sustainable management framework regarding submerged heritage, and including marine spatial planning; 3) to provide practical guidance for the stakeholders on how to implement the new methodology and management approach.

The one thing that remains underrepresented to provide the necessary tools for an efficient offshore policy is a thorough overview of the Quaternary stratigraphy of the BCS and what geological formation processes lie at its base. This is a pivotal step to understand why and where

submerged heritage may be preserved within the Quaternary deposits, i.e. the preservation potential.

The **first objective** of this research was to further complement and refine the growing database of seismic reflection and sediment core information regarding the geomorphology of the base of the Quaternary deposits (Chapter 4). In this perspective additional seismic surveys were performed between 2013 and 2017 to provide a denser seismic grid for the whole BCS. Complementary information from sediment cores and vibrocores resulted in a high-resolution depth-converted structure map of the base of the Quaternary, also known as the top of the Paleogene substratum. This pre-Quaternary surface is an important polygenetic unconformity that truncates older strata of the Paleogene and to a smaller extent some of Neogene age from the overlying Quaternary deposits. The represented surface has been diachronously shaped and reworked through Middle and Late Pleistocene times by different geological processes (e.g. fluvial, marine, estuarine, periglacial). Additionally, the offshore surface has been attached to the landward pre-Quaternary surface and was transformed into a uniform three-dimensional surface allowing new and better interpretations to be used in fundamental and applied research underpinning both scientific purposes (e.g. geology, archaeology, palaeogeography), and commercial applications (e.g. wind farm construction, aggregate extraction, dredging).

A geomorphological analysis of this surface revealed that its structure is highly complex and consists of two striking incised palaeochannel orientations bounded by escarpments: 1) a southeast-northwest orientation stretching from the Coastal Plain to the Inner and Middle Shelves (Chapter 5); 2) the second orientation of palaeochannels has a north to south-southwest orientation and is confined to the Outer Shelf (Chapter 6). The formation of the Middle Shelf is considered influenced by both geological processes that formed these two dominating palaeochannel incision orientations.

The **second objective** of this research was to understand the landscape evolution and the formational processes of the different landforms and depositional environments preserved on the Belgian Continental Shelf (Chapters 5 and 6).

Chapter 5 demonstrates that the BCS and its coastal plain occupy a key position between the depositional North Sea Basin and the erosional area of the Dover Strait as it is an area where erosional landforms and fragmented sedimentary sequences provide new evidence on northwest European landscape evolution. The Coastal Plain-Inner Shelf area host 20–30 m thick penultimate to last glacial sand-dominated sequence that is preserved within the buried palaeo-

Scheldt Valley. Here, we build on the results of previous seismo- and lithostratigraphical studies, and present new evidence from biostratigraphical analysis, OSL dating and depth-converted structure maps, together revealing a complex history of deposition and landscape evolution controlled by climate change, sea-level fluctuations and glacio-isostasy. This study presents new supportive evidence on the development, incision and infilling of the incised palaeo-Scheldt Valley landform that became established towards the end of the penultimate glacial period (MIS 6: Saalian) as a result of glacio-isostatic forebulge updoming, proglacial lake drainage and subsequent relaxation of a forebulge between East Anglia and Belgium following ice-sheet growth, disintegration and retreat in areas to the north. The majority of the incised-valley fill is of estuarine to shallow marine nature, and its age constrains the transgression and high-stand to the last interglacial (MIS 5e: Eemian). A thin upper part of the valley fill consists of last glacial (MIS 5d-2: Weichselian) fluvial sediments that show a gradual decrease and retreat of fluvial activity to inland, upstream reaches of the valley system until finally the valley ceases to exist as the combined result of climate-driven aeolian activity and possibly also glacio-isostatic adjustment. Thus, strong contrasts exist between the palaeo-Scheldt Valley and estuarine systems of the Penultimate Glacial Maximum to Last Interglacial (Saalian to Eemian), the beginning of the Last Glacial (Weichselian Early Glacial and Early-Middle Pleniglacial), and the Last Glacial Maximum to Holocene.

Chapter 6 discusses the Middle-Outer BCS in the southern North Sea. This area is characterised by a series of north to south-southwest-oriented buried, elongated and interconnected palaeodepressions, eroding last interglacial marine deposits. Despite the fact that these erosional/sedimentary features have been studied by several authors, their origin and interrelationship remained largely unknown. Here, we re-evaluate their formation and relationship by combining a large set of sediment cores with a tight high-resolution seismic grid, and by comparing their combined interpretation with recent palaeoclimatic/palaeogeographic reconstructions of northwest Europe. The sedimentary record evidences the formation of an early last interglacial shallow marine embayment that was to the south controlled by a late penultimate glacial Meuse River terrace. These sediments are truncated by the buried elongated palaeodepressions, which we interpret as erosional features carved within a floodplain of the Axial Channel trunk valley that was subjected to one or several episodes of major flooding during the Last Glacial. These episodes may have been produced by massive meltwater discharges into the North European Plain drainage system that entered the southern North Sea at the Danish–German continental shelf during time intervals when ice-sheets merged across the North Sea Basin. These meltwater pulses might have promoted ice-sheet destabilisation, releasing icebergs into the same drainage system, which may have included a large proglacial lake located in the Dutch-Danish–German continental shelf. This lake and drainage system is interpreted as the transportation medium of erratic clasts from the Scottish Grampian Highlands and the British East Coast that are currently found scattered along the southern North Sea and the Middle-Outer BCS. Indeed, erratic clasts from these locations could have reached the southern North Sea during episodes of rapid ice melting, which may have induced GLOFs, transporting icebergs and debris in the direction of the Dover Strait.

The **final objective** of this research is to understand and visualise the preservation potential for submerged heritage on the BCS (Chapters 7 and 8). Detailed knowledge about the distribution and formational process characteristics of the characterised depositional environments can now be used to assess, for the first time ever, the preservation potential of any form of submerged heritage (Chapter 7). The BCS shares geomorphological affinities with the northern part of Belgium (Flanders): 1) both areas have a thin and fragmented Quaternary cover thereby exposing the Paleogene substratum; 2) platforms, such as the Campine area, are comparable to the Inner, Middle and Outer BCS; and 3) the Flemish Valley system is connected to the downstream palaeo-Scheldt Valley, that extends to the Inner and Middle BCS. This geological affinity between both areas provides the unique opportunity to expand the current state-of-knowledge of onshore Flanders to the now submerged area of the shelf.

The **final result** is two maps that visualise the preservation potential of submerged heritage, one for archaeology and one for palaeontology (Chapter 8). The archaeological potential map indicates areas from which period archaeological material can be expected, while the palaeontological potential map highlights areas where certain types of faunas can be expected. These maps are tools that can be used to develop a sustainable management policy for submerged heritage in the near future. This way both the scientific, industry and policy communities can benefit from each other.

Samenvatting

Het Belgisch deel van de Noordzee is de kleinste exclusieve economische zone van de Noordzee en is omringd door Franse, Britse en Nederlandse wateren. Hoewel het Belgisch aandeel zeer klein is dat niet het geval voor het aandeel aan economische en maatschappelijke belangen. Deze omvatten onder meer visserij, militaire verdediging, zand extractie, onderhoudsbaggerwerken, scheepvaart, windmolen parken, pijpleidingen en kabels. Daarbovenop komt nog eens dat de economisch waardevolle Quartaire sedimenten dun en gefragmenteerd zijn. Dit laatste is de reden waarom tot op heden nog geen coherente reconstructie bestaat van de Quartair geologische evolutie van het Belgisch deel van de Noordzee. De dunne en gefragmenteerde samenstelling van deze sedimenten wordt bovendien nog eens sterk beïnvloedt door de brede waaier aan economische en maatschappelijke belangen. Om belangenconflicten te minimaliseren bestaat er een marien ruimtelijk plan, dat voor elk van deze belangen een specifieke ruimte op zee ter beschikking stelt zodat alle ecologische, economische en maatschappelijke doelstellingen gewaarborgd worden. Om deze reden is een ecosysteem visie en een betrekking van alle belanghebbende partijen (wetenschap, industrie en beleid) van groot belang.

Het Marien Ruimtelijk Plan 2013 herkend dat naast scheepswrakken alle sporen van menselijke activiteit, zij het van culturele, historische of archeologische aard, deel uit maken van het onderwater erfgoed. Ondanks deze stelling houdt het plan geen rekening hoe dit onderwater erfgoed beschermt dient te worden van schade of vernietiging veroorzaakt door de verschillende economische en maatschappelijke belangen. Bovendien houdt het plan geen rekening met paleontologisch botmateriaal, wat op zijn beurt inherent verbonden is met menselijke activiteiten en hoe het land gebruikt werd om in te overleven. Dit onderdeel is bijgevolg een fundamenteel onderdeel van het onderwater erfgoed. Om deze redenen dringt zich een duurzaam en efficiënt beleid tot het beschermen van het onderwater erfgoed op. Een eerste stap in deze richting werd genomen door het SeArch project (2013–2016). Dit project leverde een reeks van protocollen aan om tot efficiëntere activiteiten op zee over te gaan met als doel het offshore beleid te verbeteren: de belanghebbende partijen werden 1) nieuwe en verbeterde technologieën aangereikt om tot een efficiëntere survey methodologie over te gaan om onderwater erfgoed betere te kunnen inschatten; 2) een verbeterd en duurzamer beleid ter beschikking gesteld hoe men dient om te gaan met onderwater erfgoed; 3) een praktische en gebruiksvriendelijke handleiding overgemaakt hoe men de nieuwe methodologie en beleidsregels op een eenvoudige en duurzame manier kan uitwerken en toepassen.

Het enige wat tot op vandaag nog ondervertegenwoordigd blijft om tot een efficiënt beleid over te gaan is een degelijke en overzichtelijke stratigrafie van de Quartaire sedimenten en welke vormingsprocessen aan hun basis liggen. Dit is een fundamenteel onderdeel om te begrijpen waarom en waar het onderwater erfgoed bewaard is gebleven in de Quartaire afzettingen. Dit wordt ook wel het bewaringspotentieel genoemd.

De **eerste doelstelling** van dit onderzoek bestond erin de bestaande databank van seismische reflectie en boorgegevens verder uit te werken en te vervolledigen om de geomorfologie en de basis van de Quartaire sedimenten beter in beeld te brengen (Hoofdstuk 4). Om dit te beogen werden bijkomende seismische campagnes uitgevoerd in de periode 2013–2017 om een dichter seismisch netwerk voor het Belgisch deel van de Noordzee te bekomen. Complementaire informatie van boorkernen en vibrocores resulteerden uiteindelijk in een hoog-resolutie beeld van de diepte-geconverteerde structurele kaart van de basis van het Quartair, dat ook gekend is als het Paleogeen substraat. Dit pre-Quartair oppervlak is een belangrijk polygenetische discordantie dat oudere Paleogene, en tot een beperkte orde ook Neogene, sedimentlagen afsnijdt van de bovenliggende Quartaire sedimenten. Het resulterende oppervlak werd tijdens het Midden-Laat Pleistoceen gevormd en herwerkt door verschillende geologische processen (e.g. fluviatiel, marien, estuarien, periglaciaal). Voorts is het offshore deel van dit pre-Quartair oppervlak verbonden met dat van Vlaanderen en werd het getransformeerd in een driedimensionaal vlak om een fundamenteler en beter toegepaste studie ervan mogelijk te maken voor zowel wetenschappelijke (e.g. geologie, archeologie, paleogeografie) als commerciële doeleinden (e.g. windmolenparken, aggregaat extractie, baggerwerken).

Een geomorfologische analyse van dit oppervlak heeft aangetoond dat de structuur van dit oppervlak uitermate complex is en dat het twee sterk opvallende insnijdingsrichtingen van paleorivieren vertoont met een verschillende oriëntatie en die afgelijnd zijn door hellingen in het landschap: 1) een zuidoost-noordwest oriëntatie van de kustvlakte tot het Binnen en Midden Deel van het Belgisch deel van de Noordzee (Hoofdstuk 5); 2) de tweede insnijding heeft een noord-zuid tot noordoost-zuidwest oriëntatie die beperkt is tot het Buiten Deel van het Belgisch deel van de Noordzee (Hoofdstuk 6). De vorming van het Midden Deel van het Belgisch deel van de Noordzee wordt beschouwd als onderdeel van beide paleogeul insnijding oriëntaties.

De **tweede doelstellin**g van dit onderzoek had tot doel de landschapsontwikkeling en haar vormingsprocessen voor de diverse landschapeselementen en haar afzettingsmilieus beter te begrijpen (Hoofdstukken 5 en 6).

Hoofdstuk 5 demonstreert dat het Belgisch deel van de Noordzee en haar kustvlakte een essentiële positie innemen tussen het sediment accumulerende Noordzeebekken en het erosiegebied van de Straat van Dover, omdat het een gebied is waar erosieve landschapselementen en gefragmenteerde sedimentaire sequenties bewaard gebleven zijn die tot nog toe verborgen elementen van de noordwest Europese landschapsevolutie herbergen. De kustvlakte en het Binnen Deel van het Belgische deel van de Noordzee herbergen een 20–30 m dikke zandige sedimentaire sequentie bewaard in de paleo-Schelde Vallei dat dateert van het voorlaatste tot het laatste glaciaal. In dit onderzoek wordt op de resultaten van voorgaanden seismische en lithostratigrafische studies gebouwd en worden deze gecomplimenteerd met nieuwe aanwijzingen afkomstig van biostratigrafische analyses, OSL dateringen en diepte-geconverteerde structurele kaarten om zo tot een nieuwe en complexe geschiedenis van sedimentaccumulatie en landschapsevolutie te komen dat gecontroleerd werd door klimaat- en zeespiegelveranderingen en glacio-isostasie. De studie hiervan presenteert nieuwe bewijzen voor de ontstaansgeschiedenis, insnijding en opvulling van de paleo-Schelde Vallei als landschapeselement tegen het einde van het voorlaatste glaciaal (MIS 6: Saaliaan) als gevolg van glacio-isostatische opwaartse bewegingen van de aardkorst tussen East Anglia en België na het aangroeien van de ijskappen in het noorden, een proglaciaal meer dat leegloopt op catastrofale wijze en het ineenzakken van de opwaarts gestuwde delen van het landschap. Het merendeel van de vallei is opgevuld met estuariene en ondiep mariene afzettingen tijdens de transgressie en zeespiegel hoogstand van het laatste interglaciaal (MIS 5e: Eemiaan). Het bovenste deel van de vallei is opgevuld met rivierafzettingen van het laatste glaciaal (MIS 5d-2: Weichseliaan) en tonen aan dat de activiteit van de rivier geleidelijk aan afneemt en zich verplaatst naar het noordwesten ten gevolge van klimaat-gedreven eolische activiteit en mogelijks ook glacio-isostasie. De studie van de paleo-Schelde Vallei toont aan dat sterke contrasten bestaan tussen de rivier en estuariene systemen van het voorlaatste glaciale maximum tot het laatste interglaciaal (Saalian, Eemiaan), en het riviersysteem aan het begin van het laatste glaciaal (Weichseliaan Vroeg Glaciaal en Vroeg-Midden Pleniglaciaal) en het laatste glaciale maximum tot het Holoceen.

Hoofdstuk 6 analyseert het Midden en Buiten deel van het Belgisch deel van de Noordzee. Dit gebied wordt gekenmerkt door enkele karakteristieke noord tot zuidzuidwest georiënteerde begraven langgerekte en met elkaar verbonden paleodepressies, die de oudere laatste interglaciale sedimenten verwijderd hebben. Ondanks dat deze erosieve/depositionele structuren bestudeerd zijn in voorgaande studies is hun oorsprong en relatie met het omliggende landschap niet gekend. Hier bestuderen we het ontstaansmechanisme van deze paleodepressies door het combineren van een grote bestaande dataset sedimentkeren en een dicht seismisch reflectienetwerk en vergelijken we onze interpretaties met bestaande paleoklimatologishe/paleogeografsche reconstructies van noordwest Europa. De sedimentologische samenstelling van de paleodepressies toont aan dat in het Buiten Deel van het Belgisch deel van de Noordzee een ondiep mariene baai aanwezig was ten tijde van het vroege laatste interglaciaal en dat deze geconditioneerd werd door de zuidelijke rivierterras van de Maas Rivier van het voorlaatste glaciaal. Deze mariene sedimenten zijn later deels herwerkt en verwijderd tijdens de vorming van de langgerekte paleodepressies, die hier geïnterpreteerd worden als erosieve landschapselementen gevormd in de overstromingsvlakte van het Axiale (Buiten) Kanaal, een centrale riviervallei waarin vele noordwest Europese rivieren stromen ten tijde van lage zeespiegelstanden. Vermoed wordt dat de langgerekte depressies gevormd worden tijdens catastrofale glaciale meer uitbarstingen die ijsbergen en puin transporteren naar de Straat van Dover.

De **finale doelstelling** van dit onderzoek was trachten te begrijpen en visualiseren van het bewaringspotentieel van het onderwater erfgoed op het Belgisch deel van de Noordzee (Hoofdstukken 7 en 8). De gedetailleerde kennis van de distributie en de vormingsprocessen van de bestudeerde afzettingsmilieus laat toe om, voor de eerste keer, het bewaringspotentieel van het onderwater erfgoed in beeld te brengen (Hoofdstuk 7). Het Belgisch deel van de Noordzee vertoont enkele geomorfologische affiniteiten met Vlaanderen: 1) beide gebieden beschikken over een dun en gefragmenteerd Quartair sedimentdek waardoor het Paleogeen substraat soms dagzoomt; 2) platformen, zoals de Kempen, lijken sterk op het Binnen, Midden en Buiten Deel van het Belgisch deel van de Noordzee; en 3) de Vlaamse Vallei is verbonden met het stroomafwaartse paleo-Schelde Vallei. Deze geologische affiniteiten tussen beide gebieden laten toe om de bestaande kennis van Vlaanderen met betrekking tot archeologie en paleontologie te extrapoleren naar het Belgische deel van de Noordzee.

Het finale resultaat van deze studie zijn twee kaarten die het bewaringspotentieel van het onderwatererfgoed in beeld brengen, een voor archeologie en een voor paleontologie (Hoofdstuk 8). De archeologische potentieelkaart kan gebruikt worden om gebieden aan te duiden waar archeologische materiaal uit een bepaalde periode bewaard gebleven kan zijn. De paleontologische potentieelkaart toont op zijn beurt aan waar bepaalde fauna's bewaard gebleven zijn. Beide kaarten zijn belangrijke werktuigen die toegepast kunnen worden tot het ontwikkelen van een duurzaam beleid voor onderwater erfgoed in de nabije toekomst. Op deze manier kunnen wetenschap, industrie en beleidsvorming beter op elkaar afgestemd worden en zo tot voordeel van eenieder dienen.

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To the rest of RCMG: thanks for the lovely moments and small talk!

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I would like to thank VLIZ for their assistance during this research and their help in the development of an animation video I was awarded through the VLIZ Communication Prize. The video can be accessed through:

- YouTube-NL
  https://www.youtube.com/watch?v=wOg428iZJzA
- YouTube-EN
  https://www.youtube.com/watch?v=wOg428iZJzA&t=46s
- VLIZ videogalerij-NL
  http://www.vliz.be/nl/multimedia/videogalerij?album=5237&pic=130381

- VLIZ video gallery-EN
  http://www.vliz.be/nl/multimedia/videogalerij?album=5237&pic=130380

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Foreword

Chapters 4, 5 and 6 have been respectively published, and submitted for publication in peer-reviewed journals. I would like to apologise to the reader for the redundancies among the chapters due to this kind of presentation.

All figures are projected in UTM zone 31N/WGS84.

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List of Abbreviations

APM	archaeological potential map
BCS	Belgian Continental Shelf
BIIS	British-Irish Ice Sheet
DCS	Dutch Continental Shelf
DCSM	depth-converted structure map
DEM	digital elevation model
GLOF	Glacial Lake Outburst Flood
FCS	French Continental Shelf
FIS	Fennoscandian Ice Sheet
LAT	lowest astronomical tide
LEZ	low energy zone
LGM	Last Glacial Maximum
MIS	marine isotope stage
msl	mean sea level
OSL	optically stimulated luminiscence
PAZ	pollen assemblage zone
PPM	palaeontological potential map
TAW	tweede algemene waterpassing
UKCS	United Kingdom Continental Shelf

1 INTRODUCTION

1.1 Background

The Quaternary stratigraphic record of and the boundary between areas of deposition and areas of erosion within the southern North Sea is being controlled by the position of its basin shoulder (Figure 1.1). The Belgian Continental Shelf is located on this basin shoulder (Hijma et al., 2012, Knox et al., 2010) and is therefore characterised by a relatively thin and fragmented Quaternary stratigraphic record. By consequence, the Belgian shelf may be a critical area to connect the area of deposition in the north (southern North Sea as a whole) to the area of erosion to the south (English Channel and the Dover Strait).

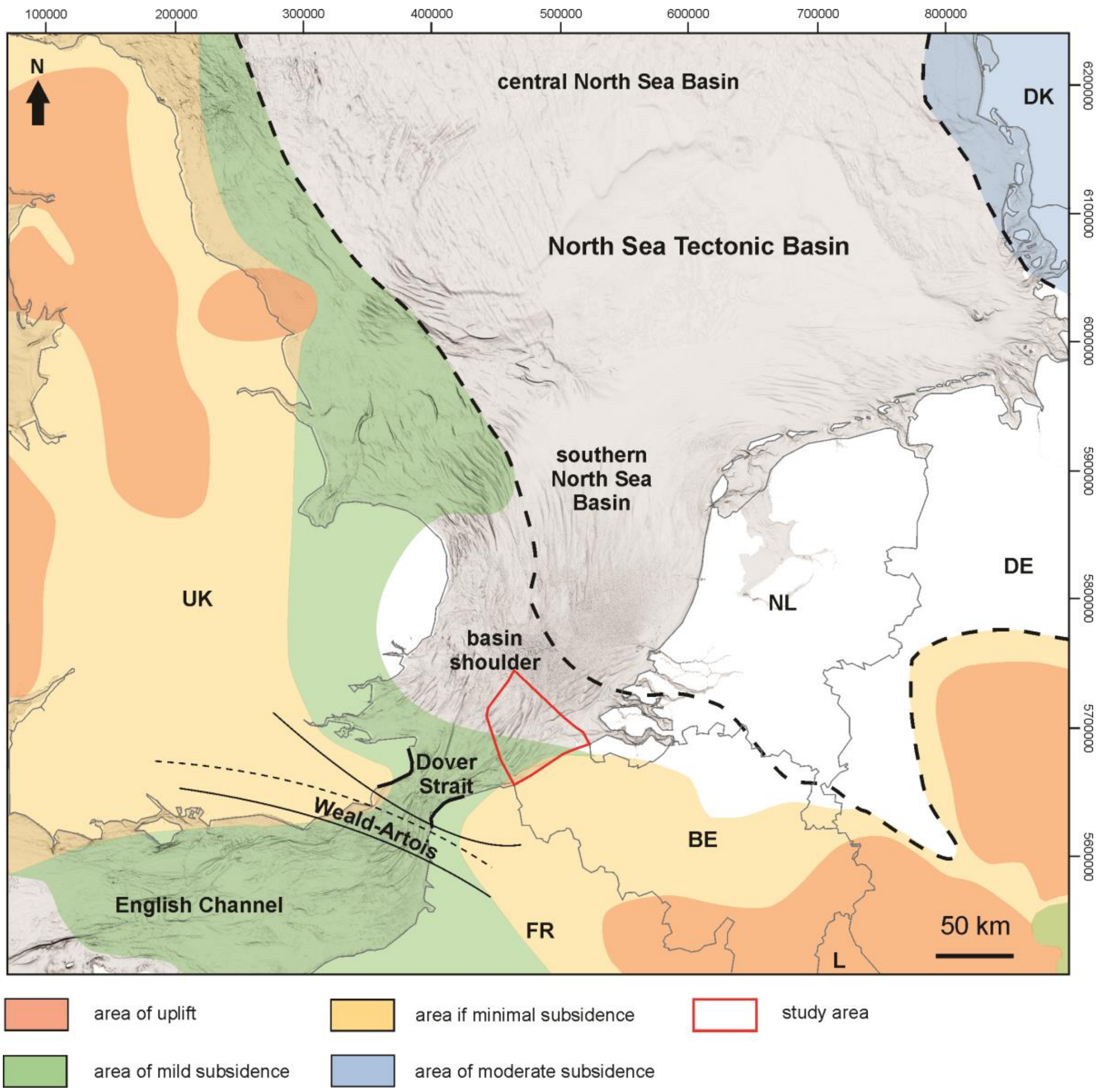


Figure 1.1 Location of the study area and the wider southern north Sea Basin and English Channel with indication of Paleogene areas of uplift and subsidence. The dashed line indicates the North Sea Tectonic Basin. The basin shoulder is located between East Anglia and Belgium-southwest Netherlands. Modified from Hijma et al., 2012 and Knox et al., 2010. For this and next figures projections are in UTM zone 31N/WGS84.

Since the 1960s the geology of the southern North Sea has been examined mainly within the boundaries of the surrounding nations of the UK and the Netherlands with a particular focus on the English Channel, the Dutch nearshore areas and the areas further to the north of interest for natural gas reserves. This has resulted in a comprehensive understanding of the Middle-Late Pleistocene geological history of the entire area starting with the initial opening of the Dover Strait and the carving of the Axial Channel and its tributaries (Thames, Rhine, Meuse, etc.: see García-Moreno (2017) for a recent overview). Belgium however, lies just south of the visible geomorphological boundary of this major palaeoriver system. Consequently, many researchers assumed that the palaeolandscape of the Belgian shelf did not have any affinity with this dynamic fluvial landscape. More so, these researchers assumed that most evidence has been removed from the geological record during the last two transgressions of the Eemian and Holocene (Balson et al., 1991; Le Bot et al., 2003; Liu 1993; Mathys, 2009; Mostaert et al., 1989). The final result is a relatively thin and complex, fragmented geological architecture that makes cross-correlation of deposits over larger areas very difficult.

Within these sediments are buried evidence of the biosphere in the form of animal and human bone material, plant fragments but also rare human artefacts (Peeters et al., 2009; Bicket and Tizzard, 2015; Hijma et al., 2012). Most of this evidence is recovered from the seafloor by industrial activities such as fishing, aggregate extraction and dredging. During the past years the geological context of these finds has become much better understood within UK and Dutch waters. In Belgium, similar finds of mammals are also found. However, the limited knowledge of its offshore Quaternary geology limits our understanding of these finds. More so, it is often these finds that provide information about these prehistoric landscapes, while it should be the other way around. The tons of bone material that have been brought ashore over the past few decades from the North Sea seafloor testify that many of these submerged landscapes are still preserved and are located close to the seafloor. Herein lies a paradox: It are these invasive and often destructive activities that bring these bones to shore without having any knowledge of them. This is especially true in Belgium where the Quaternary cover is often less than 10 m thick.

1.2 Dissertation aims and rationale

The present study first focuses on the buried prehistoric landscapes of the Belgian continental shelf and determines their relationship with the regional-scale palaeogeography of the southern North Sea Basin. In a second stage their preservation potential will be determined in relation to post-depositional processes (both local and regional), which is followed by a discussion of what this implies for submerged heritage (i.e. archaeological and palaeontological indicators) preserved within. In doing so, a series of local and regional palaeogeographic maps are made of the southern North Sea Basin, which will provide important clues to the geological context of already discovered submerged heritage and as yet undiscovered finds. In the future, these results should help pinpoint areas that have a specific preservation potential for submerged heritage ranging from absent to high. Consecutively, this strategy provides a valuable advisory tool for offshore commercial-industrial activities and the long-term management of its natural resources.

The first aim of this dissertation is to develop a new stratigraphic framework for the Belgian shelf environment with specific focus on the Quaternary. This process requires a high-resolution reconstruction of its base level (equivalent to the top of the Paleogene). This is a necessary step if we want to understand where prehistoric landscapes on the shelf environment are preserved. The most recent geological investigation dates back to Mathys in 2009. The study of Mathys (2009) mainly focused on the inner sector of the shelf and in particular on the Eemian infilling of the incised Ostend Valley system. However, many more geomorphological expressions of valley systems are present on the outer shelf environment. Each of these valley systems should be connected to the central Axial Channel within the central southern North Sea. Mathys (2009) understood this but could not clearly make these connections because at that time there was still a lack in high-resolution data in key areas which hindered the reconstruction of the base of the Quaternary, the base-level into which many of these valleys are cut. This task will be performed by integrating archived and new seismic reflection data, archived sediment cores/vibrocores and bathymetric data combined with information from literature. From this newly achieved knowledge, a palaeogeographic map series of the Belgian prehistoric landscapes will be constructed.

In a next step we will compare the results of this map series to those observed on the neighbouring shelves of the UK, the Netherlands and France. This is a necessary comparison step if we want to understand the relationship between the Belgian landscape development to that of the wider region such as the Axial Channel river system.

It is common knowledge that the development of the Quaternary landscape in northwest Europe was fundamentally driven by the expansion and retreat of ice masses extending from Scotland and Scandinavia to the centre of the North Sea (Clark et al., 2012; Ehlers et al., 2011; Lambeck et al., 2010) and by the presence and subsequent disappearance of a topographic barrier between continental Europe and Britain (see García-Moreno, 2017; Gibbard and Cohen, 2015; Gibbard and Lautridou, 2003; Gibbard and Lewin, 2016; Gupta et al., 2007, 2017; Meijer and Preece, 1995). Indeed, these northern ice masses and a southern topographic barrier fundamentally control the development of the northwest European drainage systems (including the now submerged Axial Channel) during lowstand conditions, which in turn had a major impact on the palaeogeographic evolution of the entire region (in this study we assume this includes the Belgian shelf). The structural degradation and the enigmatic removal of the Dover Strait topographic barrier long overshadowed the discussion of the Middle-Late Pleistocene palaeogeography of northwest Europe (Busschers et al., 2008; García-Moreno, 2017; Gibbard, 1988; Gupta et al., 2007, 2017; Mellett, 2012; Murton and Murton, 2012; Smith, 1985; Westaway and Bridgland, 2010). The first stage in the removal of the barrier was considered to have taken place during the Elsterian glaciation (MIS 12; Gibbard, 1988, 1995). Only recently, this first stage in the breaching of the land bridge has become better constrained and was estimated to have occurred ca. 450 ka (Gupta et al., 2017). However, the hypothesised presence of a remnant land bridge during the Saalian glaciation (MIS 6; Busschers et al., 2008; Gibbard and Cohen, 2015; Hijma et al., 2012; Murton and Murton, 2012) and the marine interactions across this structurally degraded topographic barrier during the following Holsteinian and Eemian interglacials (MIS 11 and MIS 5e; Bogemans et al., 2016; Meijer and Cleveringa, 2009; Meijer and Preece, 1995, 2000; Peeters et al., 2016) are still not well understood.

Furthermore, glacio-isostatic flexural adjustment across these palaeolandscapes (Figure 1.2) during and after the Saalian and Weichselian glaciations (but also during the Elsterian: see Moreau et al., 2015) and their impact on drainage systems and subsequent sea-level rise has only been tested quite recently (Cohen et al., 2014; Lambeck et al., 2006; Long et al., 2015; Peeters et al., 2016). Indeed, several hypotheses exist on the various pathways the major palaeodrainage systems may have taken during these periods (e.g. García-Moreno, 2017; Hijma et al., 2012; Patton et al., 2017; Sejrup et al., 2016; Toucanne et al., 2015). However, few of these are based on actual geomorphological evidence, and those that are, are either too local or based only on the seafloor morphology or on the morphology of buried palaeovalleys (e.g. Henriet and Moor, 1989; Liu et al., 1992; Mostaert et al., 1989). Other studies focus less on geomorphology and provide only sedimentary evidence (e.g. Busschers et al., 2007, 2008, Peeters et al., 2015, 2016). Today, most existing regional hypotheses about the different pathways the palaeorivers followed during the Middle-Late Pleistocene glaciations are based on interpolations of onshore palaeodrainage reconstructions (e.g. Busschers et al., 2007, 2008; Lewis et al., 2004; Peeters et al., 2015, 2016), combined with local offshore geomorphological/geological studies further downstream (e.g. Liu et al., 1992; Mathys, 2009) and with reconstructions of ice mass positions during those periods (e.g. Carr et al., 2006; Clark et al., 2012; Ehlers et al., 2011; Lee et al., 2012; Sejrup et al., 2009). Only recently, a first comprehensive regional-scale geomorphological study combining the geomorphology and crosscutting relationships of buried and exposed palaeovalleys and palaeodepressions in the southern North Sea and Dover Strait has been conducted (García-Moreno, 2017). García-Moreno (2017) however, still lacked palaeoenvironmental control from sedimentary archives. He also focuses on the maximum ice extent of the Elsterian, Saalian and Weichselian glaciations and did not fully take into account palaeogeographic changes during interglacials and intermediate ice-sheet advances, like for example during the Early Pleniglacial (MIS 4). The results of the paleogeographic map series will provide information on these geological processes and how they influence the Belgian shelf environment.

Even though the Belgian shelf is ideally positioned between the erosional region of the Dover Strait and the depositional southern North Sea Basin it was considered as a shelf area with very little accommodation space to preserve sediments and its sedimentary record was therefore not believed to be very valuable. Additionally, the area is located south of the so-called Axial Channel trunk valley, the major drainage route from the southern North Sea Basin to the Dover Strait. More so, it is considered that terrestrial depositional environments on the Belgian shelf were recycled during subsequent marine transgressions obliterating all valuable information into a reworked deposit in which the original signal is completely lost (Le Bot et al., 2003; Mathys, 2009). The increased density of new very-high seismic reflection data, the increased amounts of archived sediment cores-vibrocores on the Belgian shelf, and a much better knowledge of its fragmented Quaternary sediments now make it possible to incorporate this new information into a new regional palaeogeographic reconstruction of the southern North Sea Basin. Additionally, the new evidence from the Belgian shelf will make it possible to test the current landscape evolution hypotheses of the southern North Sea Basin and to what extent these need to be changed.

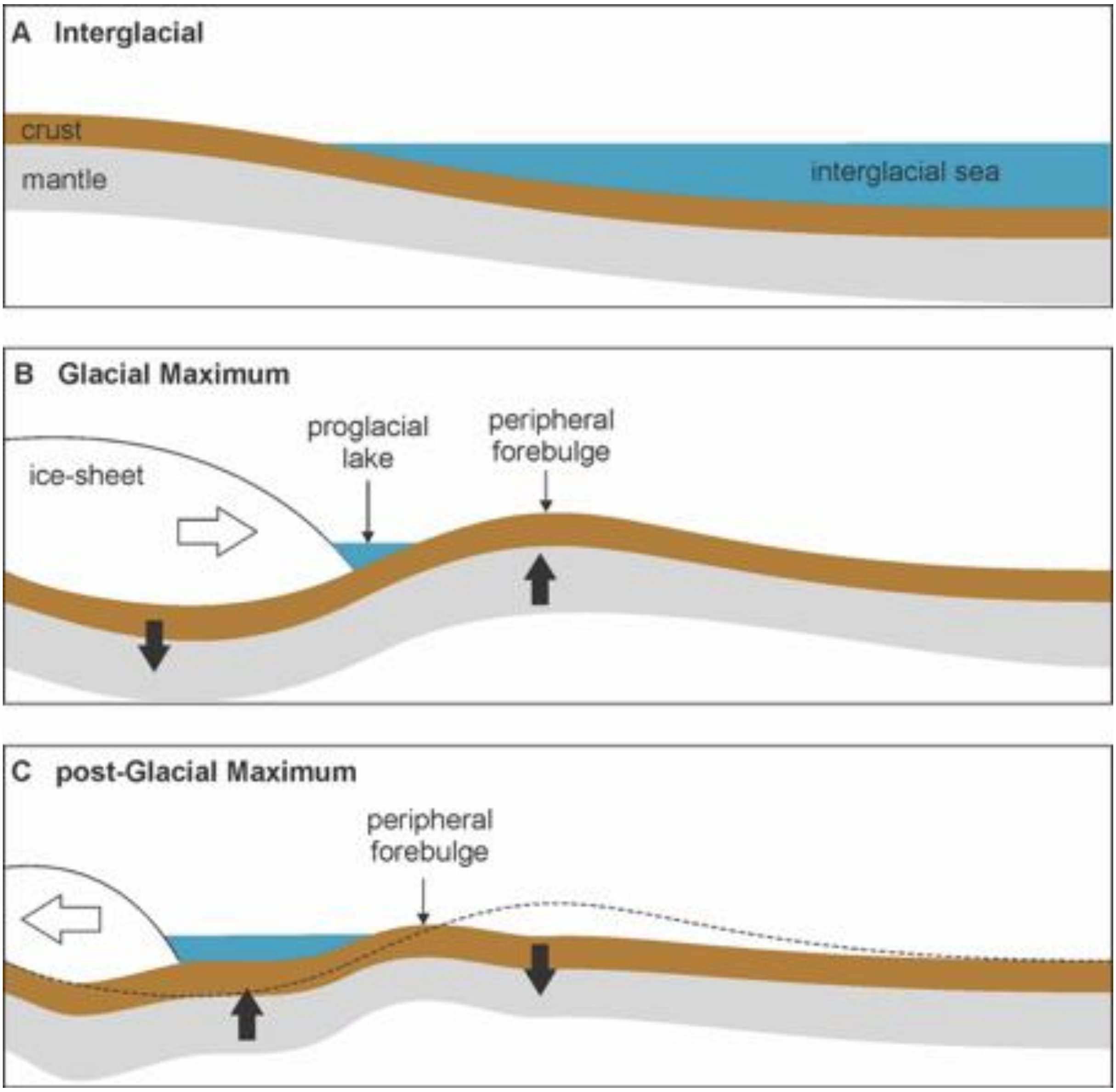


Figure 1.2 Conceptual model of glacio-isostasy for the North Sea Basin. A) During an interglacial most of the North Sea Basin is inundated by the sea as a result of high sea levels. B) When climate deteriorates sea levels decrease and water is locked in ice-sheets. As the ice-sheet grows and reaches a maximum size during the glacial maximum its weight causes the crust and mantle to subside. Mantle material flowing away from the ice-sheet causes the crust to rise and form a peripheral forebulge at a certain distance from the ice-sheet rim. Between the ice-sheet and the peripheral forebulge a proglacial lake may form. C) Climate improvement results in ice-sheet disintegration resulting in peripheral forebulge relaxation as mantle material can flow back beneath the now ice-sheet free zones. At the same time, when the ice-sheet mass remains large enough (especially during initial deglaciation or ice-sheet advance-retreat cycles) the forebulge may migrate with the ice-sheet rim. Note that the degree of glacio-isostatic adjustment in the landscape, and the rate at which it occurred, is related to differing ice-sheet volumes.

The establishment of a relationship between the local Belgian and regional southern North Sea prehistoric landscapes will help identify the processes that are responsible for their preservation or removal since the opening of the Dover Strait. During glacial sea-level lowering the southern North Sea and English Channel shelves emerged and the Axial Channel fluvial system became active. This process started net fluvial incision across the entire northwest European shelf and transported large amounts of sediments downstream. This and the combined impact on the antecedent palaeotopography of ice mass related processes (glacio-isostasy, rerouting, variable discharge) shaped the landscape for the following interglacial sea-level rise, during which the landscape inundated first, during which palaeocoastlines were formed and during which sediments were redeposited.

Once these processes have been established it is possible to determine the final aim of this dissertation: understand why submerged heritage is preserved within prehistoric landscapes and why we encounter it. For decades, tonnes of palaeontological bone material and to a smaller extent archaeological artefacts (i.e. submerged heritage) have been uncovered from the southern North Sea seafloor during commercial-industrial activities (Hijma et al., 2012; Hublin et al., 2009; Peeters et al., 2009). On the Belgian shelf hardly any evidence of submerged heritage is known today, except for some stray or accidental finds that are focused on the near-coastal and intertidal area, in the vicinity of navigation channels. In the UK and the Netherlands, a good collaboration exists between the offshore industry and the fishing and scientific communities. In Belgium such a synergy still appears to be trailing for some reason. The extensive exploitation of the Belgian shelf by the offshore industry, such as wind farm and submarine cable constructions, may be devastating to the nature of this submerged heritage. More so, due to the thin and fragmented Quaternary cover and present-day tidal forces, evidence of submerged heritage may resurface and be removed from its primary burial location, thereby losing its geological context, or even worse be destroyed. For that reason it is necessary to understand why certain prehistoric landscapes and geomorphological landforms protect depositional environments and submerged heritage can be preserved and why at other locations it is not. This will help both the synergy between the industry and policy makers and the scientific community and help improve the planning stage of offshore activities where necessary.

1.3 Palaeogeographic studies

As mentioned above the study area is positioned on the North Sea Basin shoulder between areas of deposition to the north and areas of erosion to its southwest. Both areas were part of recent palaeogeographic studies for the Middle-Late Pleistocene period. García-Moreno (2017) studied the palaeogeographic evolution of the southern North Sea as a whole (including parts of the our study area) based on the principles of geomorphology and relative chronology, while Mellett (2012) focussed on the eastern English Channel and the preservation of sediment bodies using geomorphology, sedimentology and absolute geochronology. García-Moreno (2017) provides a general overview of the evolution of the major palaeovalley systems in the southern North Sea during glacial maxima each linked to the central valley system named the Lobourg-Axial Channel. The results of García-Moreno (2017) will be integrated in this study, however here we will also focus of intermediary ice-advances such as that of the Middle Pleniglacial and study the impact of glacio-isostatic adjustment on the landscape.

Mellett (2012) demonstrates that the erosional landscape of the Eastern English Channel is able to preserve a wide variety of sediment bodies over multiple glacial-interglacial cycles and that sediments of the entire terrestrial-marine continuum are able to withstand the erosive power of multiple interglacial sea-level rises. This was demonstrated by a landscape evolution model to demonstrate the preservation potential of coastal facies. The preservation of fluvial sediments was demonstrated by the presence of sediments pockets within a major palaeovalley system carved in chalk bedrock by catastrophic and/or normal processes related to the breaching of the Dover Strait. The results of Mellett (2012) are in contrast to the present geological evolution model of the Belgian shelf where it is assumed that one single sea-level rise can erode all previous sediments even though a similar tectonic setting is present.

The present study will present its own independent palaeogeographic map series of the Belgian shelf based on geomorphology, sedimentology, acoustic facies and geochronology. This local landscape evolution model will be compared to regional observations (i.e. the southern North Sea and English Channel) and present its own conclusions for landscape development. These results will then be used to visualise the preservation potential of submerged heritage in a shallow sediment-depleted shelf environment positioned on a basin shoulder. It is expected that the intermediary positon of our study area will present results complementary to both studies.

1.4 Dissertation outline

This dissertation comprises two manuscripts that have been accepted in international peer-reviewed journals (Chapter 4 and 5). Chapter 6 and 7 will be published at a later stage. Each chapter (4 to 8) addresses a different or a combination of research questions:

Chapter 2: *Geological-stratigraphical setting*

In Chapter 2 a brief and general overview of the geographic and geological setting of the North Sea Basin and the Belgian shelf is presented. This is followed by an introduction to the Middle-Late Pleistocene palaeogeographic evolution of northwest Europe. This chapter is concluded with a literature overview of the current understanding of the Quaternary geological history of the study area.

Chapter 3: *Geophysical data and data integration*

Reports on all the available datasets and the methods used for this dissertation.

Chapter 4: *A high-resolution depth-converted structure map for Top-Paleogene surface of the Belgian Continental Shelf*

Presents the results of the base on which the Quaternary deposits are preserved, formed and recycled (i.e. the top of the Paleogene substratum or pre-Quaternary surface). This high-resolution depth-converted structure map (more commonly known as a digital elevation map) has been connected to the onshore top of the Paleogene substratum. The chapter discusses briefly the evolution of this surface to its current state and the observed geomorphological features carved within this surface. As a result of these geomorphological features, the BCS and its coastal plain are divided into four separate areas that are closely linked with each other: The Belgian Coastal Plain and the Inner, Middle and Outer BCS.

Chapter 5: *A well-preserved Eemian incised-valley fill in the southern North Sea Basin, Belgian Continental Shelf-Coastal Plain: implications for northwest European landscape evolution*

This chapter describes the geological birth and demise of the palaeo Scheldt River in the Belgian Coastal Plain and the Inner-Middle shelf down to the Axial Channel trunk valley. The chapter describes which investigations have been conducted before within the study area and how these results, in conjunction with new palaeoenvironmental evidence, age control and landscape modelling, resulted in a new stratigraphic model of the palaeo-Scheldt River in this location and its implications for the palaeogeographic evolution of northwest Europe on a regional scale.

Chapter 6: *Evidence of Late Pleistocene coastal environments and GLOFs on the Middle-Outer Belgian Continental Shelf: implications for northwest European landscape evolution*

Here we describe the formation and geological evolution of the Middle-Outer shelf. Results show that this area is part of the southeast section of the Axial Channel trunk valley system and is the direct result of fluvial activity intimately linked to the evolution and oscillations of the FIS and BIIS. In addition, evidence of far-travelled erratic clasts, sourced from the Scottish Grampian Highlands and the British east coast, reveal a close relationship with the BIIS and its fast flowing North Sea Lobe. The final result is a palaeogeographic and clast transportation model supported by new stratigraphic evidence for this section of the BCS and its implications for the palaeogeographic evolution of northwest Europe.

Chapter 7: *Preservation potential assessment for submerged heritage on the Belgian Continental Shelf*

This chapter combines the results of the palaeogeographic map series and stratigraphic model of Chapters 4 to 6 and applies it to the known submerged heritage previously encountered and expected in the near future for the entire study area. Linking the prehistoric landscapes, geomorphological landforms and their depositional environments with the location of the submerged heritage (i.e. geological context) provides clues to their origin and acknowledges, for the first time, its preservation potential from a geological point of view.

Chapter 8: *Profile type mapping: archaeological and palaeontological potential maps*

This chapter is an extension to Chapter 7 and visualises the different areas where certain archaeological and palaeontological material of a specific age can be expected based on the stratigraphic model presented in Chapters 5 and 6. This visualisation is based on the principle of profile type mapping, a technique widely used in Flanders (Belgium) that visualises the three-dimensional Quaternary sedimentary architecture on a two-dimensional scale.

Chapter 9: *Discussion, conclusions and outlook*

The final chapter of this dissertation presents a discussion that attempts to answer the research questions as explained in Chapter 1 to their fullest extent by compiling the scientific results of Chapters 4 to 8. Finally, some conclusions, considerations and scientific reflections are given concerning the ramifications of this research and where future research should focus.

1.4. Author contribution and status of manuscripts

This thesis comprises four core chapters intended for publication in international peer reviewed journals. At the time of dissertation submission, the status of these manuscripts is as follows:

Chapter 4:

De Clercq M, Chademenos V, Van Lancker V, Missiaen T. 2016. A high-resolution DEM for the Top-Paleogene surface of the Belgian Continental Shelf. Journal of Maps 12: 1047–1054. DOI: 10.1080/17445647.2015.1117992

Author contribution:

- De Clercq M – Main author. Responsible for data collection, collation and interpretations and for writing the manuscript
- Chademenos V – Processing of the depth-converted structure map and merging of the datasets
- Missiaen T – Comprehensive discussion and detailed manuscript review
- Van Lancker V – Comprehensive discussion and detailed manuscript review

Submitted: 6th May 2015

Accepted: 5th November 2015

Available online: 14th December 2015

Chapter 5:

De Clercq M, Missiaen T, Wallinga J, Zurita Hurtado O, Versendaal A, Mathys M, De Batist M. 2018. A well-preserved Eemian incised-valley fill in the southern North Sea Basin, Belgian Continental Shelf-Coastal Plain: implications for northwest European landscape evolution. Earth Surface Processes and Landforms. DOI: 10.1002/esp.4365

Author contribution:

- De Clercq M – Main author. Responsible for data collection, collation and interpretations and for writing the manuscript.
- Missiaen T – Comprehensive discussion and detailed manuscript review
- Wallinga J – OSL-dating and assistance with data analysis and manuscript review
- Zurita Hurtado O – Seismic reflection data processing
- Versendaal A – OSL-dating and assistance with data analysis
- Mathys M – Manuscript review
- De Batist M – Comprehensive discussion and detailed manuscript review

Submitted: 16th October 2017

Accepted: 6th February 2018

Available online: 21st February 2018

Chapter 6:

De Clercq M, Garcia-Moreno D, Missiaen T, Zurita Hurtado O, De Batist M. Submitted. Evidence of Late Pleistocene coastal environments and GLOFs on the Middle-Outer Belgian Continental Shelf: implications for northwest European landscape evolution.

Author contribution:

- De Clercq M – Main author. Responsible for data collection, collation and interpretations and for writing the manuscript
- Garcia-Moreno D – Comprehensive discussion and detailed manuscript review
- Missiaen T – Detailed manuscript review
- Zurita Hurtado O – Seismic reflection data processing
- De Batist M – Comprehensive discussion and detailed manuscript review

In preperation

Chapter 7:

De Clercq M, Pieters M, Langeveld B, Post K, Mol D, De Batist M, Missiaen T. In preperation. A first preservation potential assessment for submerged heritage on the Belgian Continental Shelf, southern North Sea.

Author contribution:

- De Clercq M – Main author. Responsible for data collection, collation and interpretations and for writing the manuscript
- Pieters M – Comprehensive discussion and detailed manuscript review
- Langeveld B – Comprehensive discussion and detailed manuscript review
- Post K – Comprehensive discussion and detailed manuscript review
- Mol D – Comprehensive discussion and detailed manuscript review
- De Batist M – Detailed manuscript review
- Missiaen T – Comprehensive discussion and detailed manuscript review

In preparation.

2 GEOLOGICAL-STRATIGRAPHICAL SETTING

2.1 North Sea geological setting

The North Sea is a continental shelf sea located between northwest continental Europe, the Scandinavian Peninsula and Great Britain (Figure 2.1). This sea enters the Atlantic Ocean through the Dover Strait/English Channel to the southwest, and via the Norwegian Sea (between the United Kingdom and Norway) to the north. It is commonly sub-divided in three parts; i.e. the southern, central and northern North Sea.

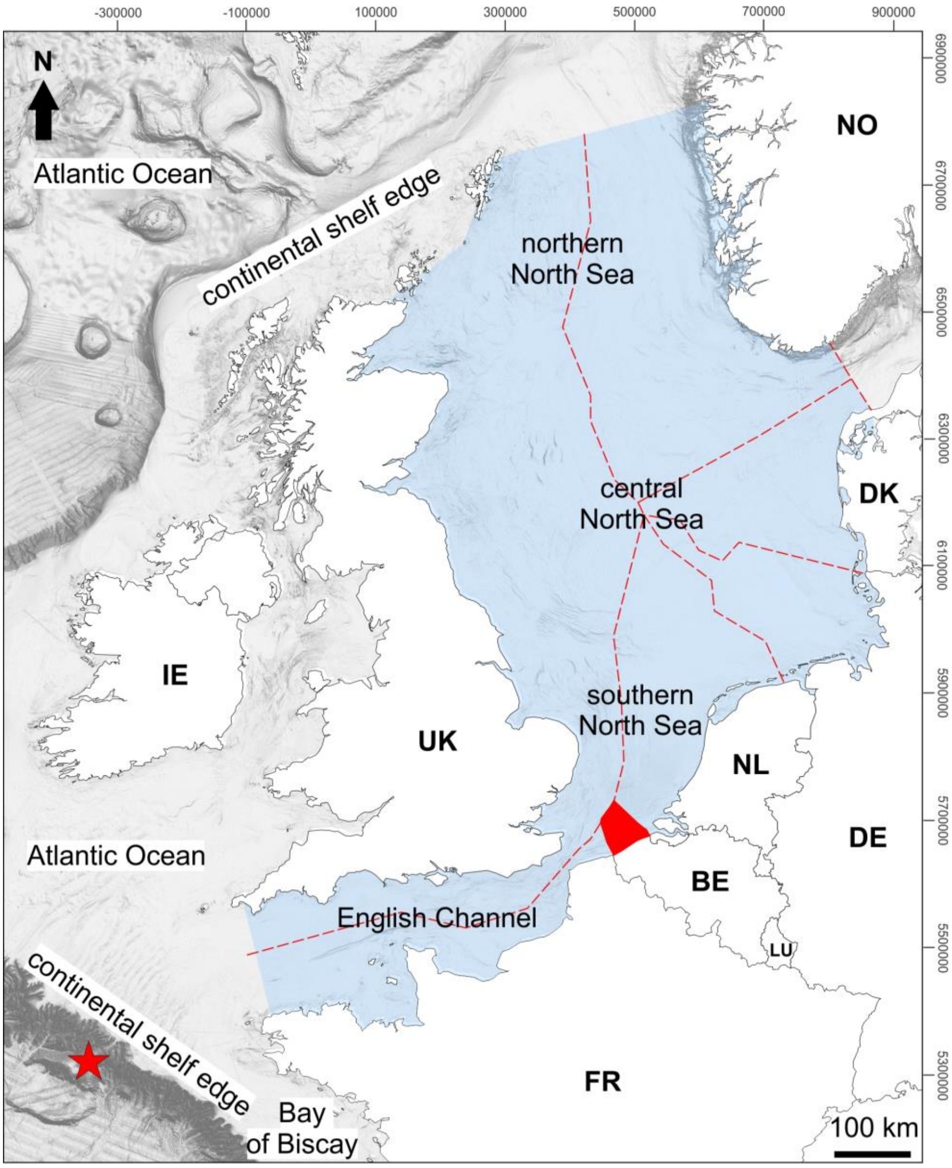


Figure 2.1 Geographical situation of the North Sea and English Channel. Offshore national boundaries are indicated (red lines). The study area, the Belgian Continental Shelf, is indicated in red. The red star indicates the study area of Toucanne et al. (2009a, 2009b, 2010, 2015).

The BCS is a key area in the southern North Sea that connects the North Sea shelf with the English Channel shelf (Figure 2.1) and is featured in part-evidence, part inference-based studies on Middle-Late Pleistocene 'proglacial drainage' and 'catastrophic land bridge removal' theories (e.g. Cohen et al., 2017; García-Moreno, 2017; Gibbard, 1995; Gupta et al., 2007, 2017; Hijma et al., 2012). During periods of maximum glaciation proglacial lakes are envisaged in the southern North Sea, blocked by an ice-sheet to the north and upland topography (i.e. a land bridge) between continental Europe and England to the south (Gibbard, 1995; Hijma et al., 2012; Murton and Murton, 2012; Toucanne et al., 2009). In the present study, we will investigate the Quaternary stratigraphy and geomorphology of the base of the Quaternary in relation to the southern North Sea landscape evolution. This will form the basis to assess for the first time the preservation potential of archaeological and palaeontological bone material of the Quaternary deposits on the BCS. In this chapter a brief summary is presented concerning the geology of the southern North Sea and the BCS and archaeology in northwest Europe.

2.1.1 Northwest European chronostratigraphy

Because we introduce and define various geomorphological features and depositional environments within the following chapters of this dissertation, it is necessary to provide an introduction to the northwest European chronostratigraphy for the various periods used within this dissertation. In marine geology, it is widely accepted that Marine Isotope Stages (MIS) are used to define the interval of time during which a geological/climatic event took place. This timescale is deduced from oxygen isotope data from deep-sea core samples, which reflect changes in the continental ice-volume over time. Stages with even numbers are characterised by high $^{16}O/^{18}O$ ratios and represent cold glacial periods; whereas, odd-numbered stages are characterised by low $^{16}O/^{18}O$ ratios and represent warm interglacial intervals (Lisiecki and Raymo, 2005). MIS are therefore typically used to define the duration of the various Quaternary glacial and interglacial stages and substages. The MIS timescale thus defines cold and warm intervals at a global scale. However, the characteristics of the various Quaternary glacial-interglacial cycles are usually based on local, land-based studies. This has resulted in the use of completely different nomenclatures to refer to those cycles in different parts of the world. For instance, MIS 6 corresponds to the glaciation that took place between ca. 200 and 130 ka (i.e. thousands of years ago). That glaciation is known as “Saalian” in northwest Europe, “Wisconsin” in North America and “Wolstonian” in Great Britain (Cohen and Gibbard, 2016). Moreover, due to local effects, the same glaciation did not start or end at the same time for each of these areas. In this dissertation we only use the northwest European nomenclature. Another issue that may lead to confusion is that the glacial maximum – defined, as the maximum extent of the ice-sheets during a glacial period – of each of the Pleistocene glaciations is diachronous.

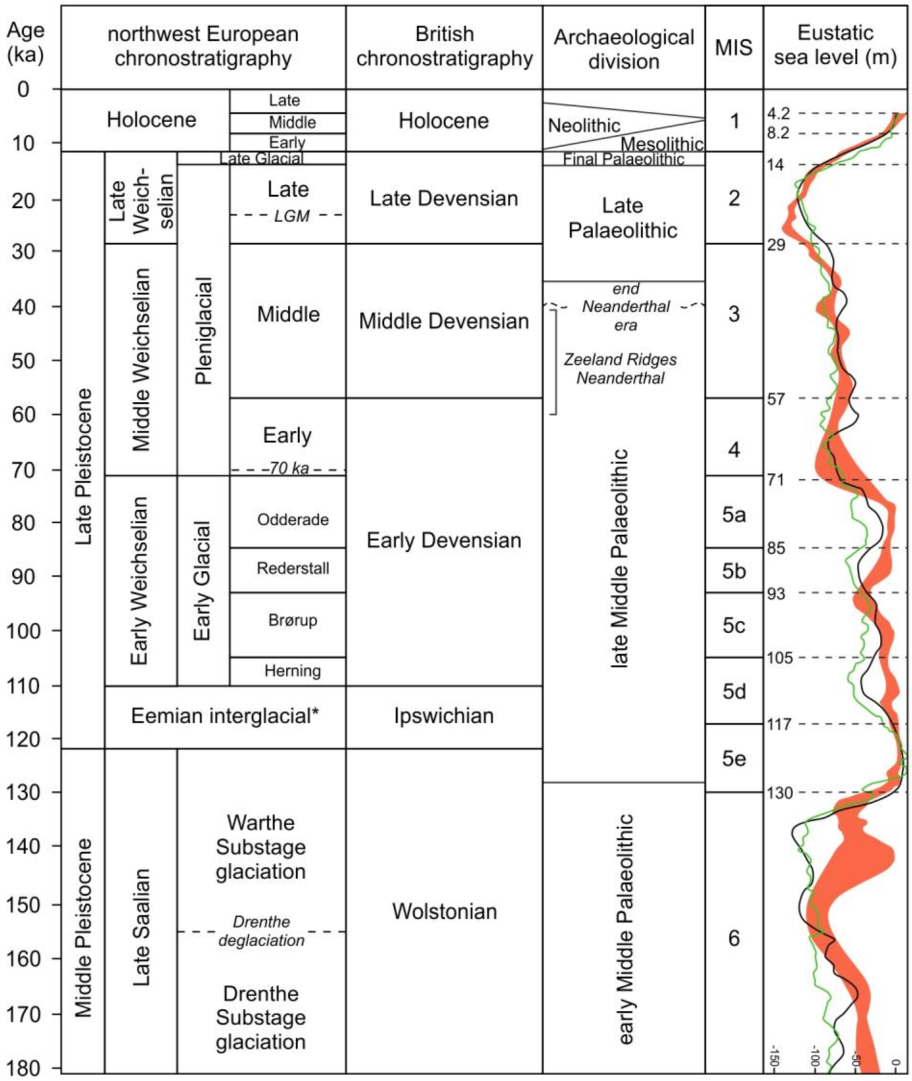


Figure 2.2 Northwest European chronostratigraphic subdivision of the late Middle and Late Pleistocene (after Peeters et al., 2015) based on van Huissteden et al. (2001) and Busschers et al. (2007, 2008), with correlations to the marine isotope record (Bassinot et al., 1994). British chronostratigraphy after Gibbard and Cohen (2016). Sea level curves from Waelbroeck et al. (2002; black line), Lisiecki and Raymo (2005; green line) and Medina-Elizalde (2013; red zone) are presented. The sea-level curve from Medina-Elizalde (2013) represents a Monte Carlo 95% confidence interval. LGM: Last Glacial Maximum. *Recently, new insights from Sier et al. (2015) placed the onset of the Eemian in northwest Europe at ca. 121 ka, thereby post-dating the MIS 6/5e transition by ca. 10 ka.

In order to avoid misunderstandings, we show in Figure 2.2 the various land-based chronostratigraphic terminologies used to define major Quaternary climatic cycles in northwest Europe and their full, partial or approximate equivalences with the MIS, absolute timing and the Quaternary time scale. Figure 2.2 only includes glacial and interglacial intervals for the timeframe during which the erosional/depositional features discussed in this dissertation were purportedly formed; i.e. from the Late Saalian to the present. Note that, in this dissertation, we use the various chronologic terms introduced in Figure 2.2 as synonyms unless otherwise defined.

2.1.2 Tectonic setting and stratigraphic architecture

The basement of the southern North Sea Basin was formed during the Precambrian and Lower Palaeozoic and has since then experienced multiple phases of rifting (e.g. Boulvain and Vandenberghe, 2018; Cameron et al., 1993; Flemming et al., 2017; Ziegler and Louwerens, 1979). The Mesozoic was characterised by rifting and basin inversion episodes, in which the North Sea first adopted its present north-northwest to south-southeast alignment (e.g. the North Sea Tectonic Basin in Figure 1.1).

By the Late Middle Jurassic an extensional episode occurred that lasted until the Middle Cretaceous. This subsidence episode enabled the deposition of Early Jurassic and Late Cretaceous marine shales, limestone, calcareous mudstones and sandstones (Figure 2.3). This crustal extension phase ended in the Middle Cretaceous, but the regional thermal anomaly that it generated continued to exist for a prolonged period of time, inducing subsidence. Notably, a thick (up to 1500 m) layer of chalk accumulated in Late Cretaceous times. Chalk deposition was interrupted in the Paleogene by tectonic uplift generated by the continental convergence during the Alpine Orogeny (Cameron et al., 1993). This collision resulted in a basin inversion, which induced basin uplift and temporary emergence of the southern North Sea Basin, resulting in an erosional phase. Today, some of these sedimentary formations outcrop in the southern North Sea Basin (Figure 2.3).

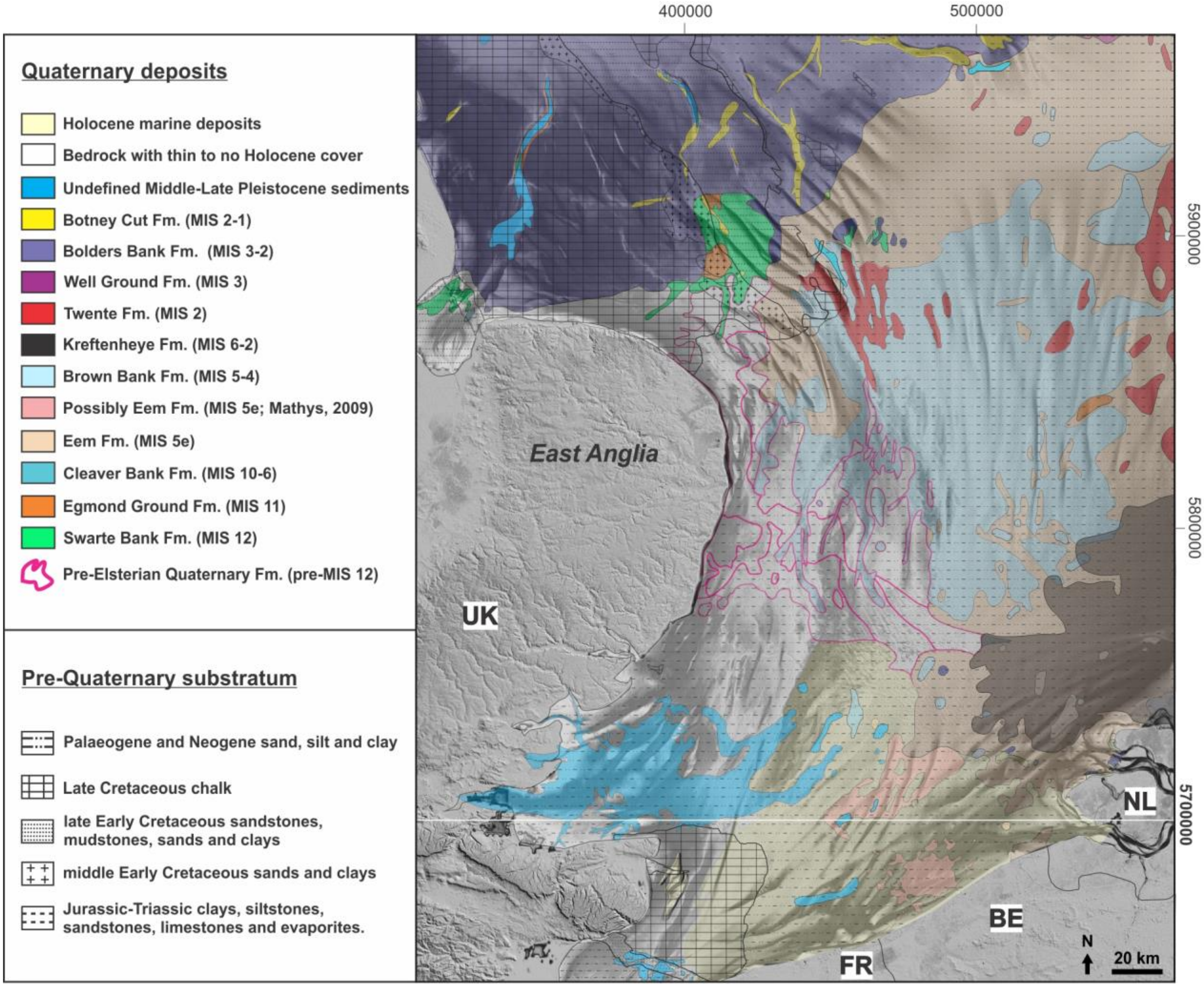


Figure 2.3 Geological map of the southern North Sea Basin (modified from García-Moreno, 2017).

Since Quaternary times the southern North Sea Basin is positioned close to and on the North Sea Basin margins. Therefore, the area appears tectonically stable and is marked by near-zero net subsidence/uplift (Cohen et al., 2014). These areas form the hinge zone of the basin. The hinge zone has received much sediment; delivered by rivers originating in upstream areas of uplift such as the Rhine and Meuse, and as a result a patchy, dissected and terraced landscape was formed with a relatively complete Quaternary stratigraphic record. Here, pre-Elsterian to Holocene sediments are preserved from various depositional environments (marine, periglacial, fluvial, aeolian, lacustrine, etc.; for an overview see Stoker et al., 2011).

More to the south, between Belgium-East Anglia and the Dover Strait, much less sediments were delivered to the shelf. Here, a relatively thin and incomplete Quaternary stratigraphic record is preserved in a dissected and terraced landscape dominated by erosion. Three main sedimentary units are identified (Figure 2.3): an undefined Middle-Upper Pleistocene unit; Eemian sediments linked to the Eem Formation; and Holocene marine deposits (see Section 2.2

for more detail). In other areas no Quaternary sediments are present and chalk bedrock or Paleogene substrate is exposed at the seafloor.

2.1.3 Pleistocene geological history

During the Early Pleistocene, Britain was connected to continental Europe, even during interglacial high-stand periods, via the extensive Eridanos-Rhine-Meuse River delta in the southern North Sea Basin and the Weald-Artois anticline at the Dover Strait (Funnell, 1995; see Figure 2.4). This land bridge was partially removed at some point during the Elsterian (ca. 450 ka) through erosion of the chalk bedrock at the Dover Strait (Figure 2.4; Collier et al., 2015; García-Moreno, 2017; Gibbard, 1995; Gupta et al., 2017). The breaching of this land bridge was one of the most significant landscape changes taking place during the Middle-Late Pleistocene history of northwest Europe and started a process that finally resulted in the insulation of Britain as an island from the European mainland (see Meijer and Preece, 1995 for a review). Such a major palaeogeographic reorganisation had a profound impact on the migration of flora and fauna, including early humans, across northwest Europe between glacial-interglacial sea-level cycles (Cohen et al., 2014; Hijma et al., 2012; Meijer and Preece, 1995). The timing of initial breach has been closely linked to glacial maxima when sea level was at lowstand and ice margins were at their maximum limits (García-Moreno, 2017; Gibbard, 1988; Gupta et al., 2017).

Commonly an Elsterian age is proposed for the main breaching event (Figure 2.5), as this stage saw the first major extension and coalescence of the FIS and BIIS (Ehlers et al., 2011; Figure 2.6 A) that completely blocked drainage into the North Atlantic Ocean via the central and northern North Sea. In response to this ice-sheet coalescence, northwest European river discharge from the North European Plain, adopted a more southern route and flowed towards the Dover Strait (Bridgland and D’Olier, 1995; García-Moreno, 2017; Toucanne et al., 2009a, 2009b). Current hypotheses state that between this land bridge and the merged ice-sheets a proglacial lake developed (Collier et al., 2015; Gibbard, 1988, 1995, Gupta et al., 2007, 2017). It is proposed that the presence of this lake was maintained by discharge from the North European Plain drainage systems along with meltwater from the merged ice-sheet. Lake outflow to the south was restricted by the Weald-Artois anticline at the Dover Strait, until a threshold was reached and the land bridge was breached (García-Moreno, 2017; Gibbard, 1988, 1995; Gupta et al., 2017). The traces of this breach are still present today in the form of deeply scoured palaeobasins of the Fosses Dangaerd at the Dover Strait (García-Moreno, 2017; Gupta et al., 2017) and of layers of terrigenous sediment that accumulated at the continental margin of the Gulf of Biscay, during the Elsterian, from capture of the North European Plain and southern North Sea Basin fluvial systems (Toucanne et al., 2009a; Figure 2.1 for location and Figure 2.7 for the sedimentary signal).

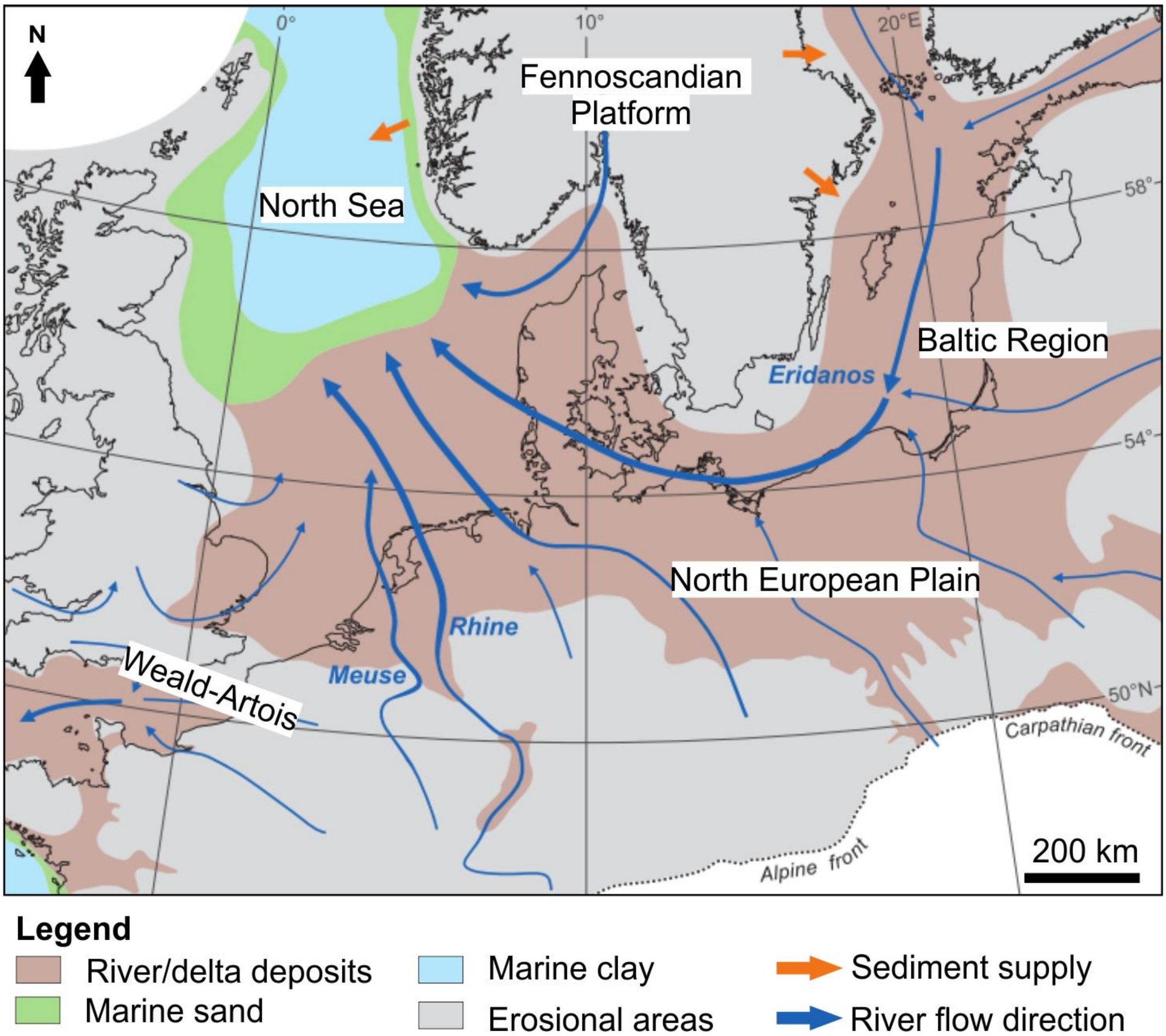


Figure 2.4 Palaeogeographic situation during the Early Pleistocene (modified from Gibbard and Lewin, 2016).

These terrigenous sediment accumulation rates at the European continental margin support an Elsterian age for initial breach. Further, the structural degradation of a land bridge at the Dover Strait is corroborated using shell assemblages from the southern North Sea Basin that imply a marine connection with the English Channel, during high-stand, at some point between the Elsterian and Saalian glaciations, i.e. the Holsteinian interglacial (Meijer and Cleveringa, 2009).

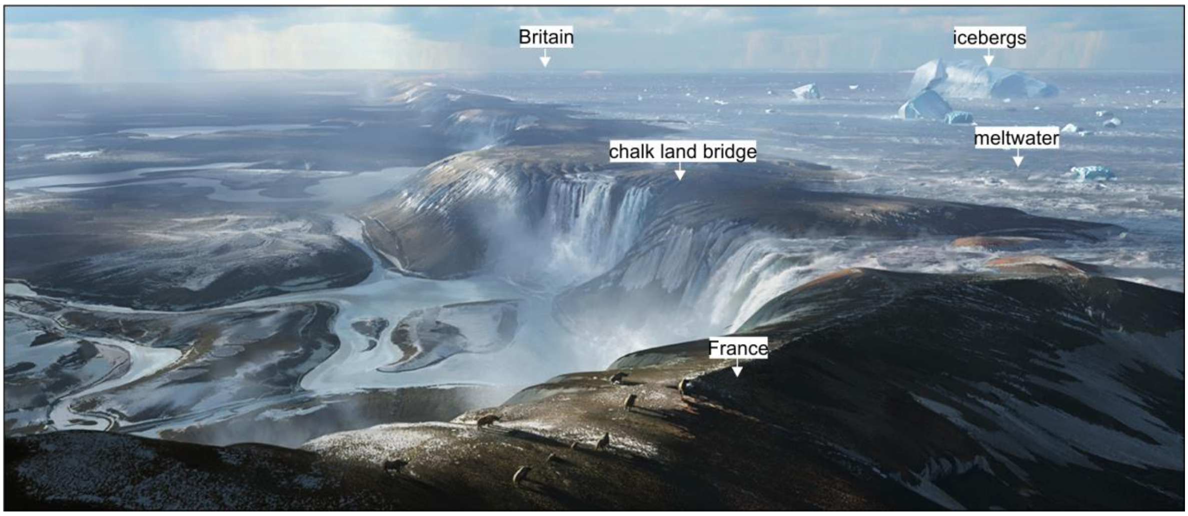


Figure 2.5 Breaching of the Weald-Artois land bridge that connects continental Europe and Britain during the Elsterian glaciation ca. 450 ka (cf. Gupta et al., 2017).

The presence of a proglacial lake in the southern North Sea Basin is often considered a precursor to breaching at the Dover Strait as these lakes provide large quantities of water, and have the ability to release them relatively instantaneously (i.e. GLOFs), therefore enabling erosion of the chalk bedrock land bridge (Gibbard, 1988, 1995), although gradual retrogressive erosion cannot be excluded as an initiation mechanism prior to breaching. Reconstructions of lake extent in the southern North Sea Basin during the Elsterian imply the lake covered an area of ca. 140.000 km^2 (Murton and Murton, 2012), however recent geophysical modelling results may suggest that the lake was smaller and had different dimensions than presumed (Moreau et al., 2015). These reconstructions are based on the distribution of glaciolacustrine deposits of Elsterian age in the southern North Sea Basin (Swarte Bank Formation: Gibbard, 2007; Long et al., 1988) and the northern Netherlands (Peelo Formation: Lee et al., 2012; van Heteren et al., 2016), and assume that the sediments were deposited contemporaneously across the basin. However, a lack of geochronological data makes it difficult to determine if a proglacial lake existed as a single body of water or if it was much more heterogeneously distributed as separate interconnected lakes. Further, it is still problematic to estimate the duration of lake existence without a chronology. Enigmatically high positions of fluvial deposits of Elsterian age in the Rhine-Meuse River catchment have been interpreted to reflect periods of high base level due to the presence of a proglacial lake in the southern North Sea Basin (Busschers et al., 2008).

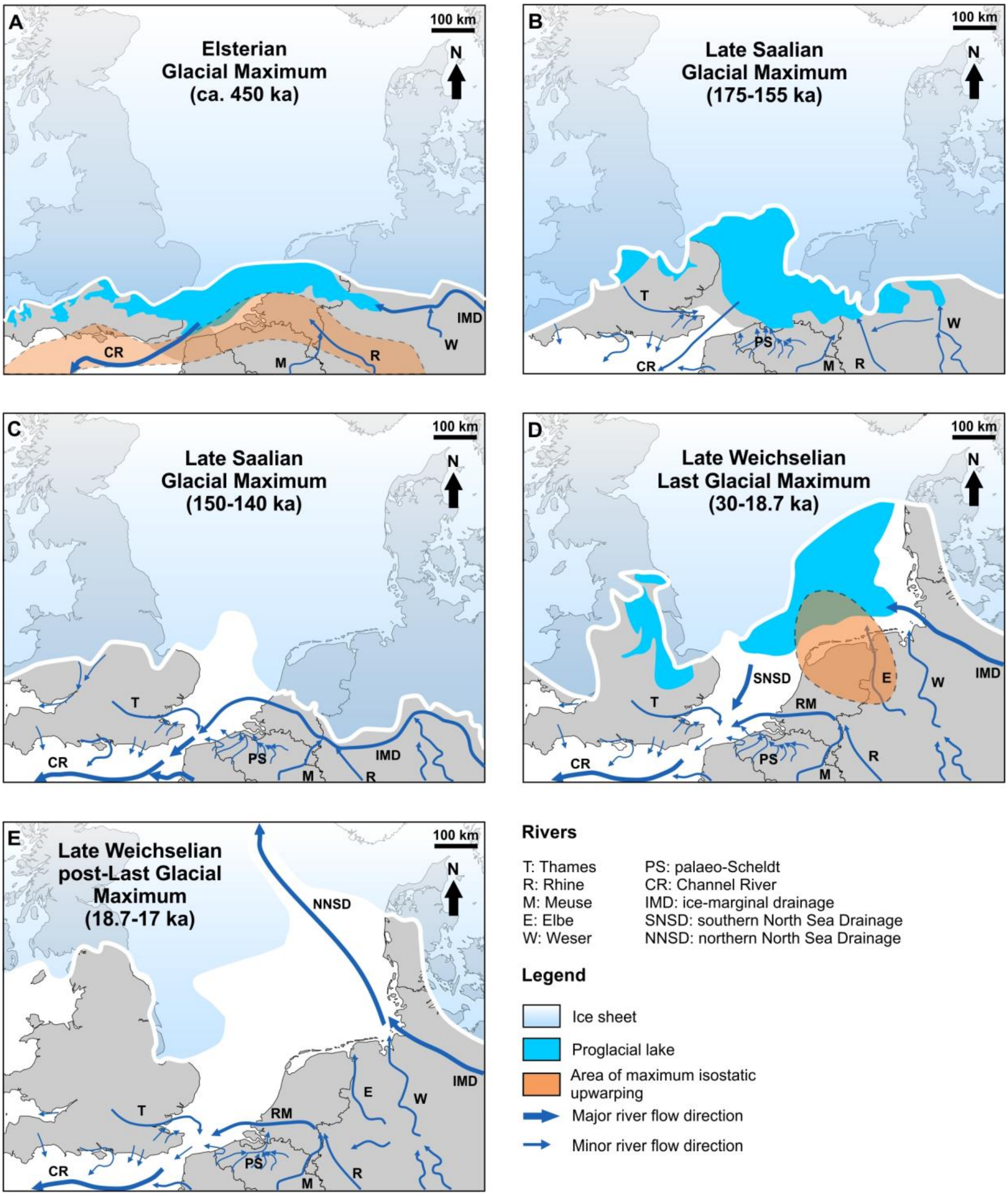


Figure 2.6 Palaeogeographic scenarios during glacial maxima of major Pleistocene glaciations. A) Elsterian Glacial Maximum; B) Saalian Glacial Maximum, proglacial lake phase; C) Saalian Glacial Maximum, post-proglacial lake phase; D) Late Weichselian Last Glacial Maximum; E) Late Weichselian post-Last Glacial Maximum. Maps modified from Busschers et al. (2007, 2008), Cohen et al. (2014), Ehlers et al. (2011), Gibbard (1988, 1995), Gibbard and Cohen (2015), Gibbard and Lewin (2016), Hijma et al. (2012), Moreau et al. (2015), Peeters et al. (2015, 2016) and Sejrup et al. (2016).

Further, a dramatic drop in base level in a subsequent proglacial lake basin during the Saalian, potentially driven by processes operating at a remnant land bridge of the Elsterian glaciation between East Anglia and northwest Belgium (Gibbard and Cohen, 2015), and draining of this proglacial lake, is recorded in the Rhine-Meuse delta (Busschers et al., 2008) (Figure 2.6 B-C). This breach is recorded as the accumulation of terrigenous sediments on the continental margin of the Bay of Biscay, during the Saalian, similar in amplitude to those of the Elsterian (Toucanne et al., 2009a; Figure 2.7), and imply the presence of a single interconnected proglacial lake. Furthermore these reconstructions are based on the distribution of glaciolacustrine deposits of Saalian age in the southern North Sea Basin (i.e. Cleaver Bank Formation: Cameron et al., 1986; Joon et al., 1990; Laban, 1995). These lines of evidence support the existence of a proglacial lake in the southern North Sea Basin during the Saalian. However, again it is difficult to reconstruct the lake's extent and chronology. The existence of a proglacial lake in the southern North Sea Basin during the Early Pleniglacial and Last Glacial Maximum is unknown. Coalescence of the FIS and BIIS during these periods (Carr et al., 2006; Clark et al., 2012; Ehlers et al., 2011; Graham et al., 2011; Patton et al., 2016; Sejrup et al., 2016) would have created the conditions necessary for proglacial lake development (Figure 2.6 D-E). However, no sedimentary evidence has been reported yet but it appears that geophysical and palaeogeographic modelling of glacial melting suggests the presence of a proglacial lake (e.g. García-Moreno, 2017; Patton et al., 2017).

Assuming the existence of a proglacial lake in the southern North Sea Basin is a key component driving incision at the Dover Strait as existing evidence would suggest an Elsterian age for breaching (García-Moreno, 2017; Gupta et al., 2017). However, it is unclear how much of the land bridge was eroded during this time and what effect the resulting topography had on the timing and duration of marine connections during subsequent high-stands. An increasing number of authors suggest the presence of a land bridge between East Anglia and northwest Belgium during the Saalian (Cohen et al., 2014; Gibbard and Cohen, 2015; Hijma et al., 2012). There is sufficient evidence to support erosion at the Dover Strait during this period as the result of a large-scale fluvio-hydraulic freshwater pulse as recorded in the Bay of Biscay (Figure 2.7). However, a more robust, highly resolved chronological framework of erosional events across the southern North Sea Basin is required to test competing hypotheses regarding the timing of breaching.

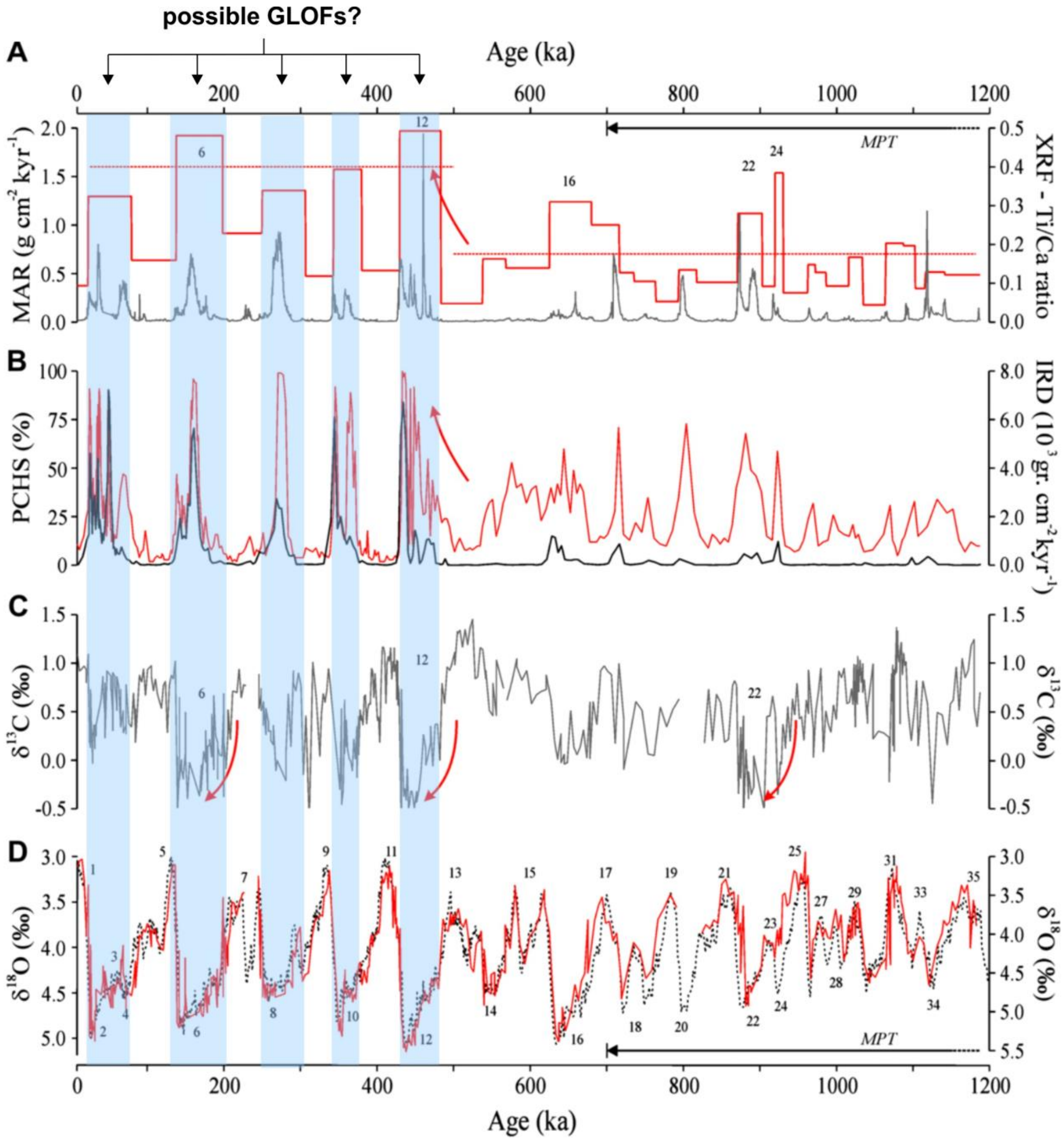


Figure 2.7 Summary of the data obtained from core MD01-2448 located in the Bay of Biscay (Toucanne et al., 2009a). (A). Terrigenous Mass Accumulation Rates (MAR, continuous red line; glacial average MAR for MIS 34-14 (0.73 $gcm^{-2}\ ka^{-1}$) and MIS 12-2 (1.62 $gcm^{-2}\ ka^{-1}$) intervals, dashed red line) and X-ray fluorescence (XRF) Ti/Ca ratio (continuous grey line); (B) Abundance of the polar planktonic foraminifera *Neogloboquadrina pachyderma senestral* (PCHS, red line) and ice-rafted detritus (IRD) flux (black line); (C) Benthic $\delta^{13}C$ and (D) $\delta^{18}O$ for core MD01-2448 (continuous red line) synchronized onto the LR04 benthic oxygen isotope stack (dashed grey line) (Lisiecki and Raymo, 2005). The arrows in the upper and lower part of the figure show the duration of the 'mid-Pleistocene transition' (MPT, from ca 1250 to 700 ka) according to the definition proposed by Clark et al. (2006). The excellent correlation of the benthic oxygen isotopic curve of core MD01-2448 with the LR04-stack strongly shows the absence of hiatuses in the studied sedimentary record. The blue blocks indicate time periods that may correlate to possible glacial lake outburst floods (GLOFs) that are clarified within this dissertation.

2.1.4 Note on Figure 2.7

Toucanne et al. (2009b) use a source-to-sink method to demonstrate a correlation between Pleistocene ice mass expansions into the Northern European Lowlands and their ability to modify the flow directions of Central European Rivers. This included massive southward discharges of fresh meltwater through the southern North Sea and the English Channel when the FIS and BIIS were confluent in the North Sea Basin. The complex network of palaeovalleys demonstrates that fluvial systems across the North Sea Basin merged in the southern North Sea Basin Axial Channel trunk valley and that downstream in the English Channel also the Somme, Seine, Solent and numerous minor French and British rivers during some Pleistocene eustatic lowstands entered this major river system only to form the Channel River. During these events terrigenous meltwater is transported to the Bay of Biscay where a continuous record of this Pleistocene climate variability (including ice-sheet oscillations) is recorded compared to the now discontinuous continental sequences across the North Sea Basin. With this observation, Toucanne et al. (2009a) acknowledge that sea-level fall and lowstand conditions favoured the seaward transfer of sediment and demonstrate that the evolution of the sediment source, superimposed on sea-level variations, acted as the main forcing of terrigenous input to the Bay of Biscay, the sink area. They also assume that the Channel River controlled the glacial terrigenous input variability and that the patterns visualised in Figure 2.7 reflect the glacial accumulations deposited at the Bay of Biscay and originate mainly from fluvial-derived sediments.

The greatest supplies of terrigenous sediment at the Bay of Biscay occurred during the Elsterian and Saalian Drenthe glaciations (MIS 12 and 6; Figure 2.7). Based on their data Toucanne et al. (2009b) demonstrate that MIS 12 corresponds both to the first time that ice-sheets were confluent, covering the North Sea Basin, and that the fluvial system was redirected southwards, thus greatly increasing terrigenous supplies to the sink area. They use this correlation to validate the hypothesis of a proglacial lake that catastrophically breached the Weald-Artois structural barrier (e.g. a GLOF). They make this correlation based on the high MAR and XRF correlation for MIS 12 (e.g. Figure 2.7). However, for MIS 6 they state that the discharge was catastrophic in character due to the low XRF measurements compared to the high MAR values. Several studies demonstrate that the late Saalian glaciation is correlated to a possible lake in the southern North Sea Basin followed by a breach event of this lake to the south in the direction of the Dover Strait (Busschers et al., 2008; García Moreno, 2017; Gibbard and Cohen, 2015; Gupta et al., 2007, 2017; Hijma et al., 2012; Murton and Murton, 2012). In the event of a breach during MIS 6 than the necessary correlation of a high XRF and MAR relationship for a GLOF may not be as valid as proposed by Toucanne et al. (2009b). By consequence the multiple glacial episodes since MIS 12 (MIS 10, 8, 6, 4 and 2) may have experienced GLOFs of uncertain magnitudes and that other processes, opposed to aa topographic barrier exert control on proglacial lake development in the North Sea Basin.

2.2 Belgian Continental Shelf

The BCS extends to a maximum of 84 km from the present-day Belgian coastline (67 km long) and covers an area of 3.454 km^2. Apart from major bedforms (sandbanks and dunes), the seafloor topography is gently dipping from 0 to 46 m below LAT in the more offshore parts of the shelf (Figure 2.8). In the near-coastal zone, depths range between 0 and 15 m; in the middle part of the shelf depths range between 15 and 35 m below LAT; and in the north, the seabed is relatively flat, not covered with large bedforms, deepening from 35 to 46 m below LAT. The sandbanks on the BCS can be tens of kilometres long, one to several kilometres wide, and up to 25 m high. They are mostly asymmetric in cross-section and their plan view commonly shows kinks. Based on their position and orientation they are grouped in the Coastal Banks, the Flemish Banks, the Zeeland Ridges and the Hinder Banks (Figure 2.8). The Coastal Banks and the Zeeland Ridges are located quasi parallel to the coastline, whereas the Flemish and Hinder Banks are clearly oblique with respect to the coastline. The sandbanks display angles of 0–20° relative to the main axis of the tidal ellipse (Mathys, 2009). In the offshore area, large dunes (2–8 m high) are also present, mostly superimposed on the sandbanks. Closer to the coast, their occurrence is more restricted and the sandbanks are generally devoid of bedforms.

The BCS basement is composed of solid layers of various ages. The Palaeozoic basement is found at a depth of ca. 250 m near the French border to 450 m near the Dutch border. It is a relatively stable continental block called the London-Brabant Massif which was only flooded since Late Cretaceous times. During this period chalk was deposited with a thickness of 50 m between Nieuwpoort and Oostende, and rapidly increasing offshore to a maximum of 220 m in the area of the Hinder Banks. The depth of the chalk top increases in a northeast direction to depths of 150 to 350 m (De Batist, 1989).

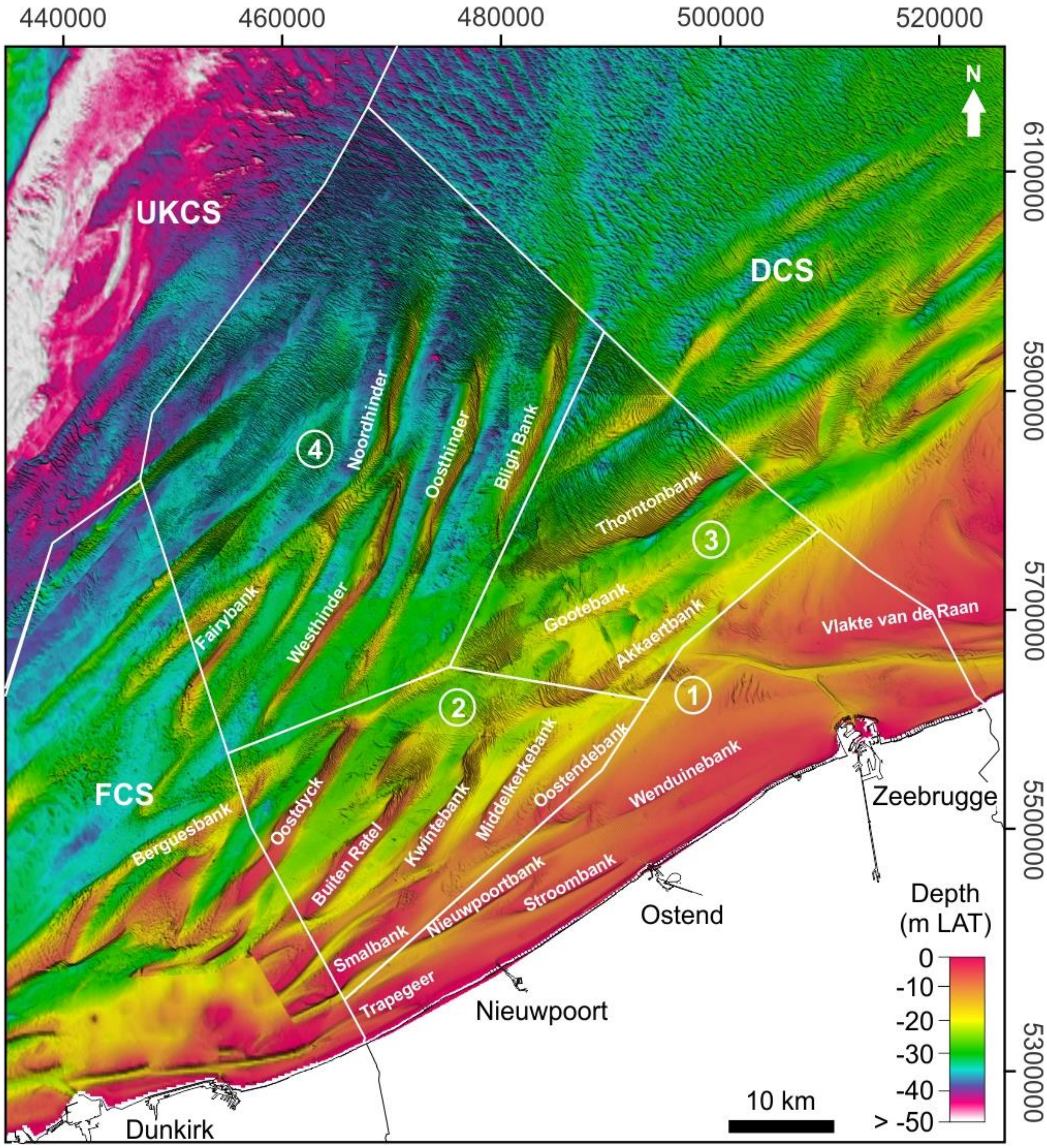


Figure 2.8 Seabed morphology of the BCS and the neighbouring shelves (compilation: Emodnet, Deltares and Flemish Hydrography and FPS Economy - Continental Shelf Service; see Chapter 3). 1: Coastal Banks; 2: Flemish Banks; 3: Zeeland Ridges; 4: Hinder Banks.

2.2.1 Paleogene substratum and its geomorphology

The Paleogene substratum of the BCS consists of Eocene deposits, mostly of marine origin, slightly tilted (0.5–1°) towards the north-northeast. In the west, the BCS is underlain by homogeneous heavily compacted clays of Ypresian age, while in the east an alternation of sand-silt-clay is present, up to the Belgian-Dutch border (Figure 2.9; see Le Bot et al., 2003 for a synthesis). From the late Neogene onwards, this substratum was exposed to marine, fluvial, estuarine, and periglacial environmental conditions (Mathys, 2009), resulting in a polygenetic, mostly erosional geomorphological surface, comprising escarpments (e.g. the Middle and Offshore Scarps; Liu et al., 1992), platforms (e.g. the Marginal, Middle and Offshore Platforms; Liu et al., 1992) and numerous palaeovalleys (e.g. the Ostend Valley, Thornton Bank Channel, Northern Valley, Coastal Valley: Liu et al., 1992).

The top of the Paleogene substrate is composed of various consolidated layers that were deposited in a marine to near-coastal depositional environment (Table 2.1) and as a result the units have strong differing geotechnical properties (Le Bot et al., 2003, 2005).

Table 2.1 Depositional environments of the Paleogene units of the BCS (modified from Le Bot et al., 2003).

Seismic unit	Depositional environment
P1	Shallow marine and barrier-protected lagoonal, wash-over and tidal flat environments
B1	Central to outer shelf environment (20–50 m water depth)
L2	Transgressive lag deposit overlying a ravinement surface
L1	Very shallow marine environment
Y5	Lagoon towards mouthbar or crevasse splay in a deltaic to intertidal environment
Y4	Eroded and then infilled depression (channel or basin?)
Y3	Nearshore mudshelf (15–30 m) with delta and tide influence
Y2	Deltaic origin deposits with a southern sediment supply
Y1	Mud-shelf environment below storm-wave base

Unit Y1 crops out in the western and central part of the BCS, while units Y2, Y3, L1, B1 and P1 crop out in the east of the BCS. Unit Y1 displays the largest thickness (150 to 180 m) and units Y2, Y3 and L1 display a maximum thickness of 30 m whereas units B1 and P1 have large respective thicknesses of 45–60 m and 40–90 m. Units Y4 and Y5 crop out in a restricted part of the shelf. Unit Y4 is located underneath the Akkaertbank and the Gootebank and corresponds to an erosion/infilling stage, confined into a channel or erosional depression, trenched in units Y1 (uppermost layers), Y2 and Y3. Unit Y5 extends from the coast to the Thorntonbank with a maximum thickness of 17 m. For a more detailed summary see De Batist (1989) and Le Bot et al. (2003).

Within unit Y1 a wide range of intraformational ‘sediment tectonic’ deformations are observed and were formed due to the relaxation of temporary states of density inversion linked to undercompaction in the early burial history of the clayey-silty sediment (for an overview see De Batist, 1989 and Le Bot et al., 2003).

The geomorphology of the top of the Paleogene substratum was shaped by multiple erosional episodes since the Late Saalian (Hijma et al., 2012; Mathys, 2009), when the southern North

Sea Basin experienced strong geomorphological alterations imposed by the merging FIS and BIIS (Cohen et al., 2014; Ehlers et al., 2011; Gibbard and Lewin, 2016).

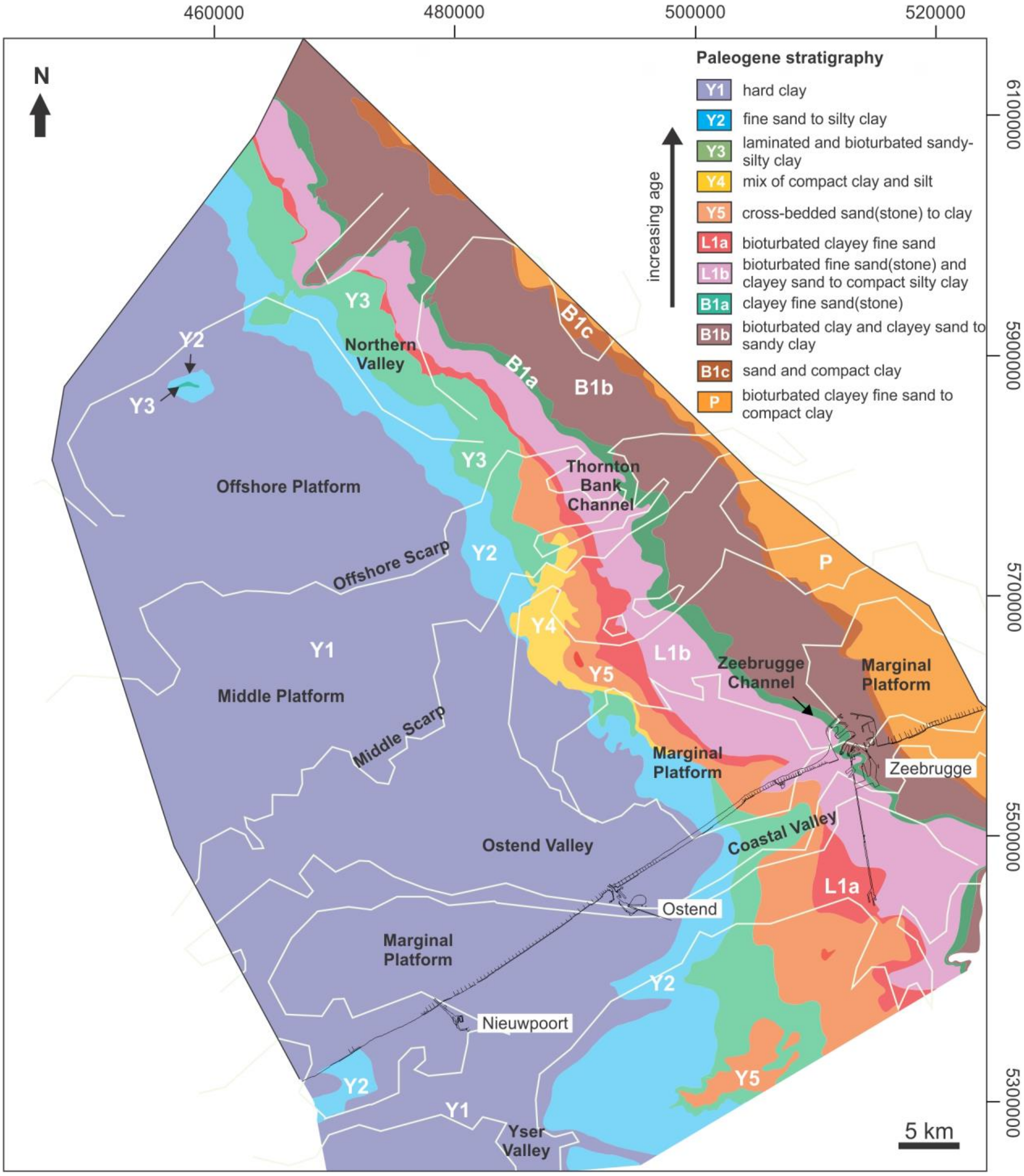


Figure 2.9 Base of the Quaternary unconsolidated deposits, alternatively the top of the Paleogene substratum. Paleogene consolidated strata modified from Le Bot et al. (2003), Paleogene morphology modified from Mathys (2009).

2.2.2 Quaternary deposits and evolution

The present Quaternary stratigraphic model of the BCS is based on the internal structure of the Middelkerke Bank after Berné et al. (1994), Trentesaux (1993) and Trentesaux et al. (1993). These authors interpreted that most of the Pleistocene deposits on the BCS had been completely reworked during the Holocene transgression and had become incorporated in the Holocene deposits. With this reasoning, it was proposed that the lower units of the Middelkerke sandbank, which make up the base of the Ostend Valley fill (i.e. the present offshore section of the palaeo-Scheldt Valley), are of Holocene age. However, this hypothesis was based on studies that focussed on the DCS (Jelgersma et al., 1979) and the eastern Belgian coastal plain. The authors that studied the Belgian coastal plain (De Breuck et al., 1969; Depret, 1981; Devos, 1984; Heyse, 1979; De Moor and Heyse, 1974; Paepe and Baeteman, 1979) even stated that Eemian marine deposits are dominantly present in the eastern Belgian coastal plain.

Trentesaux et al. (1999) on the other hand dated the Ostend Valley fill to the Middle Weichselian and the overlying tidal-flat deposits to the Holocene. They base their assumption on ^{14}C ages from juvenile marine shells by Trentesaux et al (1993), and state that these ages were rather controversial and gave ages above the ^{14}C limit at the time (ages greater than 40 ka). Stolk (1996) earlier noted that the presence of non-reworked marine shells of Weichselian age in this area is rather remarkable, because at that time sea level was always more than 35 m below present, whereas the dated samples were collected at depths of less than 27 m below present mean sea level.

The development towards a Quaternary stratigraphic model of the BCS evolved from the stratigraphic structure of the centrally located Middelkerke Bank, as this sandbank is the best studied and best ground-truthed (Trentesaux et al., 1999 and references therein). Regardless of the remaining uncertainties about their age (see above), four main depositional phases following the last phase of Ostend Valley incision (Eemian or older: Liu et al., 1992, 1993) were distinguished in the Middelkerke Bank, and were used as a model for the infilling of the entire Ostend Valley: i.e. 1) valley infilling; 2) deposition of sub-tidal or tidal-flat deposits; 3) the construction of initial ridges or coastal banks; and 4) the development of tidal sandbanks. Some of these phases have been recognised in neighbouring sandbanks, but have never been correlated or (relatively) dated to create a common stratigraphy or a genetic model for the Quaternary geological evolution of the BCS until 2009 by Mathys. Mathys (2009) distinguished seven seismic units, separated by erosional unconformities within the Quaternary deposits on the BCS (Table 2.2). The relative age of the units range from the Eemian to the Holocene, however, no Weichselian sediments have been recognised. The geological model of Mathys (2009) specifically focusses on the middle part of the BCS as this was the area with the most seismic reflection data and the presence of several deep continuous cores.

Table 2.2 Depositional environments of the Quaternary seismic units of the BCS (modified from Mathys 2009).

Seismic unit	Depositional environment
U1	Lateral-accretion within a migrating channel in a tide-dominated estuary
U2	Middle estuary
U3	Upper estuarine infill
U4	Back-barrier environment
U5	Storm-generated ridges, with a possible tidal influence
U6	Reworked tidal flat deposits within a sublittoral environment below wave action
U7	Shallow marine deposits (including tidal sandbanks)

2.3 The Flemish Valley

In general terms, the Flemish Valley is the most inland geomorphological section of a larger incised-valley system that was repeatedly scoured during sea level glacio-eustatic lowstand periods to drain Flanders and its emerged continental shelf (the BCS) towards the southern North Sea and its Dover Strait outlet (Mathys 2009; Figure 2.10 A). The Flemish Valley *sensu stricto* includes the buried valleys of the middle Scheldt and Lys to the west, and an eastern branch encompassing the Scheldt, Dender, Zenne, Dijle, Demer and Nete Valleys. All tributaries join near Ghent to form the downstream part of the Flemish Valley, opening widely towards the north-northwest. Based on seismic reflection data, Mathys (2009) shows that, in the present southern North Sea Basin, the system is divided into the west-northwest-flowing Dutch Channels (a precursor of the Lys and Waardamme Valleys) and the main west-southwest-trending Coastal-Ostend Valley (Liu, 1990; Liu et al., 1992; Mostaert et al. 1989) flowing northwest into the major Axial Channel of the southern North Sea.

The youngest comprehensive synthesis concerning the fluvial terrace sequences that are preserved above the present floodplains (Saalian and older) of the Flemish Valley dates back to Tavernier and De Moor (1974). Similarly, a qualitative geomorphological analysis of the valley system dates back to De Moor and Heyse (1978). Within the Scheldt-Lys valley system the pre-Saalian terraces are not terraces *sensu stricto* but rather correspond to alluvial gravel deposits scattered thriftily on planation surfaces on top of the Flemish Hills (Tavernier and De Moor, 1974; Figure 2.10 B). A major problem with these higher located alluvial gravel deposits is their high degree of chemical weathering, which make it rather difficult to assign their true depositional nature (e.g. marine, coastal, fluvial, etc.) and to provide a true age estimate (e.g. pre-Pliocene age).

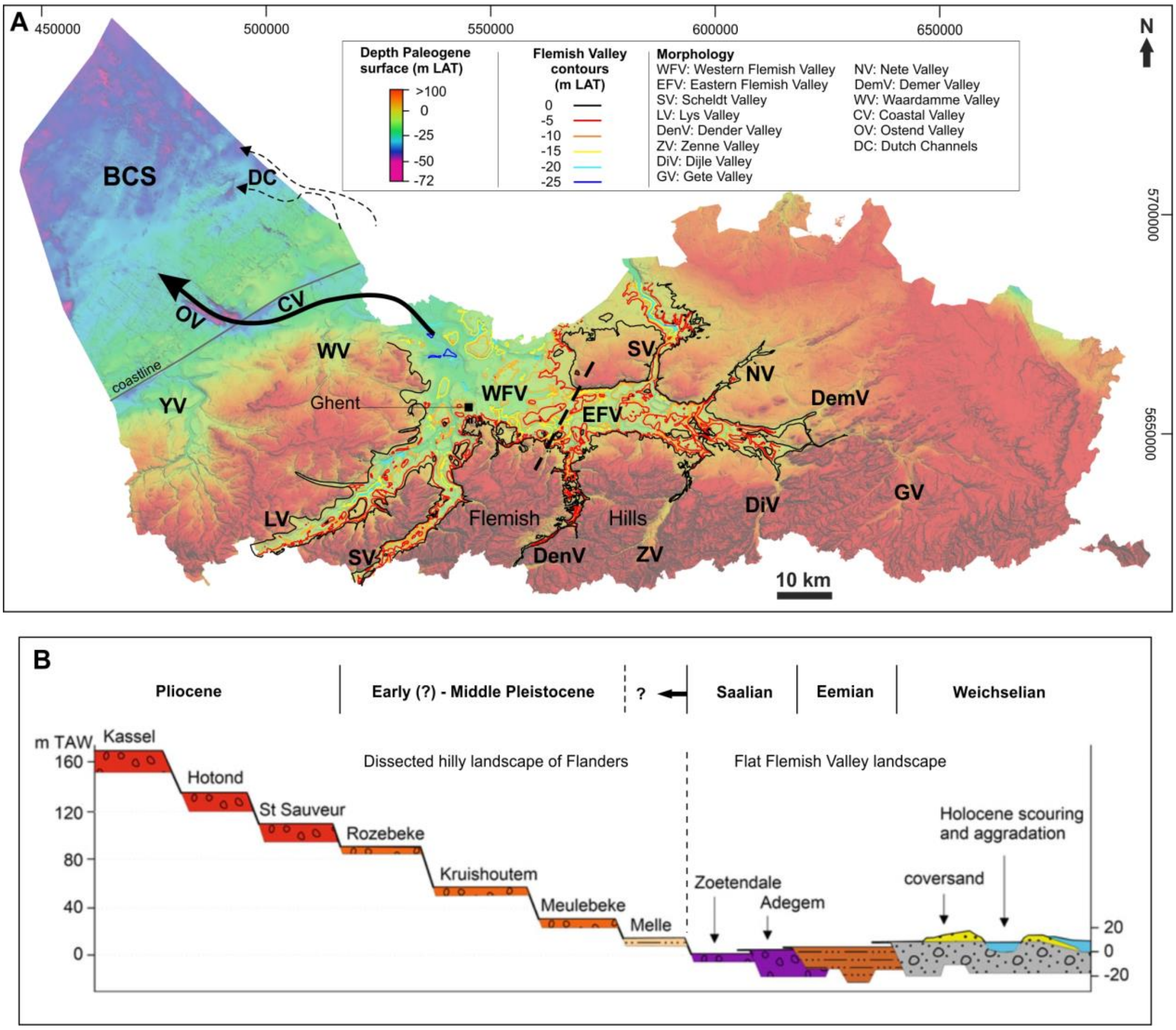


Figure 2.10 A) The pre-Quaternary surface with indication of the Flemish Valley contour lines below present-day levels. The black arrow indicates the drainage pathway of the Flemish Valley in the direction of the southern North Sea and the Dover Strait before it was diverted towards its present-day course (Mathys, 2009). B) A synthetic cross-section of the terrace sequences within the middle Lys-Scheldt Valley area (from Heyse and Demoulin, 2018).

The gravels positioned at 20–100 m above present-day floodplains can't be assigned to a specific river valley (Heyse and Demoulin, 2018). Only local remnants of the lower Meulebeke Terrace clearly pertain to the Lys Valley. Beyond indications of progressively deteriorating climate conditions and tentative correlations with glacial periods (De Moor 1963; Tavernier and De Moor 1974), no reliable age data is so far available to establish any chronology of the Pliocene to Middle Pleistocene incisional history of these 'terraces' preserved within the palaeo-Flemish Valley (Heyse and Demoulin, 2018). Finally, the stratigraphic position of the lowest palaeo-Flemish Valley terrace, the Melle Terrace, is known only very locally, southeast of Ghent. According to De Groote (1977), the presence of *Azolla filiculoides* in peat layers is indicative to the Holstein interglacial (MIS 8). This view was later contradicted by Vanhoorne (1992) who placed it in the Middle Saalian Wacken interstadial (MIS 7, according to the latest stratigraphy of Stephan, 2014), leaving some uncertainty about the timing of the present Flemish Valley development marked by cut-and-fill processes (Heyse and Demoulin, 2018).

2.3.1 Plio-Pleistocene evolution *(based on Heyse and Demoulin, 2018)*

While the Late Saalian and post-Saalian basic characteristics and nature of the processes of the Scheldt valley system have been identified, its formation history beyond the Holsteinian-Early Saalian (MIS 8–7) of the last stage preceding the creation of the Flemish valley (Melle Terrace; De Groote, 1977; Vanhoorne, 1992) remains speculative. In general terms, the history of fluvial incision and terrace formation are the result of the long-term northeast tilt towards the North Sea Tectonic Basin (Figure 1.1). Before the Late Saalian (MIS 6) reorganisation, up to the Late Miocene (11.6–5.33 Ma) a consequent northeast-oriented river system is observed that is now buried beneath the Diestian Formation in northern Belgium (Vandenberghe et al., 2014). The present north-northeast direction of Belgian River system is likely the result of modifications and alterations of an ancient drainage system by stream captures when the main regional tilt component turned to north-northwest in parallel with the more and more eastward location of maximum uplift in the Ardennes during the Plio-Pleistocene (5.33 Ma–11.7 ka; Houthuys, 2014). In this context, of a long period of slow valley downcutting since the Oligocene (33.9–23.03 Ma), it is until today hardly possible to assign ages to the successive terrace sequences of the palaeo-Scheldt Valley. While it is relatively certain that the lower terraces result from the combined effect of climate oscillations and accelerated uplift since the Middle Pleistocene (465–126 ka; Heyse and Demoulin, 2018), the upper terraces are believed to be the result of fluvial response to a slow Neogene (23.03–2.58 Ma) tilt toward the North Sea Tectonic Basin (Heyse and Demoulin, 2018). However, unlike in the terrace staircase of the Ardennian valleys, there is here no clear hiatus in the vertical spacing of the successive terraces within the palaeo-Scheldt Valley that would point to the remote effect of the Middle Pleistocene uplift pulse in the Ardennes (Heyse and Demoulin, 2018). Based on these observations, Heyse and Demoulin (2018) consider the palaeo-Scheldt River response to be related to regional-scale tectonic signals and hypothesise that, if applicable in the palaeo-Scheldt Valley, this pulse, which occurred prior to the formation of the Melle Terrace, might have been responsible for the ca. 30 m incision leading from the Kruishoutem to the Meulebeke Terraces (Figure 2.10 B). The

Meulebeke Terrace is indeed the first terrace for which sedimentological remnants are preserved within the valley system rather than on top of its interfluves, suggesting the Middle Pleistocene uplift/incision event fixed the position of valleys that were previously laterally mobile (Heyse and Demoulin, 2018). The equally large incision above the Kruishoutem Terrace might then be understood as a response to the Weald–Artois uplift that closed the Dover Strait around 1.4 Ma (Van Vliet-Lanoë et al. 1998; Figure 2.4). By contrast, and although age issues are still debated, the more recent history of the Flemish Valley is better documented (Heyse and Demoulin, 2018). After slowdown of downcutting in all tributaries of the Flemish Valley for the Elsterian-Saalian period (e.g. Hijma et al., 2012, De Moor and Heyse, 1978; Tavernier and De Moor, 1974; Vandenberghe and de Smedt, 1979), a phase of generalised renewed incision commenced during the Saalian and culminated with the Early Eemian transgression in the Ostend Valley and its continuation toward the inland Flemish Valley (e.g. Mathys, 2009). However, recent findings tend to confine this multi-stage incision to the Saalian (Bogemans et al., 2016). After this phase, a cut-and-fill evolution was initiated, which was no more interrupted by significant further deepening of the Flemish Valley. While the post-Elsterian waning of river incision, giving progressively way to Flemish Valley cut-and-fill points to a phase of tectonic quiescence in the area, a specific cause has yet to be found for the Saalian incision (Heyse and Demoulin, 2018). In the absence of tectonic evidence, it was very soon related to the opening of the Dover Strait (Paepe and Vanhoorne, 1976; Vandenberghe and de Smedt, 1979). This picture was later confirmed and refined by Gupta et al. (2007, 2017) and showed the latter underwent two *megaflood* events. These events most likely occurred during the Elsterian (MIS 12) and the last glacial phase of the Saalian (MIS 6; e.g. Gibbard, 1995; Toucanne et al., 2009a; Cohen et al., 2014; Figure 2.6 A to C), when most northwest and central European rivers, including the palaeo-Scheldt River, drained into a large proglacial lake in the southern North Sea, dammed by the merged FIS and BIIS in the north and the uplifted Weald–Artois axis in the southwest, that acted as a land bridge between Continental Europe and England. Spilling of the lake waters over the lowest point of the Weald–Artois axis rapidly evolved in high fluvio-hydraulic flow and caused *catastrophic* breaching of the Dover Strait topographic barrier. As the base level of the resulting Channel River was located ca. 120 m below the present sea level, downstream at the Bay of Biscay, and the Channel floor lay significantly lower than the top of the topographic barrier, each breaching event involved a wave of regressive erosion that propagated far into the Flemish Valley system. Strangely enough, the Elsterian event did not leave any identified traces, perhaps because it interfered with the ongoing tectonically driven incision of the valley (Heyse and Demoulin, 2018). The Saalian event on the other hand, induced a 15–20 m deepening of the Channel River (Gupta et al., 2007), and is held most responsible for the inland deep incision that created the Flemish Valley (Beerten, 2010; Beerten et al., 2013). Once the present form of the Flemish Valley was established, the incision and aggradation stages basically reflected the combined controls of glacio-eustatic variations of the Flemish Valley base level, alternations of transgressive-regressive regimes, and climate-driven variations in hillslope sediment delivery (Heyse and Demoulin, 2018). Remarkably, the oldest sediments preserved in the lower Western Flemish Valley are Saalian in age, while incision and early deposition in the upper Eastern Flemish Valley occurred slightly later, during the Eemian (Bogemans, 1993). Heyse and Demoulin (2018) postulate that the precise timing of the Saalian incision of the Flemish Valley at first glance appears unclear, not only because the ages of the Zoetendale and Adegem Terraces, as proposed by Tavernier and De Moor (1974), are inferred rather than positively determined but also because the Saalian chronology is still debated, notably with respect to the inclusion of MIS 9 and 10 in the Saalian complex (e.g. Busschers et al., 2008; Stephan, 2014). However Busschers et al. (2008) demonstrate, based on the OSL-chronology of Rhine-Meuse River sediments in the Netherlands, that the Drente maximum ice advance of the FIS and BIIS over the North Sea Basin occurred in MIS 6. Therefore Heyse and Demoulin (2018) deduce that, even if Beets et al. (2005) discuss the possibility of an MIS 8 ice advance in the North Sea Basin, the *catastrophic* breaching of the Dover Strait topographic barrier and the subsequent Flemish Valley incision most probably occurred during MIS 6. Moreover, if the Melle Terrace was indeed formed during the Wacken interstadial (i.e. MIS 7, see Stephan, 2014), rather than depending on glacio-eustatic oscillations within MIS 6, the two-stage (Zoetendale/Adegem) incision of the Flemish Valley and certainly also the stepped flanks of the Ostend Valley (Liu, 1990; Liu et al., 1992; Mathys, 2009) probably resulted from a similar multi-stage deepening of the breach across the Dover Strait topographic barrier by repeated lake overspill and emptying, which led to temporary interruption(s) and partial removal of the sediments before the fluvio-periglacial filling of the valley resumed (Heyse and Demoulin, 2018). Then, at the Saalian-Eemian transition, when sea levels were still low, the strongly decreased supply of hillslope sediment to the rivers under warming conditions allowed the drainage system to re-incise in its Saalian filling (Heyse and Demoulin, 2018). Prolonged by the action of migrating tidal channels during the following transgression (Mathys, 2009), this removed the largest part of the filling and locally further carved the Paleogene substratum down to -25 m, and even a -28 m maximum in the downstream located Ostend Valley (Figure 2.10). With the Eemian transgression, a succession of fluvial, estuarine and shallow marine environments migrated landward, progressively invading and infilling the Flemish Valley with sediments. Except south of Ghent or in more protected areas of the northern part of the river system (e.g. inherited palaeotopography), the initial fluvial sediments were largely removed from the Flemish Valley by the succeeding high-energy tidal currents in the propagating estuary, which left only a basal layer of residual gravel below 5–20 m of tidal channel and tidal flat sediments (Heyse, 1979). During the Eemian high-stand, sea level was a few metres above the current one; the marine influence was felt as far as 40 km inland, with Eemian marine deposits preserved up to +2,5 m TAW in the Flemish Valley (De Moor and Heyse, 1974). During the Early Weichselian sea-level fall (ca. -40 m and ca. -50 m during MIS 5d and 5b, respectively; Waelbroeck et al., 2002) entailed renewed erosion in the Flemish Valley and partial removal of the Eemian infill. Based on pollen data, Verbruggen et al. (1991) demonstrate that the climate was cool and very wet in central and northwest Belgium during the Early Weichselian (ca. 115–70 ka), which, combined with the base level fall, allowed rivers to incise deeply, locally down to -17 m (Tavernier and De Moor, 1974), in the still unfrozen fill. A similar incision is concurrently observed downstream in the re-emerging Ostend Valley (Mathys, 2009), whereas the upstream part of the Eastern Flemish Valley shows only hints of limited erosion by meandering channels (Bogemans, 1993). The deteriorating climatic conditions of the Early Pleniglacial (74–59 ka: Huijzer and Vandenberghe, 1998), colder than in the Early Weichselian but still wet according to Verbruggen et al. (1991), caused a lot of hillslope material to be delivered by periglacial slope processes to the downstream reaches of the valleys, where braided rivers were most often incapable to cope with such a load. This led first to the accumulation of alluvial fans in the Flemish Valley, going rapidly laterally and vertically into cryoturbated sandy channel fills during the Middle Pleniglacial (59–27 ka: Huijzer and Vandenberghe, 1998), which in turn were overlain by overbank deposits testifying to quiet floodplain environments that still largely determine the modern flat topography of the Flemish Valley. At the turn of the Late Pleniglacial (27–13 ka: Huijzer and Vandenberghe, 1998), while most of the up to 20 m-thick Weichselian fill of the Flemish Valley was accumulated, the regional climate not only was very cold but became also extremely dry, leading to the onset of a predominant aeolian activity that would give the final touch to the Flemish Valley landscape during the Late Pleniglacial and the Late Glacial (Heyse and De Moor, 1979).

2.4 Archaeology

Palaeolithic archaeology is a discipline that is for the largest part biased towards accidental discoveries; with little theory predicting where new finds are to be expected based on the current distribution of traces of hominin presence (Finlayson, 2004). Recent discoveries along the East Anglian North Sea coast (United Kingdom) have once more raised questions concerning the strength of former theories on the earliest occupation of northwest Europe, such as the inferred 'habitat tracking' of the earliest occupants of the mid- and northern latitudes (Roebroeks, 2005). Flint artefacts from Pakefield and Happisburgh demonstrate that hominins were already present in northwest Europe before 500 ka (Ashton and Lewis, 2012; Parfitt et al., 2005, 2010; Preece and Parfitt, 2012). Recently, this was corroborated by the discovery of Early Pleistocene (0.78–1 Ma) human (*Homo antecessor*) footprints preserved in estuarine clays at Happisburgh (Ashton et al., 2014).

The Palaeolithic occupation of northern Europe can be seen as an intricate balance between climate change and the progressive technological development of humans able to cope with gradual deteriorating climate conditions. The archaeological record in northwest Europe, starting at over 800 ka, suggests repeated colonisations occurred starting from the ameliorating glacial-interglacial transitions and thereon and retreats or local extinctions when climate deteriorated again (Parfitt et al., 2005, 2010; Stringer, 2006; White and Schreve, 2000). The technological development to deal with long, cold winters, such as clothing, shelter, fire and more effective hunting, from perhaps ca. 500 or 400 ka eventually led to the ability to survive in harsher climates, such as in northwest Europe, which was prone to climate change (Ashton and Lewis, 2012; Hosfield, 2016). By the Last Glaciation humans developed in such a way they were able to tolerate glacial conditions other than during the coldest periods of the Early Pleniglacial (MIS 4) or the LGM (MIS 2; Boismier et al., 2012; Roebroeks et al., 2011).

As a peninsula or island of northwest Europe, Britain is and has always been a *dead end* for northern human migration. Due to the catastrophic permanent opening of the Strait of Dover at the end of the Elsterian (MIS 12), access to peninsular Britain became restricted especially during interglacial high sea levels (Gibbard, 1995; Gupta et al., 2007; Smith, 1985; Toucanne et al., 2009a; White and Schreve, 2000). A basic understanding of the dramatic regional palaeogeographic evolution of the southern North Sea Basin (see 2.1.3 Pleistocene geological history) is therefore critical to understand regional human migration patterns and why artefacts or occupation sites are encountered across various depositional environments. Recent palaeogeographic considerations by Hijma et al. (2012) suggest that during the Holsteinian (MIS 11) and possibly up to the Wacken interstadial (MIS 7: Stephan, 2014) there may have been a semi-permanent land bridge in the southern North Sea Basin. This land bridge acted as a watershed between the Rhine-Meuse and East Anglian Rivers flowing to the north and the

Thames and the Scheldt Rivers flowing to the south. It is suggested that this land bridge corridor was permanently removed during the Late Saalian (MIS 6), after which time the Rhine and Meuse Rivers flowed south to join the Thames, Scheldt and Channel Rivers (Hijma et al., 2012; Toucanne et al., 2009a).

A land bridge corridor, as suggested by Hijma et al. (2012), between Continental Europe and Britain, may have been an important landform element for human migration and occupation. Recently, a comprehensive analysis of the complete dataset of the British Lower and Early Middle Palaeolithic artefacts and findspots was used to test these migratory incursions into Britain (Ashton et al., 2018). The study shows that already by the end of MIS 8 or early MIS 7 early Neanderthals colonised southeast England from northwest Europe via a narrow landbridge located between Continental Europe and Britain.

This suggests that in Flanders archaeological evidence of humans should be present. In fact, hundreds of archaeological sites and artefacts have been found across a wide variety of depositional environments (https://onderzoeksbalans.onroerenderfgoed.be) dating from the Late Middle Palaeolithic (Eemian: MIS 5e) onwards, with the oldest evidence dating back to ca. 300 ka upstream the Meuse River at the Belgian-Dutch border. Unfortunately, a comprehensive study of all sites and artefacts, like Ashton et al. (2018) in Britain, is still lacking in Flanders.

Contrary to evidence from land hardly any evidence of human presence has been found in the southern North Sea Basin and its deposits. The finds that have been encountered though always received much media attention and revealed much about human migration across northwest Europe. One of the most intriguing examples was a Neanderthall skull fragment dredged up in 2005 from an abandoned Middle Pleniglacial (MIS 3) Rhine-Meuse river terrace not far outside the BCS (Hublin et al., 2009). This find testifies to the fact that human artefacts and bone material are preserved within the diverse suite of drowned landscapes now buried beneath the North Sea seafloor.

3 GEOPHYSICAL DATA AND DATA INTEGRATION

3.1 Introduction

The present study is based on the analysis and interpretation of marine geophysical data. These consist of bathymetric data and two-dimensional single-channel (and to a smaller extent also multi-channel) seismic-reflection profiles. The geophysical dataset gathered for this study comprises the seismic reflection database compiled by Mathys (2009) and new data acquired by the author of this dissertation and colleagues during geophysical surveys during the period 2013–2017. In addition, seismic reflection datasets were obtained from various commercial-industrial organisations (NemoLink, Fugro) that helped expand the complete seismic reflection database.

The interpretation of the geophysical data was carried out by combining the seismic interpretation software IHS Kingdom Suite, with Geographic Information System (GIS and QGIS) and Global Mapper. Seismic-stratigraphy, velocity models and geological mapping were constructed by combining the seismic reflection and bathymetric data with a series of archived sediment core descriptions gathered from the onshore (DOV) and offshore (TILES) areas.

3.2 Seismic reflection data

Marine seismic imaging is a geophysical technique used to obtain detailed information of the physical characteristics of the subsurface (Figure 3.1). Due to their accuracy and resolving power, seismic methods have increasingly become a major investigation tool in marine geoarchaeological studies (Zurita Hurtado and Missiaen, 2016). Recently, within the framework of the IWT SeArch project (Missiaen et al., 2017), different seismic sources and receiver combinations were tested on the BCS to assess, for the first time ever, their suitability to visualise buried prehistoric landscapes, geomorphological landforms and depositional environments for geoarchaeological potential assessment. The results of this study are summarised in Zurita Hurtado and Missiaen (2016).

The seismic reflection data used in this study consist of single-channel (very) high-resolution seismic reflection profiles collected from the Belgian and neighbouring British, Dutch and French shelves We have combined these data in a single dataset used for the characterisation of the geological architecture and geomorphology of the BCS and its surrounding areas (Chapters 4 to 6), with the goal to construct a new stratigraphic framework. Based on the recommendations of the SeArch project the majority of the seismic reflection data gathered for the present study were acquired by means of sparker and boomer sources in combination with a single-channel streamer (Table 3.1).

Table 3.1 Comparative table of the performance of the seismic sources used in this study (modified from Zurita Hurtado and Missiaen, 2016).

Source	Frequency Range (kHz)	Vertical Resolution (m)	Penetration (m)
SIG sparker 1200	0.8–0.9	>0.50	<200
GSO 360 sparker	0.8–1.2	>0.50	<200
RCMG Centipede sparker	1.0–1.2	>0.35	<50
IKB-Seistec boomer	1.0–5.0	>0.25	<20

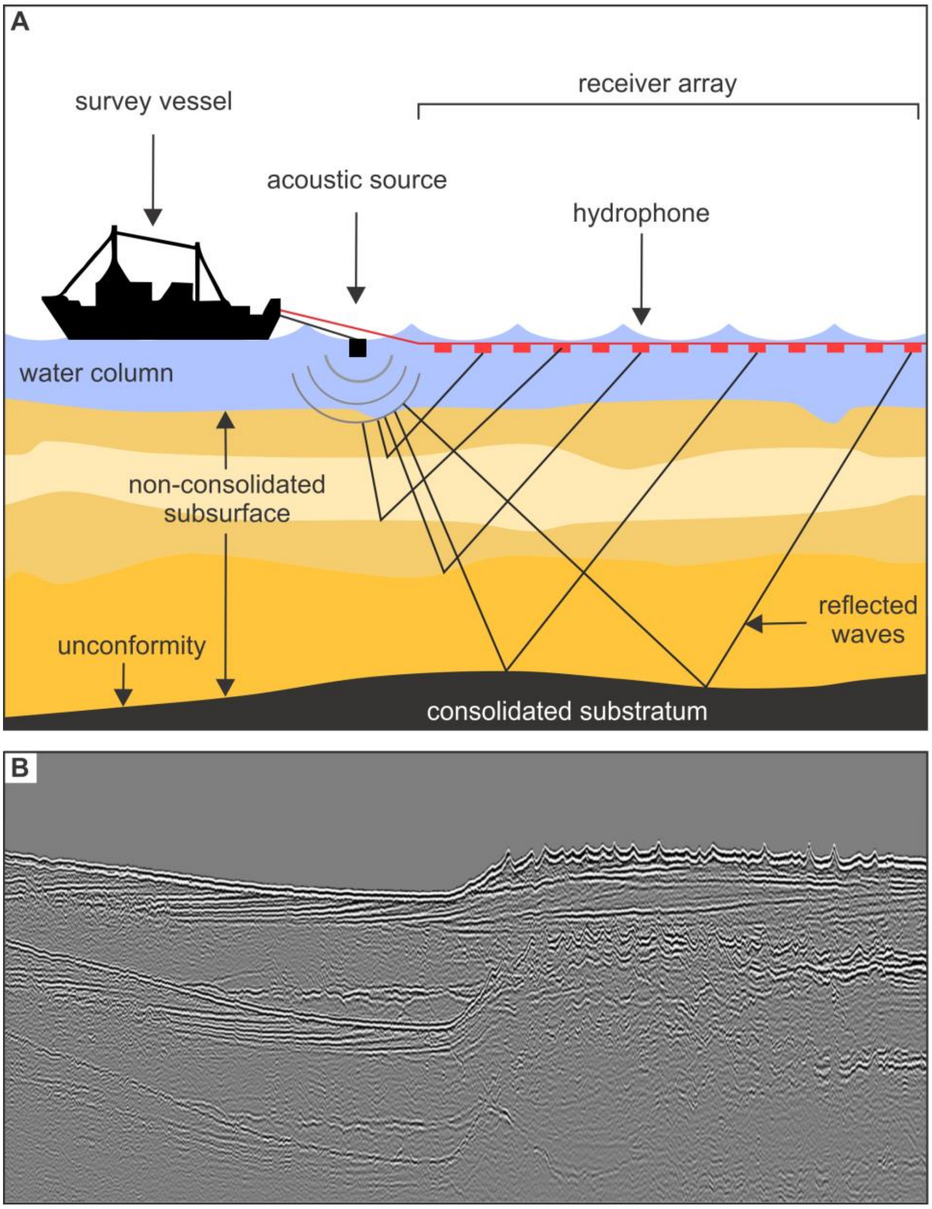


Figure 3.1 A) Schematic illustration of seismic reflection data collection. B) Example of a two-dimensional seismic reflection profile collected for this study. Vertical scale is measured in two-way travel time. Horizontal scale shows geographic position.

Sub-bottom acoustic sources can be classified on the basis of the frequencies and of the emitted acoustic energy strength. These two parameters affect the performance of the instrument, which is largely defined by the (vertical) resolution and penetration depth of the signal into the subsurface. The resolution is the ability to differentiate closely spaced layers or objects and is mostly dependent on the frequency content of the signal. In general, low frequency sources, like an air gun, have a low vertical resolution (> 5 m) but can penetrate several hundred meters into the subsurface. Medium frequency sources like boomers or sparkers, as used in this study (see Table 3.1), can penetrate to depths of a few tens of metres up to a hundred meters with a relative good resolution, ranging between 0.5 and 5 m. Finally, high frequency sources like pingers, chirp sonars and echosounders provide detailed information (< 0.5 m) of the shallow subsurface down to a few tens of meters below the seafloor.

Penetration, on the other hand, is controlled by the strength of the emitted signal. Increasing the power of the acoustic signal without altering the frequency allows for greater penetration into the subsurface. Penetration depth is also affected by the sedimentary composition of the seafloor and subsurface. In general, penetration increases for fine-grained and/or less-consolidated sediments but decreases for coarse-grained and/or consolidated sediments. Increasing the energy of the source when working in areas with a sandy seafloor, like on the Belgian shelf, is even counterproductive as it stimulates more energy to be reflected back off the seafloor, resulting in a decreased signal to noise ratio or a phenomenon called ringing. The acoustic sources used in this study typically operate at frequencies below 10 kHz targeting geological layers at depths up to ca. 50 m below the seafloor (Table 3.1). In ideal circumstances depths of 100 m (or even more) may be reached.

The dataset for this study covers the entire territory of the Belgian continental shelf (ca. 3.500 km^2) and extends into the neighbouring British, Dutch and French sectors of the continental shelf and consists of ca. 12.000 line-kilometres of two-dimensional seismic reflection data (Figure 3.2). It is mainly composed of the dataset gathered by Mathys (2009) that already included 5.300 line-kilometres of two-dimensional seismic reflection data. This dataset consists of several seismic reflection datasets collected between 1980 and 2007. These data were acquired mainly using sparker and boomer sources. Vertical resolution of this dataset ranges between 0.5 and 3 m and penetration ranges between 12 and 80 m. About 4.000 line-kilometres of this database were created by digitising paper seismic reflection profiles and converting them to SEGY format. Technical specifications on the acquisition, processing and digitisation of these data are described in the doctoral dissertation of Mathys (2009).

This dataset was complemented with ca. 1.600 line-kilometres of seismic profiles collected in collaboration with FPS Economy - Continental Shelf Service. These seismic reflection profiles were acquired with either the IKB-Seistec Boomer, or the Centipede or SIG 1200 sparker in combination with a single-channel streamer. These data have a resolution better than 0.5–1 m and a penetration of ca. 30 m. Part of the technical specifications on this dataset acquisition and processing were described in reports Depret-G-tec (2009) and Mathys et al. (2009). Seismic reflection data for the construction of wind farm and cable construction from the private industry was also collected and consists of ca. 1.900 line-kilometres of reflection data. Detailed information about the NemoLink Interconnector cable dataset was provided by Elia and NGIL (2012). Finally, ca. 2.800 line-kilometre seismic reflection data were collected in the period 2013–2017. These surveys were undertaken within the framework of the RCMG–Deltares–VLIZ collaborative project "IWT SBO–120003 'SeArch': Archaeological heritage in the North Sea". This dataset includes data acquired with various sources (see Table 3.1) and both single-channel and multi-channel receivers (multi-channel along one single test line). Vertical resolution ranges from 0.15 to 1 m, and penetration from a few hundred meters to 15 m. Technical specifications on the acquisition and processing can be found on http://www.sea-arch.be/nl/resultaten. A summary of these surveys is provided by Zurita Hurtado and Missiaen (2016).

Figure 3.2 Overview of the seismic reflection profiles used in this study.

In the present study, we focus on the geomorphology of the pre-Quaternary erosional surface and the geologic structures and seismic-stratigraphy from the Quaternary sediments. In Chapter 4 a DCSM of the pre-Quaternary surface is generated (cf. De Clercq et al., 2016) by combining and re-evaluating the interpretations from Mathys's (2009) database and the additional data described above.

Processing of the newly acquired seismic reflection data was performed by Oscar Zurita Hurtado using the Radex Pro software package (see also Zurita Hurtado and Missiaen, 2016 for an overview). Below we provide a generalised overview of the processing applied to the gathered data for the period 2013–2017.

3.2.1 Signal processing

Single-channel (SC) data was processed with Deco Geophysical seismic processing software RadExPro using the following flow:

1) Common Mid-Point (CMP) binning: navigation information is added to the seismic data. In this study, all the navigation information corresponded to the x and y positions of the GPS of the vessel. Source and receivers were assumed to lie at a constant distance on a linear path behind the vessel allowing us to calculate the source, receiver and reflection point locations in a simple way. The error introduced by this assumption is minimal and avoids placing extra positioning systems on the actual streamer

2) Data editing: noisy and/or dead traces are edited or simply deleted

3) Deconvolution: signal wavelet is compressed in order to increase resolution

4) Noise Attenuation: removal of noisy frequencies

5) Seabed picking: the seafloor reflector was highlighted and the arrival time stored in order to be used by other processes such as muting, swell correction, tidal corrections, multiple attenuation, etc.

6) Swell correction: a swell filter was applied to attenuate heave effects. This helps to flatten the seabed and the subsequent horizons

7) High amplitude noise attenuation: removal of local narrow frequency band noises observed as vertical stripes on the seismic record

8) Spatial noise removal: removal of side scattered noise and spatially random noise

9) Amplitude correction: gain was applied to increase the amplitudes of the data at greater depth (later arrivals) when compared to the earlier arrivals

10) Multiple attenuation: multiples are delayed reflections that interfere with the “primary” reflections we want to image. The delay occurs because the reflection energy has taken a more complex and longer ray path from source to receiver, reverberating between two layers that are highly reflective. While any two layers with high reflectivity can create a multiple, in general it is energy reverberating in the water column (i.e. between the seafloor and the sea surface) that is the most important (so-called seafloor multiple). In single-channel data, seafloor multiple reflections are quite strong due to the incapacity of these systems to discriminate between events based on their arrival time or direction of arrival. When working in very shallow water, they hide primary reflections and therefore it is desirable to attenuate them as much as possible. Advanced processing techniques can be applied in order to improve the quality of the data and overcome some of these issues but because of the intrinsic nature of the single-channel system most demultiple techniques are not as efficient on SC data as on MC data

11) Muting: data above water bottom was removed

12) Tidal correction: tidal changes introduce time variant shifts to the seismic trace that need to be corrected

13) Data archive: data was exported in SEG-Y format so it could be later read on most interpretation software

Multi-channel (MC) data was processed with the same software using the following flow. Some processing steps are common to single-channel acquisition geometries and are indicated with an asterisk (*)

1) CMP binning: navigation information for each source, channel and reflection location are calculated from the supplied x and y positions of the GPS of the vessel. Source and receivers were assumed to lie at a constant distance on a linear path behind the vessel

2) Data editing (*)

3) Deconvolution (*)

4. Noise Attenuation (*)

5. High amplitude noise attenuation (*)

6. Velocity analysis every 250 CDPs: creates a model of the seismic velocities in the sub-surface. It is performed in the CDP domain, where the assumption of hyperbolic move-out of reflections is often reasonable. The aim of the velocity analysis is to find the velocity that flattens a reflection hyperbola and thus provides the best result when stacking is applied

7) Data sorting: sort the data from common shot point (CSP) to CMP gathers

8) Normal move-out (NMO) correction: When the distance between the source and receiver increases, the travel time also increases. On the seismogram the effect of this delay in travel time is visible as a hyperbolic curve. The NMO correction is applied to correct for this effect

9) Muting: Frequency distortion occurs as a result of NMO correction, particularly at shallow depths and large source-receiver offsets. A stretch mute is applied (manually picked or based on angle of reflection) to remove sections that are affected by NMO stretching

10) Stacking – All the traces from the same CMP are summed up into one trace, generating a pseudo zero-offset section (similar to the one obtained with a single-channel streamer)

11) Spatial noise removal (*)

12) Amplitude correction (*)

13) Multiple attenuation (*)

14) Muting (*)

15) Tidal correction (*)

16) Data archive (*)

3.3 Bathymetric data

Acoustic geophysical devices are universally used to determine the depth of the seafloor and generate a digital elevation model map of the seafloor, i.e. a bathymetry map. They are divided into single-beam and multibeam systems. Single-beam systems are not as widely used anymore but we will explain the principle on which they are based here shortly. They are generally configured with a transceiver (transducer/receiver) that also contains a transmitter, which emits a high-frequency acoustic pulse in a beam downward into the water column directed at the seafloor (Figure 3.3 A). The transceiver controls the pulse length and provides electrical power at a given frequency. The emitted acoustic energy is reflected off the seafloor beneath the vessel and received again at the transceiver. The continuous recording of the time difference between the transmission and the reception of the acoustic signal yields (very) high-resolution two-dimensional topographic profiles along a set of survey tracks. Original data thus provide depths in TWTT (two-way travel time, i.e. the time needed for a transmitted acoustic signal to travel from the transmitter (transducer) to the seafloor and back to the receiver), which is depth-converted into meters by applying the sound velocity in water. Sound velocity in water is a function of the temperature, salinity and pressure/depth ratio of the water, but can for most conventional applications be approximated to 1.500 ms-1 (Stewart, 2011). These parameters can be measured by performing sound velocity profiles before or during the data acquisition. To produce a three-dimensional bathymetric map, single-beam data requires an incredibly high-resolution of data points. Multibeam echosounders (MBES) on the other hand, emit sound waves in the shape of a fan from directly beneath a vessel's hull. These systems measure the TWTT. In this way, multibeam sonars produce a "swath" of soundings (i.e. depths) for broad coverage of a survey area (Figure 3.3 B-C; e.g. Hughes-Clarke et al., 1996).

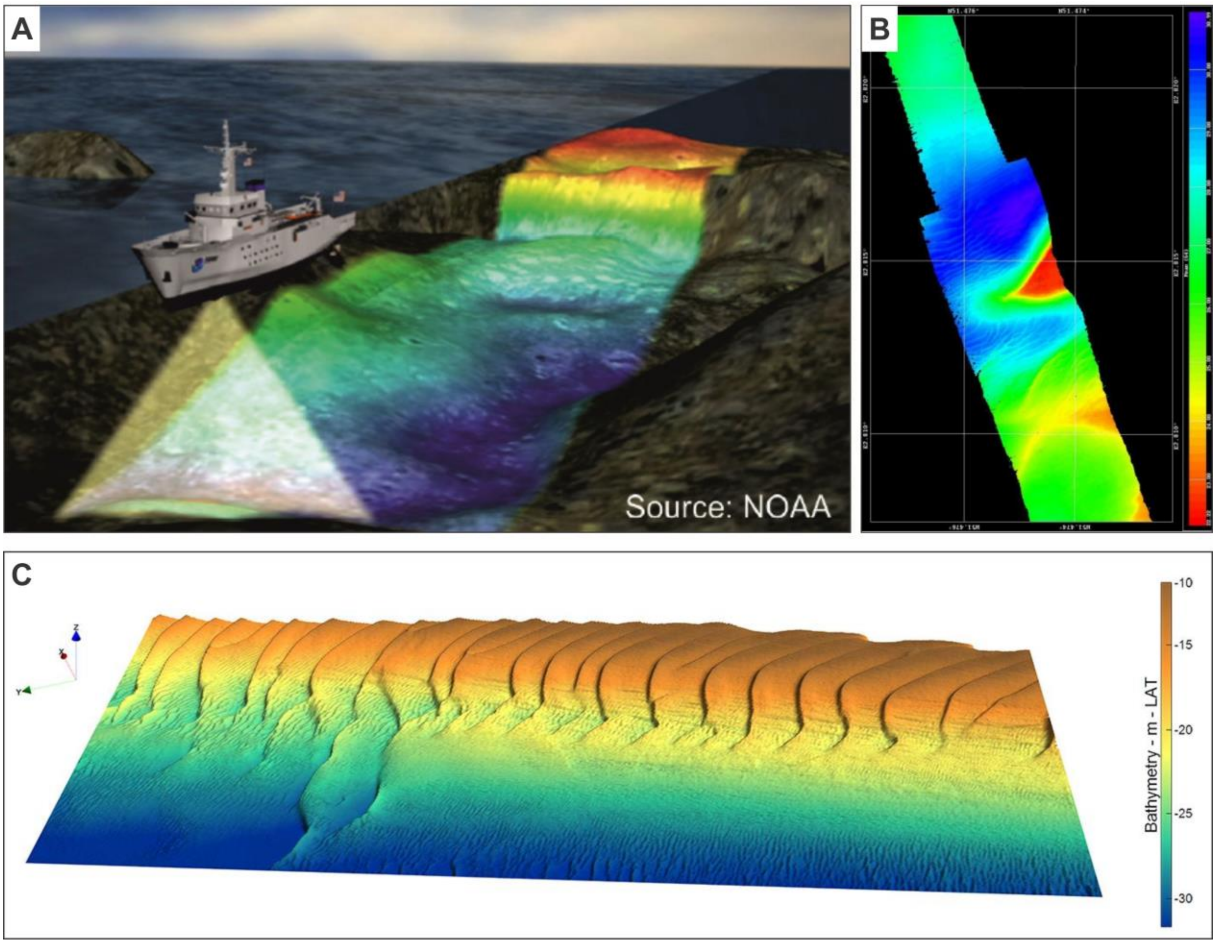


Figure 3.3 A) Schematic visualisation of MBES data collection using a hull mounted system. B) Example of a single swath of MBES data of the Belgian Thornton sandbank by the RV Simon Stevin (source: COPCO). C) Example of an interpolation of MBES data from multiple swaths represented as a three-dimensional view on the aggregate extraction zone 4 of the Oosthinder sandbank (source: FPS Economy - Continental Shelf Services).

The area on the seafloor covered by multibeam echosounders depends on the depth of the water, and typically corresponds to two to four times the water depth. Various transmitted frequencies are utilised by different MBES systems depending on the sea floor depth. A time-depth conversion based on MBES data is a complex process as the orientation of the swath varies with the movements of the ship, which in turn also depends on the sea state. In order to receive a high-precision three-dimensional bathymetry map of the seafloor a correction is needed for the movements of the ship, which again depends on the state of the sea. Today, this problem is resolved by using RTK-GPS and motion sensors.

The bathymetric data used in this study consists of three separate datasets with different resolutions: The North Sea-English Channel dataset (Figure 3.4); the DCS dataset; and, the BCS dataset (Figure 3.5), all consisting a mixture of single-beam and multibeam bathymetry data. The North Sea-English Channel bathymetric data consists of a harmonised elevation model generated (230 m bin size) for all European sea regions from selected bathymetric survey data sets and composite DEMs, while gaps with no data coverage are completed by integrating the GEBCO Digital Bathymetry. The DEM with its information layers is made freely available for browsing and downloading through http://www.emodnet-bathymetry.eu/data-products. More information about the DEM details can be found at: www.emodnet-bathymetry.eu/internal_html/qaqc-and-dtm-production-details/9.

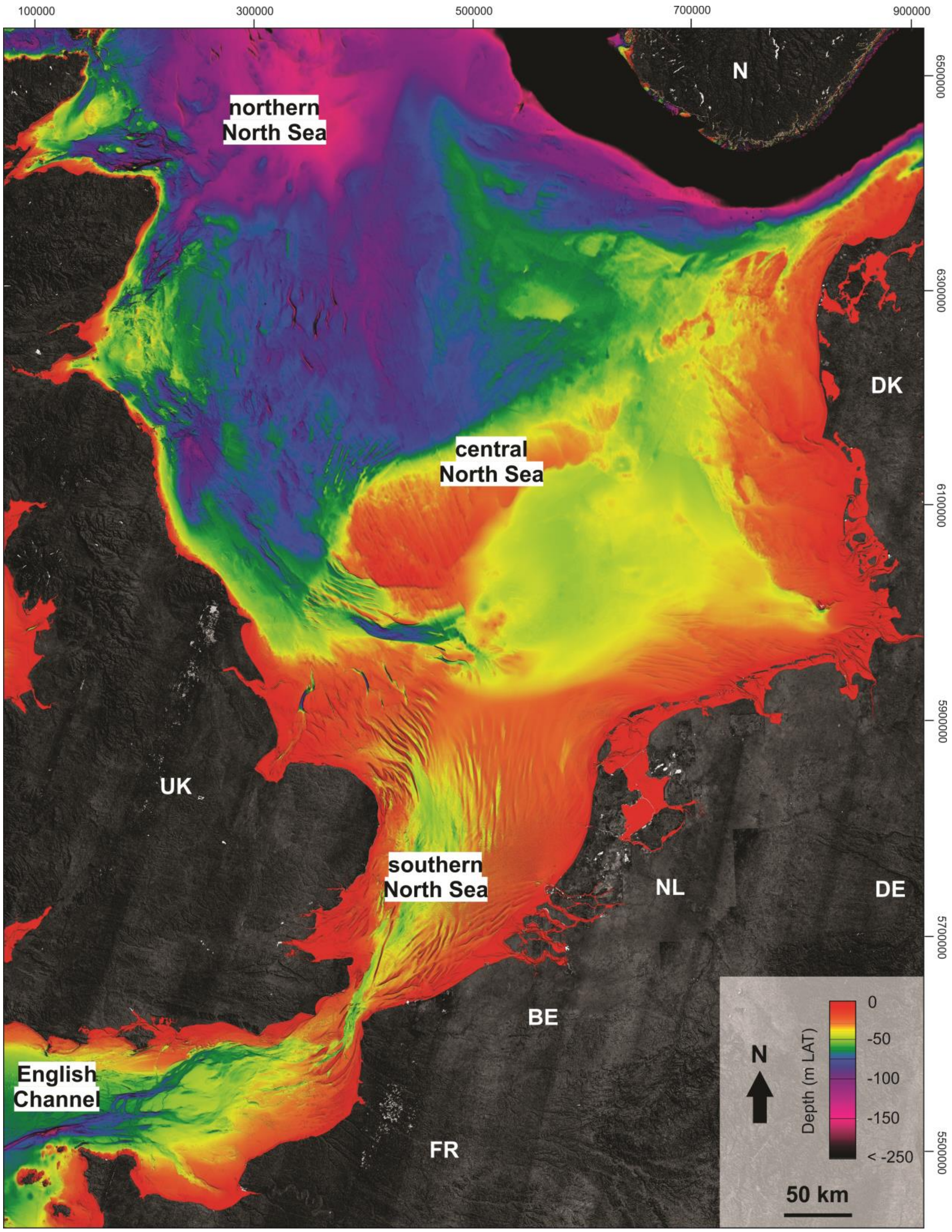


Figure 3.4 The North Sea-English Channel bathymetric dataset used in this dissertation.

The Dutch institute Deltares provided the DCS bathymetry dataset (data available at www.deltares.nl). It comprises single-beam and multibeam bathymetric data gathered by Deltares until 2013. Deltares provided this dataset fully processed and gridded at 25 m bin size.

The BCS bathymetric dataset is a compilation of single-beam and multibeam bathymetric data acquired between the 1990s and 2015 by the Belgian Flemish Hydrography and the Belgian FPS Economy − Continental Shelf Service. The Belgian Flemish Hydrography Service fully processed and gridded at 20 m bin size provided this dataset as ArcGrid files.

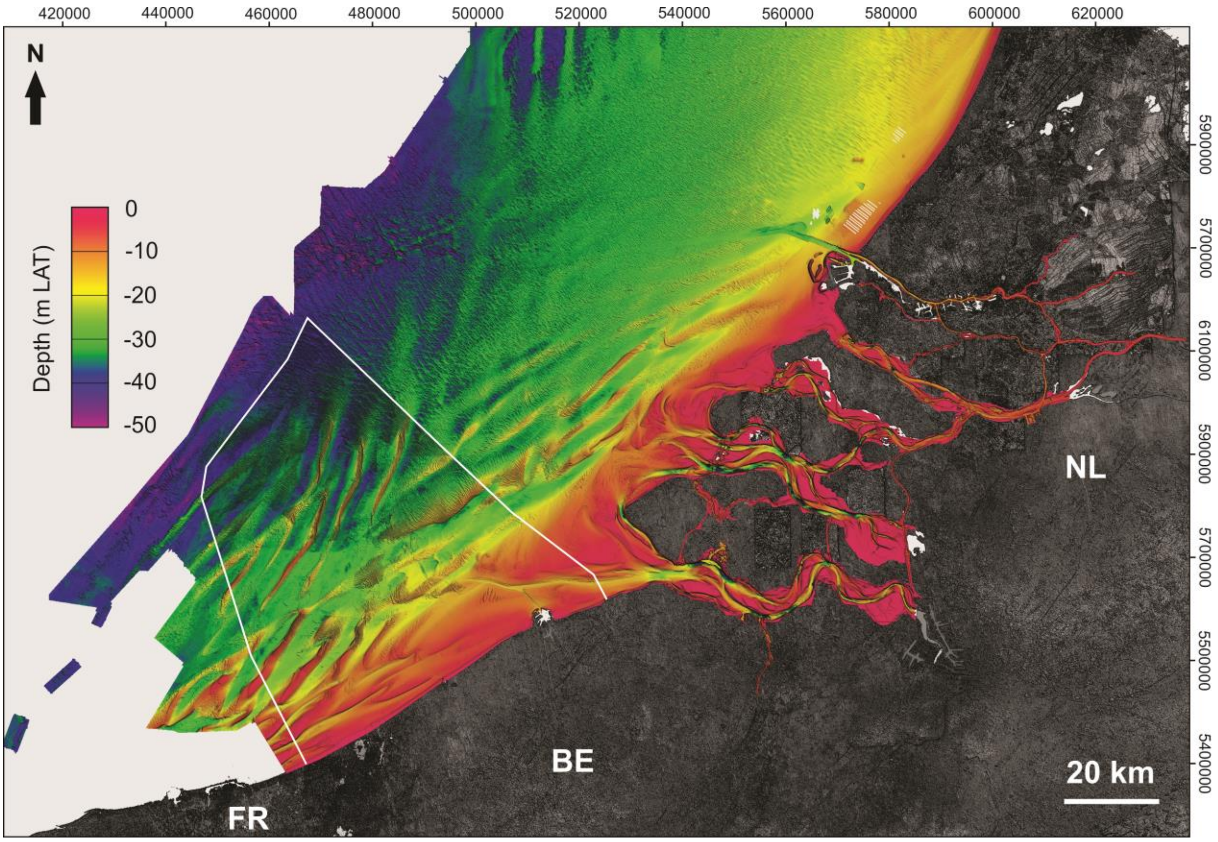


Figure 3.5 The BCS (indicated by a white border) and the DCS bathymetric datasets used in this dissertation.

3.4 Sediment core and vibrocore data

Coring is a method for recovering subsurface sediments. The coring equipment or vibrocorer is deployed from a platform or vessel and upon reaching the seafloor, applying various techniques, a plastic liner, positioned in a metallic casing, is pushed into the seafloor sediments. Depending on the technique applied and the cohesive characteristics of the sediments a specific recovery, preferably a hundred percent, will be reached and brought up to the platform/vessel.

The offshore sediment core-vibrocore database used in this study is a recently developed database that is still being refined and which will later become available through www.mumm.ac.be/datacentre. The database consists of lithological descriptions from available archived sediment cores and vibrocores across the BCS, DCS, FCS and UKCS (Figure 3.6; Kint and Van Lancker, 2016). To construct this database public available data sources were consulted, including data from neighbouring countries (mostly via national data portals). Data from offshore industry are continuously being incorporated as well. Data from industry projects (if needed, upon permission) have been added through digitisation from available reports. All entries in the database were carefully verified and metadata are maximally added for uncertainty analyses. Metadata coding followed EU guidelines (www.seadatanet.org). Main fields were positioning, timestamp (vintage), methodological and analytical techniques used. Quality flags were entered (following www.geoseas.eu).

The onshore core database used in this study were collected from the Flanders subsurface database (DOV, 2014) and can be found on www.dov.vlaanderen.be. The cores within this database were collected from various sources for subsurface research, sediment sampling or determining ground water quality, determining geophysical parameters, environmental research, etc.

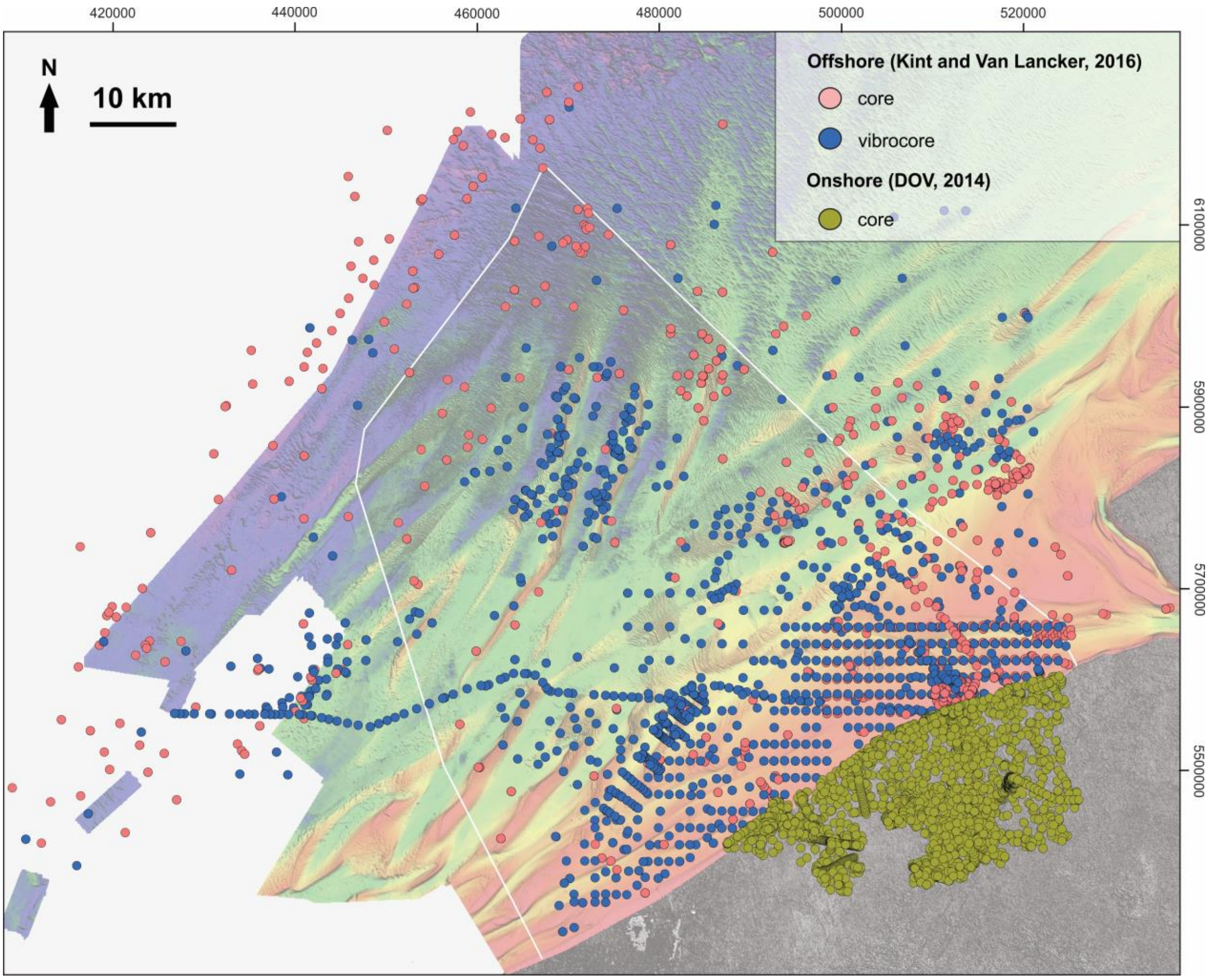


Figure 3.6 Overview of the cores and vibrocores used in this study.

Please note that within this dissertation no archived sediment core or vibrocore was neglected, even if its information was fragmented or was of lower quality (some onshore sediment cores date back almost 100 years). The fact that large gaps are present within both databases was used as an argument to uphold all information in a first stage of interpretation (profile construction). Only if the information caused serious deviations from the overall geology the sediment core or vibrocore was rejected. This process involved the following steps:

1) exploration of the database to determine the areas with the highest sediment core/vibrocore density, the best quality of description and metadata

2) these areas, in conjunction with the pre-Quaternary geomorphology (Chapter 4), were used to the best knowledge to construct geological cross-sections (Chapters 5 and 6)

3) offshore geological cross-sections were supported with seismic reflection analysis. This process was not adequate for onshore cross-sections as no seismic information was available. Therefore, all sedimentary cores, even low quality cores, were drawn along each section. When gaps were present in cross-sections these low quality cores proved sometimes valuable when e.g. a clay-sand interface needed to be continued. This is especially important at the Holocene-Pleistocene interface. When a low quality sediment core caused correlation issues it was removed or discarded from the cross-section.

3.5 Data integration

The stratigraphic model of landscape evolution presented in this dissertation (Chapters 4 to 6) was developed through the interpretation of bathymetric data and shallow subsurface two-dimensional seismic reflection profiles calibrated with sediment cores and vibrocores. The data were collected over the past few decades by various survey companies, scientific institutes and government agencies with different objectives, and demonstrated variations in quality, format, and resolution, as technologies have advanced over the years. Integrating such a dataset was problematic. Landscape reconstruction follows a more or less universal process of data integration. Here, we followed the process as explained by Mellett (2012) and García-Moreno (2017) with some minor adaptations to the own study area (cf. Mathys, 2009).

3.5.1 Data analysis and interpretation

3.5.1.1 Seismic reflection data

Each seismic line, including the already interpreted seismic dataset from Mathys (2009), was interpreted following the principles of Mitchum et al. (1977a, 1977b, 1977c) and Mitchum and Vail (1977) for acoustic facies analysis (Figure 3.7). For each seismic reflection line the characteristic seismic facies, using its geometry, frequency, amplitude and continuity, were identified and cross-correlated with cross-lines. By identifying key bounding reflections as seismic horizons, using the software package Kingdom Suite, boundaries for seismic units were identified.

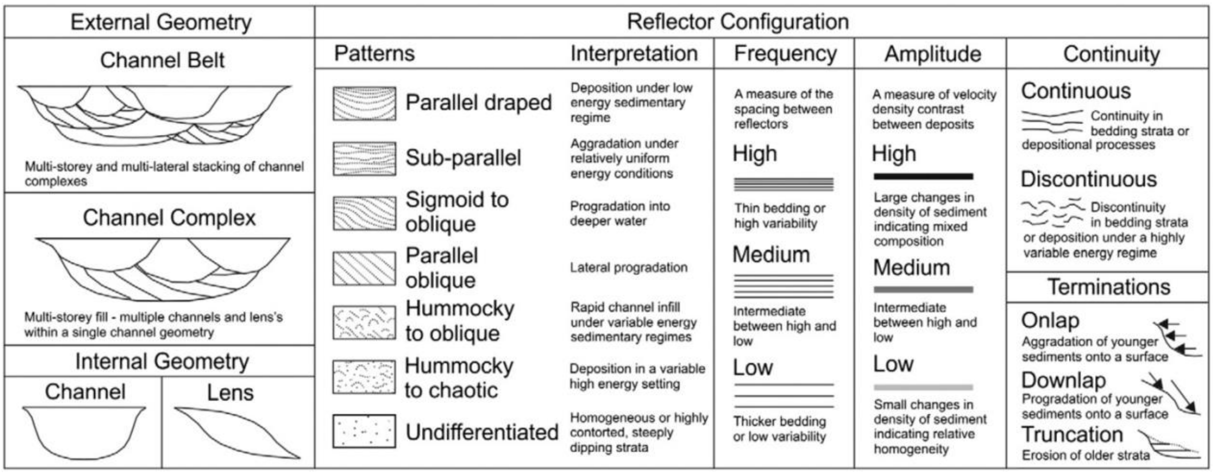


Figure 3.7 Guide on the interpretation of seismic reflection data based on Mitchum et al. (1977a, 1977b, 1977c) and Mitchum and Vail (1977).

The ability to recognise seismic-stratigraphic facies/units on seismic reflection profiles from different surveys depended not only on the data quality (that varies with weather, sea state, water depth, source-receiver characteristics, lithology) but also on the geometry of the seismic reflection profiles compared to the sedimentary units/acoustic facies. Finally, processing and filtering of the acquired data also influenced its quality depending on the target that was being studied (see Zurita Hurtado and Missiaen, 2016). Most of these aspects were conditioned at the time of surveying and can’t be changed afterwards. A sizeable proportion of the seismic reflection dataset used in this dissertation consists of digitised paper records (see Mathys, 2009). These digitised paper records contain inherent data deterioration due to ageing of the paper records and flaws in the digitisation process. Furthermore, it was noticed that some of the coordinates of the tracks may contain mistakes rendering the digitised record useless. The final result therefore is often of a lesser quality, sometimes leading to the removal of that particular seismic reflection data.

The success rate of interpreting seismic-stratigraphic facies/units varied across the Belgian shelf depending not only on data density but also on the intrinsic properties of the facies and how it appears on the acquired seismic data through the decades. An appropriate way to tackle this problem was by comparing cross-lines with one another and if possible to visualise them in a three-dimensional environment (VUPAK module in Kingdom Suite) in order to better understand and comprehend the geological architecture. This is a suitable technique because facies and reflection geometries vary greatly depending on the angle of visualisation. To achieve the best visualisation a well-oriented seismic survey planning of the seismic lines was needed. During the interpretation stage isolated lines were difficult to interpret and correlate to neighbouring high-density data areas and resulted in generalisations. At the end of the interpretation phase the seismic units were visualised using algorithms within the Kingdom Suite software (e.g. topography and thickness maps). Because the seismic units on the BCS are highly fragmented (Mathys, 2009) this was a difficult process.

3.5.1.2 Bathymetry

The multibeam datasets were used to identify remnant geomorphological landforms preserved on the sediment-starved seafloor, i.e. incised-valleys, terraces, escarpments, etc. This task was performed using Global Mapper 2013 and 2016.

3.5.1.3 Sediment cores-vibrocores

For this study no new sediment cores or vibrocores were acquired. Instead, two extensive archived sediment core-vibrocore databases (onshore: DOV; offshore TILES) were used to meet the needs of this study (see Chapters 4 and 5 for more detailed information). Each database contained sedimentary and structural descriptions. The quality of these descriptions depended on the end-user’s terms (geotechnical, geological, commercial-industrial, etc.). For some sediment cores and vibrocores photographs were available. These were helpful to characterise facies changes in more detail (see Bogemans and Baeteman, 2014). A final interpretation of the depositional environment, when possible, was achieved through characteristic sedimentary structures (cf. Bogemans and Baeteman, 2014). At two locations, sediment samples were acquired for palaeoenvironmental analyses and one for OSL-dating (see Chapter 5). These results were used to back the palaeoenvironmental interpretations and chronological control of the landscape evolution.

3.5.2 Data interpolation

3.5.2.1 Seismic reflection data

The final step consisted of constructing a DCSM of the pre-Quaternary surface and two isopach maps for the master units Pleistocene and Holocene using ArcGIS. This was done by interpolating between data points using a nearest neighbour algorithm in ArcGIS using a bin size of 50 m for the pre-Quaternary surface inside the BCS and 25 m outside the BCS (see Chapter 4). The latter was applied in this area because of a low seismic reflection data density and lack of sediment cores-vibrocores. For the first order units Pleistocene and Holocene, a bine size of 50 m was applied. Up to this stage the thickness of each seismic unit is presented in ms TWTT. To convert time to actual depth a dynamic depth conversion tool in the seismic interpretation software Kingdom was used, which allowed to integrate the interpreted horizons from the seismic data (in time) and sediment core-vibrocore information (in depth) into a grid-based layered velocity model. The average velocity model was calculated using the depth from the sediment cores-vibrocores in order to calculate the depth of the interpreted horizons. All available data were then plotted using ArcGIS, where the sediment cores-vibrocores were combined with the calculated depth model. Nearest neighbour was used as the interpolation method. Spatial analysis of the pre-Quaternary DCSM was carried out using Global Mapper. Topographic profiles were produced to help with morphological interpretations of buried features.

3.5.2.2 Data integration

During the final step, before a palaeogeographic map series for submerged heritage assessment was constructed, the obtained seismic information was combined with the information obtained from the archived sediment core-vibrocore databases (Figure 3.8). Through this process the spatiotemporal variation of depositional environments preserved within prehistoric landscapes could be traced over greater distances by tracing the layer boundaries. Once the spatiotemporal interaction between the depositional environments was understood a palaeogeographic map of the landscape could be constructed. This technique was however limited by the availability and quality of both the seismic and the sediment core-vibrocore datasets. For example, in certain areas of the BCS (see Figures 3.2 and 3.6) the data density is low, making it difficult to trace seismic and/or sedimentary facies across greater distances, resulting in a generalisation. This means that in some cases seismic facies lacked sedimentary information and *vice versa*. In spite of this disadvantage the datasets presented in this dissertation were the best available for the

BCS at the time of writing and are believed to provide a strong foundation for the development of a complete stratigraphic model, a palaeogeographic map series and the assessment of the preservation potential for both prehistoric landscapes, geomorphological landforms and depositional environments as well as for the submerged heritage within.

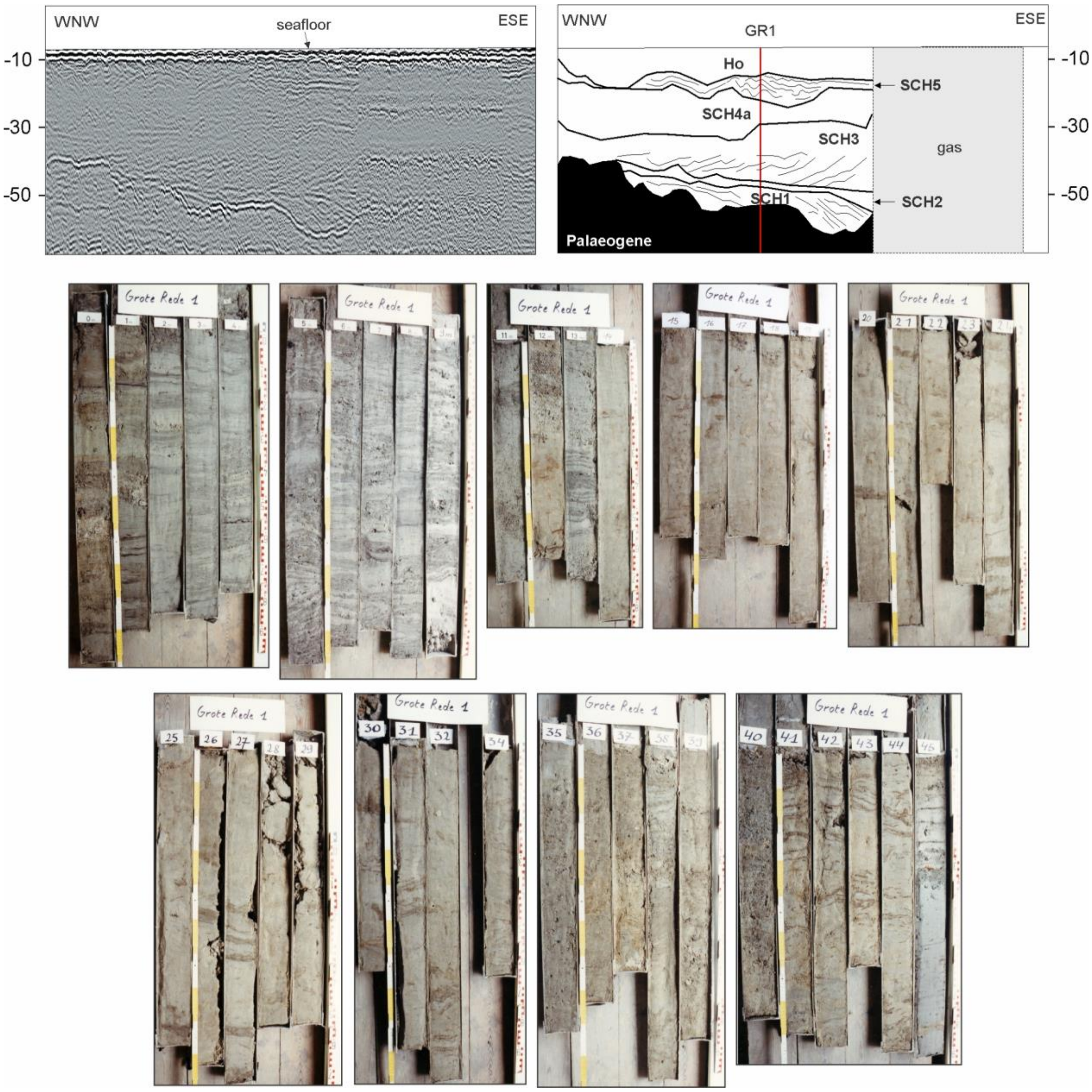


Figure 3.8 Integration of core data with seismic data. Depths are in m LAT.

4 A HIGH-RESOLUTION DEPTH-CONVERTED STRUCTURE MAP FOR THE TOP-PALEOGENE SURFACE OF THE BELGIAN CONTINENTAL SHELF

This chapter is a slightly modified version of the manuscript that was published as:

De Clercq M, Chademenos V, Van Lancker V, Missiaen T. 2016. A high-resolution DEM for the Top-Paleogene surface of the Belgian Continental Shelf. Journal of Maps 12: 1047–1054. DOI: 10.1080/17445647.2015.1117992.

Abstract

A 1:250.000 scale depth-converted structure map of the pre-Quaternary surface for the Belgian Continental Shelf was created based on extensive analyses of older and recent geological and geophysical datasets. The pre-Quaternary surface is an important polygenetic unconformity that truncates older strata of the Paleogene and to a smaller extent some of Neogene age from the overlying Quaternary deposits. As such it represents the base of the latter. The represented surface has been diachronously shaped and reworked through Late Quaternary times by different geological processes (e.g., fluvial, marine, estuarine, periglacial). Additionally, the offshore surface has been attached to the landward pre-Quaternary surface and was transformed into a uniform 3D surface allowing new and better interpretations to be used in fundamental and applied research underpinning both scientific purposes (e.g. geology, archaeology, palaeogeography), and commercial applications (e.g. wind farms, aggregate extraction, dredging).

4.1 Introduction

The BCS is a sediment-depleted shallow shelf environment without a distinct shelf break and is located in the southern North Sea Basin (De Batist, 1989). Its coastal plain is characterised by a small non-linear, glacio- and/or hydro-isostatic subsidence component that increases significantly towards the north-east in the direction of the Fennoscandian land mass (Kiden et al., 2002; Vink et al., 2007). As a consequence there is very little accommodation space to accumulate newer and preserve older sediments, which in turn caused important reworking and redistribution of these older sediments. The lack of accommodation space led to a complex thin, discontinuous and patchy Quaternary sediment cover (Mathys, 2009). Still, it protects some significant palaeolandscape elements that connect the erosional area of the Dover Strait in the south and the depositional southern North Sea Basin to the north (Hijma et al., 2012). Knowledge of the location, preservation conditions and spatio-temporal distribution of the remaining Quaternary deposits is critical to provide a better stratigraphical context for past and future submerged heritage finds not only on the BCS but also in the surrounding continental shelves (see Hijma et al., 2012).

In addition, the pre-Quaternary surface in Belgium regained interest to meet new societal and economic challenges. On the one hand, the deposits directly above this surface might contain a unique archive of submerged heritage. On the other hand, the pre-Quaternary surface provides the base for this sedimentary archive that is economically exploited by the Belgian offshore industry. Two Belgian research projects recently addressed these issues, each requiring detailed mapping of the pre-Quaternary surface of the BCS. The 'SeArch' project (Archaeological Heritage in the North Sea, www.sea-arch.be) wanted to provide the industry with an efficient survey methodology, a sustainable management framework and a practical guide for the assessment of submerged cultural sites. To characterise the preserved palaeolandscape elements and to quantify their depth distribution, a detailed map of the pre-Quaternary unconformity was required. The main focus of the Search project has been on buried palaeovalleys, planation surfaces, slope breaks and escarpments. A majority of these elements was already described in previous studies (e.g. Liu, 1990; Liu et al., 1992, 1993; Mathys, 2009, Mostaert et al., 1989), though often not in sufficient detail and/or precision, and a high-resolution DCSM of the pre-Quaternary surface was urgently needed to pinpoint areas with high preservation potential for submerged heritage. For this purpose the offshore and landward pre-Quaternary surfaces were fully integrated, resulting in a level of detail never achieved before. Such a connection allows for better interpretations concerning landscape evolution of buried palaeolandscapes. The second research project ‘TILES’ (Transnational and Integrated Long-term Marine Exploitation Strategies, www.odnature.be/tiles) focused on sustainable management of marine aggregate resources through the development of four-dimensional resource models as decision support tools, linking three-dimensional geological models, knowledge and concepts to numerical impact models. Here, voxel models of the Quaternary sediments are created in which lithological and other sedimentological characteristics are quantified to get a better understanding of the distribution, composition and dynamics of the sparsely available geological resources (Van Lancker et al., 2014). This will enable detailed and long-term calculations of the economic resource potential of the Quaternary deposits overlying the pre-Quaternary surface. Both scientific projects fit well within the aims of the recently recognised research field called Continental Shelf Prehistoric Research and the SPLASHCOS COST Action (2009–2013) to build a multidisciplinary research community in Europe through better knowledge on investigation methods, interpretation and management of underwater archaeological, geological and palaeoenvironmental evidence of prehistoric human activity. As such, an interdisciplinary and international research framework is provided to archaeologists, heritage professionals, scientists, government agencies, commercial organisations, policy makers and a wider public.

4.2 Study area

The BCS extends to a maximum of 84 km from the present-day Belgian coastline (67 km long) and covers an area of 3.454 km^2. The seafloor topography is characterised by the presence of sandbanks and swales (Figure 4.1). In the swales, water depths reach 30–40 m below LAT, whilst the offshore sandbanks shallow up to 5 m below LAT. Previous studies highlighted that the Top-Paleogene surface is polygenetic and that the sediments below this surface are composed of compacted clays and sands to sandy clays that were deposited in a near-coastal to outer shelf environment from the Eocene to the Neogene (between 53 and 2.6 Ma; see Le Bot et al., 2003 for a synthesis). In areal extent, consolidated Ypresian clays from the Eocene dominate, but overall there is a lateral difference in lithology, both between and within the units. Farther offshore finer sediments such as silt and clay are predominant, but also silty sand, muddy sand and even calcareous sandstone beds occur (Le Bot et al., 2003). During the Quaternary (2.6 Ma to present) the surface was diachronously ‘reshaped’ in marine, fluvial, estuarine and periglacial environments influenced by climatic change (Mathys, 2009).

4.3 Data and methods

To construct a high-resolution (1:250.000 scale) DCSM of the Top-Paleogene surface, available offshore data, as well as relevant land data, were compiled (Figure 4.1). This comprised in-depth processing and re-interpretation of sediment cores, in conjunction with extensive seismic records. The digital seismic database, acquired over the last 45 years, contains a total of over 7.800 km seismic lines of which 2.534 km were digitised from paper records. After quality assurance, a series of inconsistencies were revealed that caused problems in developing a DCSM of the Top-Paleogene surface. Apart from the uneven distribution of the data over the study area, some of these inconsistencies were related to incorrect or missing corrections for the tidal level, and poor positioning of the old analogue seismic lines, both causing a move or lift of the Top-Paleogene surface by several metres. Additionally, a considerable area of low acoustic penetration was present in the nearshore area, due to shallow gas in the subsurface. Here, the modelled surface is mainly based on sediment core-vibrocore data. From the 749 sediment cores/vibrocores located on the BCS only 130 reached the Top-Paleogene surface; most of them were only 2–3 m deep (i.e. vibrocores). In order to reconstruct the Top-Paleogene surface, the sediment cores together with ca. 6.670 km of seismic data, were analysed in an industrial software package (Kingdom Software, v8.8). The 'Dynamic Depth Conversion' tool within this software package was used to integrate the interpreted horizons from the seismic data (in time) and sediment core-vibrocore information (in depth) into a grid-based layered velocity model. The average velocity model was calculated using the depth from the sediment cores-vibrocores in order to calculate the depth of the interpreted horizons. All available data were then plotted using Esri ArcGIS (v10.1), where the sediment cores-vibrocores were combined with the calculated depth model. Nearest Neighbour was used as the interpolation method. To allow optimal reconstruction of the nearshore Top-Paleogene surface, gridded data from land were incorporated in the geographical information system. The former, constructed on the basis of sediment cores and cone penetration tests, is free for download and needed a transformation of the reference system from Belgian Lambert 72 (EPSG: 31370) to UTM WGS84 zone N31 (EPSG: 32631) (Matthijs et al., 2013). Since this grid was available at a 50×50 m resolution, the offshore data were interpolated to this resolution. The result is shown on Figure 4.2. To depict morphological elements in the terrain model a slope analysis was performed and is useful to define the differences between large- and small-scale features, which can be highlighted by changes in slope.

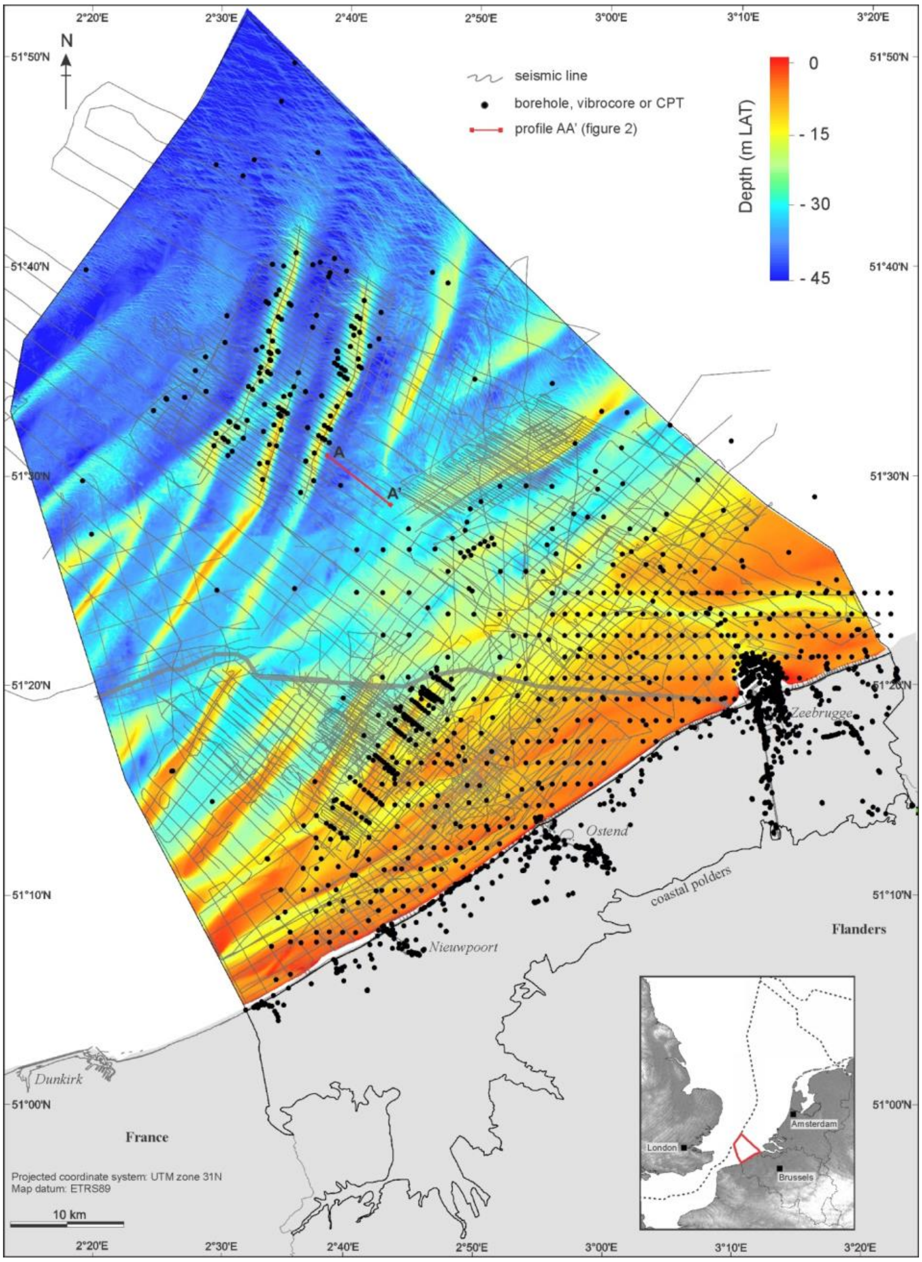


Figure 4.1 Database used to construct the Top-Paleogene surface on Figure 4.2, superimposed on a DEM of the present-day seafloor bathymetry (compiled from Flemish Hydrography and FPS Economy - Continental Shelf Service, 2015). A small number of onshore data (Retrieved from Databank Ondergrond Vlaanderen) were added to the offshore dataset to better interpolate onshore and offshore surfaces. Grey lines are seismic data and black dots are information from boreholes, vibrocores and cone penetration measurements (no distinction made).

4.4 Geomorphology

The top of the Paleogene substratum is defined by an angular unconformity marked by an overall strong seismo-stratigraphic reflector that cuts off the underlying sediments of the Paleogene and Neogene strata (De Batist, 1989; Liu, 1990; Mathys, 2009). This unconformity represents a significant stratigraphic hiatus, specifically here between the gently northeast dipping Lower- and Middle-Eocene formations (Thanetian to Rupelian in De Batist, 1989) and the overlying Quaternary deposits. The depth to the surface varies between -9 and -72 m LAT and its geomorphology is characterised by a series of features ranging from planation surfaces, bounded by escarpments and slope breaks, to buried palaeovalleys and (elongated) palaeodepressions and -ridges.

A number of the discussed morphological features have been described in the literature (Liu, 1990; Liu et al., 1992, 1993; Mathys, 2009; Mostaert et al., 1989). New evidence from a denser sediment core-vibrocore and seismic reflection network has allowed us to redefine some of the previously described features and resulted in the identification of new ones. In order to link the offshore and onshore situations, a single surface was created that can now be used as a transition zone for future palaeogeographical reconstructions between the depositional north and the erosional south of the southern North Sea and provide additional clues to its geological evolution and their link with the Dover Strait opening. It is obvious that, regardless of the large amounts of data, some areas of the BCS are still poorly understood (i.e. the edges of the BCS).

4.4.1 Planation surfaces, escarpments and slope breaks

Three major planation surfaces have been mapped on the BCS. Escarpments or slope breaks bind them and are cut across different Paleogene layers of varying resistance (Figure 4.2) indicating that lateral erosion processes were active at a successively lower base level. The overall orientation of the escarpments and slope breaks is southwest to northeast. The most nearshore platform, the Marginal Platform, is a remnant of the higher inland Top-Paleogene surface that extends offshore. It is divided into three smaller platforms, each given an addition to their naming based on their geographic position (i.e. Western, Central and Eastern Marginal Platform). These platforms have little relief and have a total depth that varies between 10 and 20 m below LAT. At the edges, the slopes range from a steep 14.5% to less than 0.5%. The steepest slopes are cut by palaeovalleys whilst the gentlest slopes face north, away from past fluvial activity. The platforms themselves do not seem to have a straightforward relationship with the underlying Paleogene lithostratigraphic units (Figure 4.2), however some contour lines, such as the 10 m and 22–24 m below LAT, locally follow the general geometry and distribution of the lower Paleogene units (see Le Bot et al., 2003 for more information). Offshore, the Middle Scarp gives way to the Middle Platform, which farther offshore prolongs into the Offshore Scarp and Offshore Platform. Both platforms and escarpments are more continuous compared to the Marginal Platform, which is dissected by large palaeovalleys with a distinct southeast to northwest orientation.

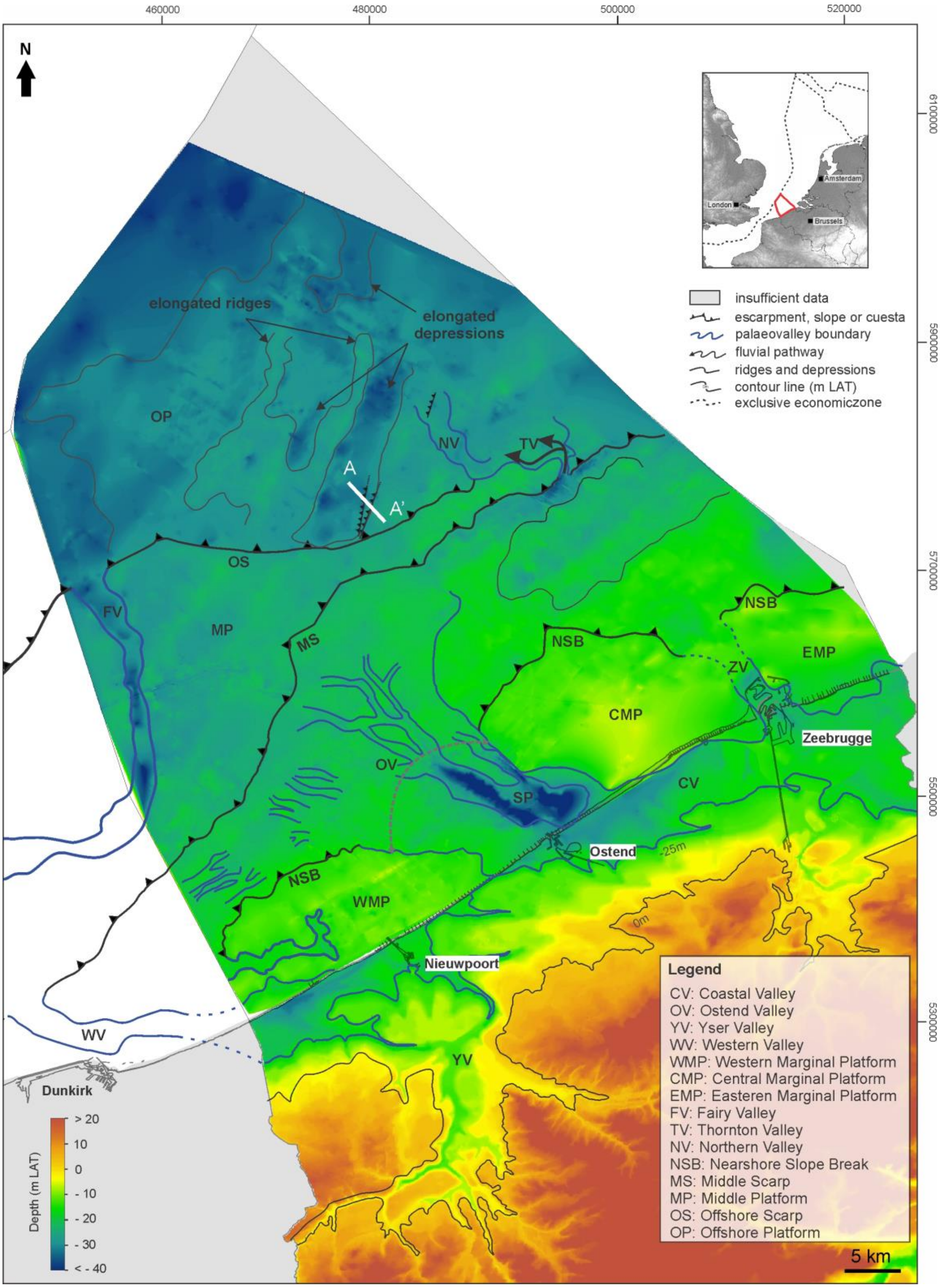


Figure 4.2 The Top-Paleogene surface of the Belgian Continental Shelf attached to northwest Flanders. The upper inset shows the Paleogene layers that define the base of the Quaternary sediments. The lower inset shows the bathymetry with indication of sandbanks. Cross-section A-A’ is presented in Figure 4.3. This figure is modified from De Clercq et al., 2016).

An exception is the narrow Thornton Valley cutting the eastern side of the Middle Scarp, beneath the Thornton sandbank (Figure 4.2). The overall depth of the Middle Platform ranges between 28 and 33 m below LAT, whilst the escarpment displays a gentle drop of some 3 m. The northern boundary of the Marginal Platform, the Offshore Scarp, is characterised by a drop of 4 to locally 10 m. The depth of the Offshore Platform varies between 30 and 40 m below LAT and shows a greater variety in relief due to the presence of elongated depressions and ridges. Comparing the orientation of these morphological features to the underlying Paleogene lithostratigraphic structures shows no obvious direct relationship of differential erosion (Figure 4.2; De Batist, 1989). Beyond the Offshore Platform, the Top-Paleogene surface almost resembles the present-day bathymetry with a maximum depth of 45 m below LAT and continues into a large palaeovalley known as the Axial Channel (Liu et al., 1992; Mostaert et al., 1989). It can be concluded that both the Middle and Offshore Scarps are larger features that traverse the whole width of the BCS. Based on studies from the southern North Sea it is assumed that both scarps are remnants of ancient river terraces of the Rhine and Meuse rivers (Bridgland and D'Olier, 1995; Busschers et al., 2005, 2007, 2008; Gibbard, 1995; Hijma et al., 2012; Peeters et al., 2015; Rijsdijk et al., 2013). This conclusion is similar to what Mathys (2009) suggested about the Offshore Platform being the valley floor of an incised-river valley.

4.4.2 Elongated depressions and ridges

On top of the Offshore Platform a group of tidal sandbanks, referred to as the Hinderbanks, is located. Beneath this group of sandbanks several north to south-southwest oriented elongated depressions and ridges can be recognised in the Top-Paleogene surface. Comparison between digitised analogue data from the 1980s and the present-day bathymetry revealed a direct link with the sandbanks. The elongated ridges are (largely) located beneath the actual sandbanks, while the depressions are located in the swales between or on the edges beneath the sandbanks. Generally the depressions are semi-parallel and typically a few km wide, 10–20 km long and scoured down 5–10 m lower than the surrounding planation surface. The southeast depression is long and narrow (2.3–4.9 km wide, 20 km long) and reaches a depth up to 25 m below the surrounding planation surface. This depression is responsible for two local steep elevation drops in the Offshore Platform just north of the Offshore Scarp (see Figure 4.3). These 'secondary and tertiary escarpments', as revealed by recent seismic data from April 2015, are 3.8 km long and marked by a local drop of ca. 10 m (from the surrounding planation surface to the secondary scarp) and ca. 11.5 m (from the secondary to the tertiary scarp) respectively. They merge with the Offshore Scarp in the southwest but divert from it when going northeast (maximum 1.2 km).

These features clearly provide evidence that this depression is both related to the current hydrodynamic processes and other non-recent geological processes. The northwest depression in the Offshore Platform is generally less narrow (2.2–6 km wide) and not as elongated (17 km long). Farther towards the northeast both elongated depressions connect and create a wider depression in the Top-Paleogene surface, continuing into Dutch waters. Other depressions on the western side of the Offshore Platform also seem to have a relationship with the present-day bathymetry and continue into French waters. This is not surprising considering the overall limited Quaternary cover (less than 3 m in the swales) and the strong hydrodynamic currents present in these zones (Van Lancker et al., 2015).

4.4.3 Palaeovalleys

Besides elongated depressions and ridges several large palaeovalleys occur in the Top-Paleogene surface. Three of these palaeovalleys, the Ostend, Zeebrugge and Yser valleys, cross the present-day Belgian coastline and can be related to (formerly) active rivers. All three valleys are marked by a general north-northwest to south-southeast orientation, with an increasing east-west orientation towards the west, that is, the Strait of Dover. The most prominent palaeovalley is the funnel shaped Ostend Valley (Figure 4.2). It is connected to the inland ancient Scheldt pathway through the Coastal Valley, which runs parallel to the present-day coastline. The Ostend Valley cuts its pathway deep through the Marginal Platform, locally eroding up to almost 50 m into the Paleogene sediments. The slopes of the valley are variable, but generally the western slope is much more gentle (< 0.5%) than the eastern slope which is marked by an internal channel. Here the slope varies between 2% and locally 8%. Two deeply incised channels occur within the valley, respectively 20/10 km long and reaching up to 58/66 m below LAT at their deepest incisions. These two internal channels are characterised by three deep depressions (so-called Sepia Pits) in which the deepest depths are found (Mostaert et al., 1989).

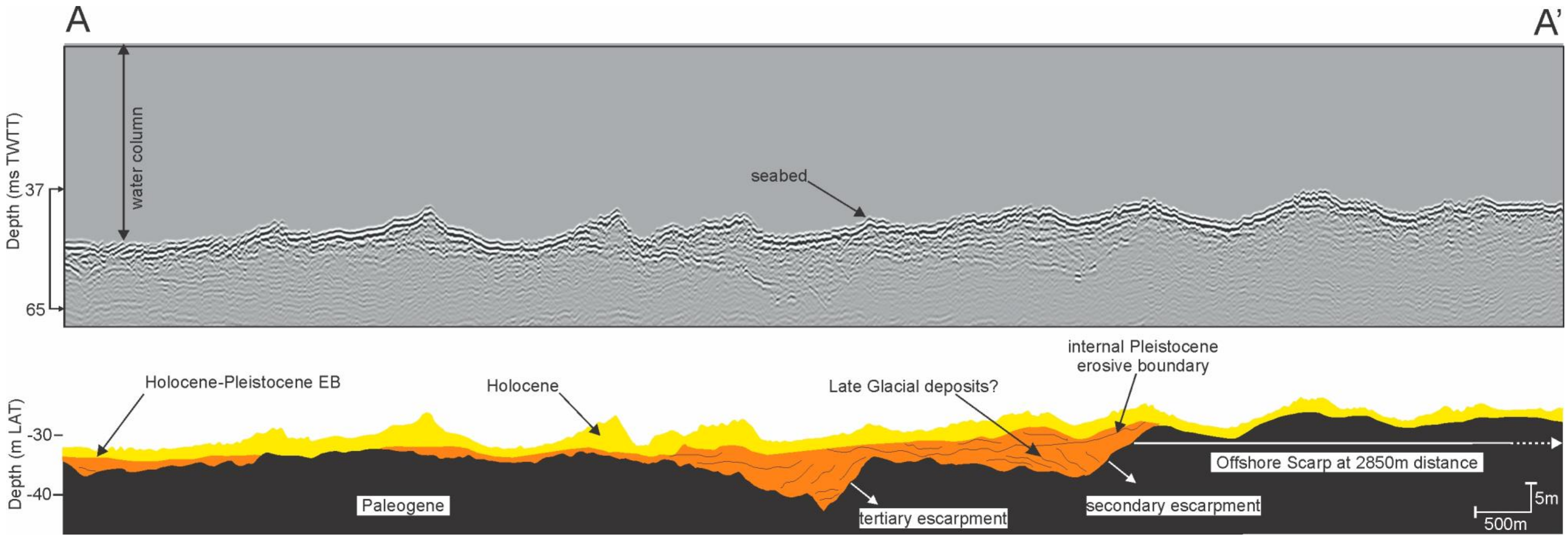
The Zeebrugge Valley (Figure 4.2) is a smaller outflow of the ancient Scheldt pathway (38 km^2, compared to 280 km^2 of the Ostend Valley) and is mapped for the first time. The valley separates the Central and Eastern Marginal Platform. Similar to the Ostend Valley, its eastern slope is much steeper (14.5%) than its western slope (> 1%), which can be directly related to the presence of the very hard and compact erosion-resistant Bartonian clay (Jacobs et al., 2002). The deepest point in this valley is located in a centralised depression called the Zeebrugge Pit (Liu, 1990), reaching up to 34 m below LAT.

The last palaeovalley that is buried beneath the present coastline belongs to the Yser Valley (Figure 4.2). The Yser River is still active today within the buried valley system and acts as an independent drainage system within the West Flanders (Baeteman, 1999). The Western Marginal Platform separates it from the Scheldt drainage system. Seismic evidence offshore of the Yser Valley is at best difficult to interpret (biogenic gas combined with shallow geology, and low water depths creating a strong multiple effect), but the general contours of the valley can be distinguished. It is clear that the Western Marginal Platform blocks the pathway north of the Yser palaeovalley, diverting it westwards into the French part of the North Sea (i.e. the Western Valley, Figure 4.2 after Liu, 1990). The deepest parts of this shallow palaeovalley locally reach 30 m below LAT, but generally range between 17 and 25 m below LAT.

Offshore, two more palaeovalleys occur near the Belgian-Dutch border, the Thornton and Northern Valley (Figure 4.2). The Thornton Valley is a small and narrow (0.5–1 km wide) incision that cuts through the Middle Scarp and up to 12 m deep through a small high in the Top-Paleogene surface below the Thornton sandbank. The maximum depth of this valley is 41 m below LAT whilst its flanks reach up to 30 m below LAT. The slopes of the valley vary between 2% and 6%. Beyond the Middle Scarp this valley opens up to a 6 km wide open low-lying plain, and bifurcates in two 2.5 km wide valleys with their deepest points ca. 11 m below the surrounding planation surface (see Figure 4.2). The crosscutting relationship of the Thornton Valley with the Middle Scarp indicates that the latter is older. This has important implications for future palaeogeographical reconstructions of the area (e.g. Busschers et al., 2005, 2007, 2008; Hijma et al., 2012; Peeters et al., 2015; Rijsdijk et al., 2013). Farther north the Thornton Valley connects to the Northern Valley (see Figure 4.2). This palaeovalley is very shallow and ranges between 30 and 34 m below LAT, with a maximum of locally 38 m below LAT in a depression to the north. The width of the valley varies between 0.8 km just off the Thornton Valley and 3 km farther north where it connects to the southeastern elongated depression of the Offshore Platform. The slopes of the valley reach up to 1.5% in the south, but are generally less than 1% further to the north. The northern part of this river valley is marked by a small cuesta with a length of 1–1.5 km and a local drop of 3–3.5m making an angle of 135° with the axis of the valley. At its northern edge the valley seems to be cut off by a swale in-between the present-day Hinder sandbanks indicating that erosion took place at a later phase.

Recent seismic data revealed a narrow elongated buried valley on the western border of the BCS that enters the FCS ca. 20 km offshore the present coastline (Liu, 1990; Figure 4.2). The valley runs semi-parallel to the Belgian-French border and widens from 0.8 km to ca. 3 km over a distance of 13.6 km. The depth of the valley is variable and changes between 8.2 and 13.2 m below the surrounding planation surface.

Figure 4.3 (next page) Seismic reflection profile shot with a GSO 360 tips Sparker (top) and interpreted panel (below) illustrating seismic facies character and association for deposits overlying the Top-Paleogene surface. The location of profile A-A’ is shown in Figure 4.2.


A
A'
Depth (ms TWTT)
37
65
water column
seabed
Holocene-Pleistocene EB
Holocene
Late Glacial deposits?
internal Pleistocene erosive boundary
Depth (m LAT)
-30
-40
Paleogene
tertiary escarpment
secondary escarpment
Offshore Scarp at 2850m distance
5m
500m

The V-shaped valley in the south changes abruptly (after only 1.4 km) to a flat-floored valley towards the north. In the north, beneath the Fairy Bank, the valley shallows and seems to terminate just north of the Offshore Scarp. The origin and nature of the deepening of the Top-Paleogene in the northwestern part of the BCS is until now unclear, but is likely related to the proximity of the Axial Channel (see Liu, 1990). This 'new' western valley appears to cut off the Offshore Scarp at the western edge of the BCS beneath the Fairy Bank where, according to Mathys (2009), it is supposed to connect with the Quaternary Basin in French waters. It suggests that the palaeovalley is younger than the Offshore Scarp and the Quaternary Basin. In analogy with the Thornton Valley, located beneath the Thornton sandbank, this valley was named the Fairy Valley. The south-north orientation of the valley is parallel to the lithological variability of the underlying Paleogene layers possibly indicating a relationship. The orientation of the valley seems to differ somewhat from the other valleys on the BCS. No explanation for this orientation can be given yet. A larger scale view beyond the study area might provide more information on this detail.

4.5 Conclusions

A digital high-resolution DCSM of the Top-Paleogene surface of the BCS and neighbouring coastal zone has been constructed based on extensive data comprising offshore boreholes and seismic records, together with onshore boreholes and cone penetration tests. The mapping allowed quantifying the depth and extent of existing and new morphological features in detail, which is important for assessing the geoarchaeological potential of the area, as well as to provide an accurate base level of the Quaternary deposits, being a resource with economic viability. The relative heights of a series of drowned river valleys, platforms, slopes and scarps were better defined, and for the first time, a smaller valley below the harbour of Zeebrugge was mapped, as well as a valley near the Belgian-French border. Linking the relative height of the Quaternary deposits on top of these geomorphological features to sea level changes will shed new light on the timing of their formation. In the near future, the DCSM will be merged with data from the Netherlands to re-evaluate the evolution of the Rhine and Meuse rivers on the BCS, two rivers that controlled sedimentation and erosion processes in the southern North Sea area between Belgium-East Anglia and the Dover Strait (Hijma et al., 2012).

4.6 Acknowledgements

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4.7 Post-scriptum

This chapter presents the latest version of the BCS pre-Quaternary surface (e.g. Top-Paleogene) series that builds upon the results of Liu (1990), Liu et al. (1992, 1993), Mathys (2009) and Mostaert et al. (1989) and the state of knowledge at that time of writing (e.g. end of 2015). In the period 2016–2017 additional geophysical campaigns were conducted to improve our knowledge of this surface across the BCS and into the neighbouring shelf areas. This resulted in a continuous updating cycle of the surface as presented in this chapter. The final results of this updating process are presented in Chapters 5 and 6. It is important to note that each updated version of the pre-Quaternary surface map resulted in new insights of the observed geomorphological features and the processes responsible in their formation (f.e. it is believed that the Northern Valley may not be a actual valley as its morphological expression diminishes with a closer spaced dataset; it has also become clear that incised-valleys are present offshore the Yser Valley but they are difficult to track because of strong multiples and subsurface gas).

From here on the BCS will be divided into three areas: the Inner, Middle and Outer BCS. The Inner BCS stretches from the present coastline to the Middle Scarp, while the Middle BCS is bounded between the Middle and Offshore Scarps. Finally, the Offshore Scarp and the BCS outline bound the Outer BCS to the north. These names were chosen to provide a better geographic orientation of the different geomorphological sections of the BCS and are partly based on the east-west (Inner BCS) and north-south (Outer BCS) orientations of observed geomorphological features.

The pre-Quaternary surface is used in the next chapters to develop the geological evolution of the BCS for the Late Saalian to the LGM (MIS 6–2) and thus forms the foundation for this dissertation. In Chapter 5 we first discuss the evolution of the palaeo-Scheldt Valley in an area that covers the eastern coastal plain to the Middle BCS, while in Chapter 6 we discuss the evolution of the Middle-Outer BCS. This order of business was chosen because the Outer BCS data density was still too low to start a detailed study and additional surveys were needed (performed in 2016–2017). The Middle BCS is covered in both chapters because the area is influenced by geological processes originating from the land interior (east-west orientation) but also from the southern North Sea (north-south orientation).

5 A WELL-PRESERVED EEMIAN INCISED-VALLEY FILL IN THE SOUTHERN NORTH SEA BASIN, BELGIAN CONTINENTAL SHELF-COASTAL PLAIN: IMPLICATIONS FOR NORTHWEST EUROPEAN LANDSCAPE EVOLUTION

This chapter is a slightly modified version of the paper De Clercq M, Missiaen T, Wallinga J, Hurtado ZO, Versendaal A, Mathys M, De Batist M. 2018. A well-preserved Eemian incised-valley fill in the southern North Sea Basin, Belgian Continental Shelf-Coastal Plain: implications for northwest European landscape evolution. Earth Surface Processes and Landforms. DOI: 10.1002/esp.4365

Abstract

This paper demonstrates that the Belgian Continental Shelf and coastal plain occupy a key position between the depositional North Sea Basin and the erosional area of the Dover Strait as it is an area where erosional landforms and fragmented sedimentary sequences provide new evidence on northwest European landscape evolution. The study area hosts 20–30 m thick penultimate to last glacial sand-dominated sequences that are preserved within the buried palaeo-Scheldt Valley. Here, we build on the results of previous seismo- and lithostratigraphical studies, and present new evidence from biostratigraphical analysis, OSL dating and depth-converted structure maps, together revealing a complex history of deposition and landscape evolution controlled by climate change, sea-level fluctuations and glacio-isostasy. This study presents strong new supportive evidence on the development of the incised palaeo-Scheldt Valley landform that became established towards the end of the penultimate glacial period (MIS 6; Saalian) as a result of glacio-isostatic forebulge updoming, proglacial lake drainage and subsequent relaxation of a forebulge between East Anglia and Belgium following ice-sheet growth, disintegration and retreat in areas to the north. The majority of the incised-valley fill is of estuarine to shallow marine depositional context deposited during the transgression and high-stand of the last interglacial (MIS 5e: Eemian). A thin upper part of the valley fill consists of last glacial (MIS 5d–2: Weichselian) fluvial sediments that show a gradual decrease and retreat of fluvial activity to inland, upstream reaches of the valley system until finally the valley ceases to exist as the combined result of climate-driven aeolian activity and possibly also glacio-isostatic adjustment. Thus, strong contrasts exist between the palaeo-Scheldt Valley and estuary systems of the Penultimate Glacial Maximum to Last Interglacial (Saalian, Eemian), the beginning of the Last Glacial (Weichselian Early Glacial and Early-Middle Pleniglacial), and the Last Glacial Maximum to Holocene.

5.1 Introduction

The BCS is a key area in the southern North Sea that connects the North Sea shelf with the English Channel shelf (Figure 5.1 A) and is featured in part-evidence, part inference-based studies on Middle-Late Pleistocene 'proglacial drainage' and 'catastrophic land bridge removal' theories (e.g. Gibbard, 1995; Gupta et al., 2007, 2017; Hijma et al., 2012; Cohen et al., 2017; García-Moreno, 2017). During periods of maximum glaciation proglacial lakes are envisaged in the southern North Sea, blocked by an ice-sheet to the north and upland topography (i.e. a land bridge) between continental Europe and England to the south (Gibbard, 1995; Toucanne et al., 2009; Hijma et al., 2012; Murton and Murton, 2012). The waters collected in these lakes were routed southward to the English Channel shelf as lake overspill. This carving of a proglacial lake spillway valley is used to explain the formation of the Dover Strait (Roep et al., 1975; Smith, 1985). In the aftermath of the final land bridge removal a proglacial valley in the southern North Sea, partially located on the BCS, remained in use by the Meuse River (Busschers et al., 2008). Some of these theories explicitly point to the Belgian palaeo-Scheldt-related palaeovalley records in their stories on the erosion history of the BCS and her surroundings (Gibbard, 1995; Hijma et al., 2012). However, evidence from the infills of the palaeo-Scheldt Valley, as presented in this study, is not used to its full potential to constrain such theories and support or disprove them.

The BCS is regarded as a sediment-depleted shallow marine environment (Mathys, 2009) characterised by a modest non-linear, glacio- and/or hydro-isostatic subsidence component (Kiden et al., 2002; Vink et al., 2007) resulting in limited accommodation space during multiple glacial-interglacial cycles. This relatively small accommodation space is larger than the available sediment supply, repeatedly reworking older sediments outside incised-valley systems before final burial takes place (Mathys, 2009). As a result the average thickness of the Quaternary sediments outside valleys and sandbanks is often 10 m or less (Balson et al., 1992; Deleu and Van Lancker, 2007; Le Bot et al., 2003; Liu, 1990; Mathys, 2009). However, the eastern coastal plain of Belgium (Figure 5.1 B) is characterised by a ca. 20–30 m thick accumulation of Pleistocene sediments, which occurs as the infill of a buried palaeovalley (De Breuck et al., 1969; De Moor and Heyse, 1974; Depret, 1981; Devos, 1984; Heyse, 1979). This palaeovalley extends further onto the BCS and represents the downstream part of the palaeo-Scheldt Valley, which flowed into the southern North Sea Basin through a series of valley segments, known as the Flemish, Coastal and Ostend Valleys (Liu, 1990; Liu et al., 1992; Mathys, 2009; Mostaert et al., 1989).

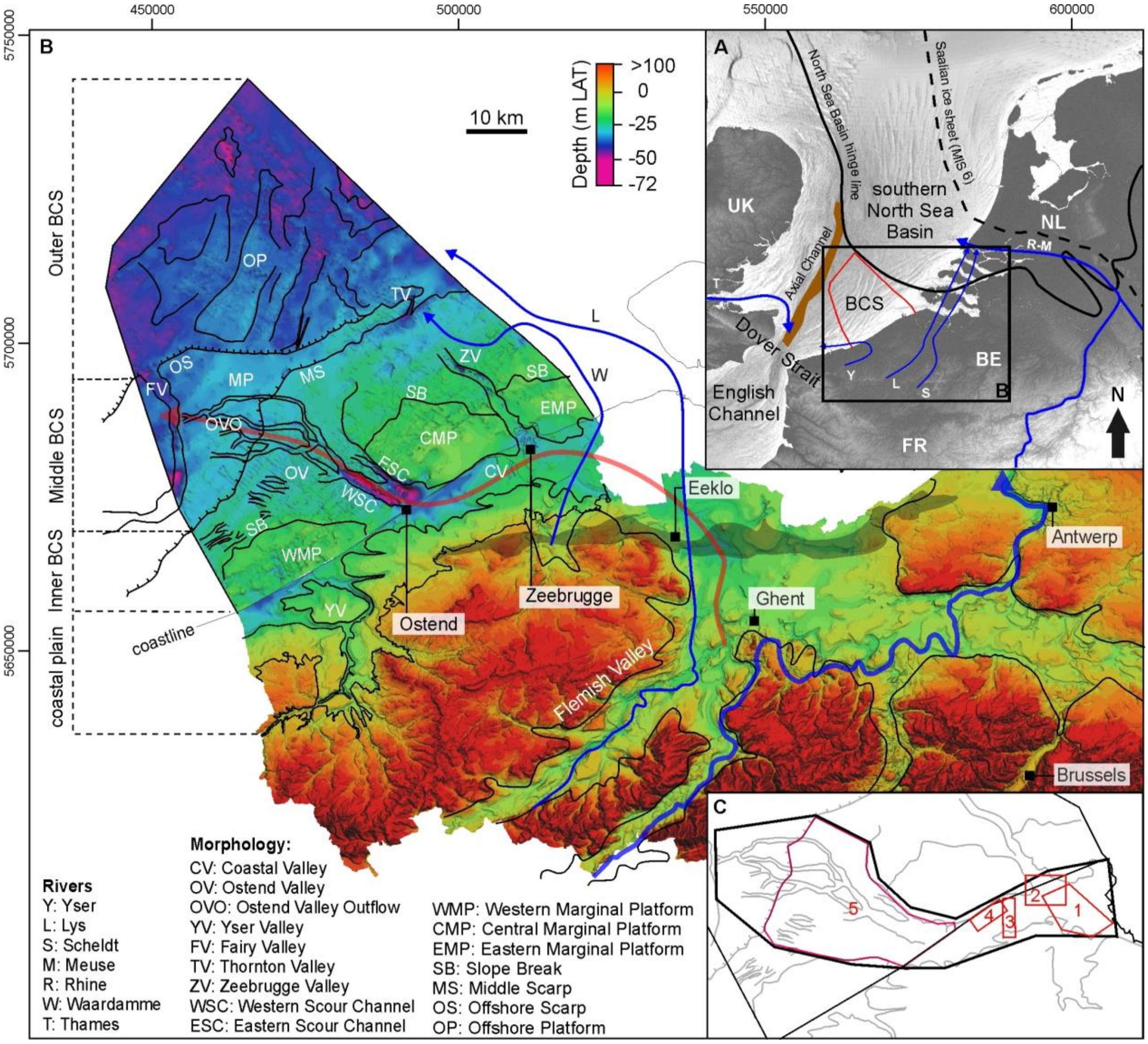


Figure 5.1 A) Location of the larger study area in relation to the (approximate) position of the North Sea Basin hinge line (Knox et al., 2010) and the maximum extent of the Late Saalian (MIS 6) ice-sheet (Ehlers et al., 2011). Also the direction of the Rhine and Meuse (Busschers et al., 2008) and several Belgian rivers (De Moor and Pissart, 1992; Kiden and Verbruggen, 2001) is given up to the Late Saalian Drenthe deglaciation (ca. 160 ka: Busschers et al., 2008). The Dover Strait course of the Yser River is based on Liu (1990) and Bogemans et al. (2016). B) The present-day Scheldt River (thick blue line) and palaeo-Scheldt River (red line) projected on top of the pre-Quaternary surface alongside its morphological elements (black lines) (modified from De Clercq et al., 2016). In translucent black the extent of the Late Pleniglacial Maldegem-Stekene sandridge is given (modified from Derese et al., 2010). C) The study area (black line) and previous local investigations areas (red boxes 1 to 5).

The sedimentary infill of the palaeo-Scheldt Valley so far has not been used to full potential in the context of the Pleistocene evolution of the southern North Sea Basin. The few studies in the eastern coastal plain were mostly local investigations for geotechnical or hydrogeological purposes (e.g. De Breuck et al., 1969; De Moor and Heyse, 1974; Depret, 1981; Devos, 1984: Figure 5.1 C), resulting in the definition of local lithostratigraphic units that later turned out to be often incoherent (Table 5.1). Moreover, during these early investigations, it was not known that a buried palaeovalley existed at that location. The recognition of the palaeovalley followed from morphological analysis of the offshore pre-Quaternary surface (Liu, 1990; Mostaert et al., 1989). Indeed, the discovery of a deeply scoured (ca. 35–50 m) buried palaeovalley extending from the present-day coastline in a southeast-northwest seaward direction, i.e. the Ostend Valley, required the presence of an onshore extension, i.e. the Coastal Valley (Liu, 1990; Mostaert et al., 1989; Figure 5.1 B). The offshore deposits of the Ostend Valley were studied in great detail by Mathys (2009).

Table 5.1 Overview of the terminology used in the literature to describe the Pleistocene deposits of the palaeo-Scheldt Valley in northwest Belgium (expressed as formations; see also ncs.naturalsciences.be/quaternary). In between brackets the depositional environment is indicated: a: aeolian; e; estuarine; i: intertidal; f: fluvial; m: marine. The locations of areas 1 to 5 are indicated on Figure 1 C.

Series	**Stage**	**Substage**	**MIS**	Area 5 Mathys (2009)	**Area 4** Devos (1984)	**Area 3** De Breuck et al. (1969)	**Area 2** Depret (1981)	**Area 1** De Moor and Heyse (1974); Heyse (1979; Hijma et al. (2012)
Late Pleistocene	Weichselian	Late Glacial	2					Maldegem (a)
		Pleniglacial	4–2		Eeklo (f) Damme (f)	Zuienkerke (i-m)	Eeklo (f)	Eeklo (f)
		Early Glacial	5d–a		Zeebrugge	Houthave (i) Wenduine (f)	Damme (f) Zeebrugge (f)	Damme (f)
	Eemian		5e	U3 (e-m) U2 (e) U1 (f-e)	K2 (?) K1 (?)	Meetkerke (i) Uitkerke (a-f) Oostende (m)	Moerkerk e(m) Kaprijke(e)	Meetkerke (i) Moerkerke (m) Kaprijke (e) Adegem? (f)
	Saalian		6					Adegem (f)

Current understanding (in part hypothesis) from previous studies, in partial analogy to the Yser Valley system to the south (Bogemans et al., 2016) and the Thames-Medway system across the North Sea (Bridgland and D'Olier, 1995; Gibbard, 1995), is that the palaeo-Scheldt Valley connects to a larger incised trunk valley, the Axial Channel, between Belgium and England (Hijma et al., 2012; Mathys, 2009). This connection was first established at the end of the Saalian glaciation (MIS 6). In this period a proglacial lake was formed between a merged ice-sheet across the North Sea Basin and a land bridge between England and southwest Netherlands-Belgium (García-Moreno, 2017; Gibbard, 1995; Hijma et al., 2012; Mathys, 2009). Large rivers, such as the Rhine and Meuse, alongside other smaller rivers, such as the palaeo-Scheldt, entered this proglacial lake. The rise in lake level initiated overspill and valley incision at the land bridge and eventually caused it to breach (Cohen et al., 2014; Gibbard and Cohen, 2015; Hijma et al., 2012). According to some authors the fall in base-level initiated the incision of the palaeo-Scheldt Valley that can be traced up to the inland Flemish Valley (Heyse and Demoulin, 2018; Mathys, 2009). Similar upstream retrogressive erosion is also observed for the Meuse River in the Netherlands (Busschers et al., 2008; Hijma et al., 2012; Peeters et al., 2016).

In and around the North Sea Basin, the proglacially carved valleys were transgressed during the Eemian interglacial, and this period left considerable depositional evidence in the form of incised-valley fills. The subsequent infilling history of the palaeo-Scheldt Valley was the result of a single marine incursion during the Eemian interglacial (MIS 5e) via the Ostend Valley followed by minor incision and deposition during the following sea-level drop (Mathys, 2009). However, prior to the lake drainage, the palaeo-Scheldt River conjointly with other Belgian rivers, flowed in a southwest-northeast direction towards the Netherlands in the direction of the Saalian ice-sheet (De Moor and Pissart, 1992; Hijma et al., 2012) where they may have entered a proglacial lake (Figure 5.1 A). One can question if proglacial trunk valley cutting adequately explains a drastic reorganisation of the Belgian river systems as suggested by Mathys (2009) and Heyse and Demoulin (2018). Other authors, such as Hijma et al. (2012), suggest that the palaeo-Scheldt River already flowed southwest in the direction of the Dover Strait during and presumably also after the Elsterian glaciation (MIS 12). With this they infer that a drainage divide connected East Anglia and the southwest Netherlands/Belgian Campine area, with the palaeo-Scheldt system passing through a saddle region of the divide.

During the Eemian transgression the palaeo-Scheldt Valley infill is believed to be the result of a rapid relative sea-level rise that includes a sudden seal-level acceleration of ca. 4 m culminating into a rapid and almost instantaneous drowning of the palaeo-Scheldt Valley (Mathys, 2009). Most of the Eemian records studied around the North Sea cover only the last bit of transgression because of limited depth of valley fill at inland positions. Such studies cannot confirm nor exclude acceleration (Long et al., 2015). The nearby onshore Yser Valley, located southwest of the palaeo-Scheldt Valley, shows a normal transition from fluvial to estuarine sediments without indication of sea-level acceleration (Bogemans et al., 2016). The palaeo-Scheldt Valley is unique because it is a deeply incised-valley located both half onshore-half offshore. This way the palaeo-Scheldt Valley should be able to preserve evidence of any possible acceleration. The final result, following the 4 m acceleration model of Mathys (2009), is a more than 20 m thick sedimentary sequence of marine deposits in the Coastal Valley (e.g.

De Breuck et al., 1969; De Moor and Heyse, 1974; Depret, 1981; Devos, 1984; Heyse, 1979). This concept of a sudden drowning of the palaeo-Scheldt Valley, however, contradicts data interpretations from other buried palaeovalleys connecting to the palaeo-Scheldt Valley (e.g. Zeebrugge Valley: Mathys, 2009; Dutch Channels: Ebbing and Laban, 1996) and the very complex sedimentary architecture throughout the Coastal Valley (e.g. De Breuck et al., 1969; De Moor and Heyse, 1974; Heyse, 1979; Depret, 1981; Devos, 1984) as well as observations made in the southern North Sea (Long et al., 2015).

Arguably the following Weichselian sea-level lowering combined with climate deterioration promoted intense deep fluvial incision up to 20 m deep, in the inland Flemish Valley (Heyse and Demoulin, 2018), whereas the downstream palaeo-Scheldt Valley experiences ‘limited’ incisions of 15 m deep (De Moor and Heyse, 1974; Mathys, 2009). Lastly, the lower reach of the palaeo-Scheldt Valley was gradually abandoned in favour of a new course via Antwerp and is associated with major periglacial aeolian activity (Maldegem-Stekene sandridge; Figure 5.1 B) and decreased fluvial discharge (Crombé and Robinson, 2017; Derese et al., 2010; Heyse, 1979; Heyse and Demoulin, 2018). However, geophysical modelling results of forebulge warping show that minor elevation differences of several meters between Belgium and the Netherlands are established following glacio-isostatic upwarping across the northern Netherlands (Peltier, 2004; Steffen, 2006). This suggests that glacio-isostasy cannot be ruled out as a causal element of the final abandonment of the palaeo-Scheldt Valley west of Ghent.

In this study, a combination of a recent high-resolution DCSM of the pre-Quaternary surface of the BCS (De Clercq et al., 2016) and (onshore) Flanders (Matthijs et al., 2013) with a newly compiled very-high-resolution bathymetry map of the BCS is supplemented by old and new seismic reflection data and archived sediment core descriptions. This has provided us with the opportunity to make a comprehensive sedimentological and morphological study that has led to new insights on a valley-wide scale and allowed us to extend the course of the palaeo-Scheldt River even further into the southern North Sea Basin. Our sedimentological interpretations are supported by new pollen and malacological analysis. Chronological control is supplied by OSL dating of a sediment core located at the transition between the Coastal and Ostend Valley, almost penetrating the deepest parts of the sedimentary infill (sediment core GR1: Figure 5.2 B). These results are combined into a palaeogeographic map series of the BCS from the Late Saalian to the Late Pleniglacial with specific focus on the Eemian interglacial sea-level rise.

5.2 Geomorphological-stratigraphical setting

The palaeo-Scheldt Valley extends from northwest Belgium, over the Belgian coastal plain onto the BCS (Figure 5.1 A). It is located south of the North Sea Basin hinge line (Knox et al., 2010), a setting that resulted in a long-term stable pre-Quaternary surface relatively little affected by tectonic subsidence or uplift (D'Olier, 1981; Funnell, 1996; Kiden et al., 2002; Vink et al., 2007). Accommodation space to deposit transgressive sediments on the relatively flat pre-Quaternary surface was limited prior to the Saalian proglacial lake spillway erosion followed by the incision of numerous valleys across the BCS (e.g. the formation of the Offshore Scarp by the Meuse River and incised-valleys such as the palaeo-Scheldt, Zeebrugge and Thornton). During the following Eemian transgression the inherited Saalian palaeotopography provided sufficient accommodation space, especially in incised-valleys, to deposit transgressive and regressive sediments up to the eustatic high-stand position. This inherited setting resulted in the formation of important erosional and depositional features: erosional features such as tidal channels and scour hollows in incised-valleys (Liu et al., 1993; Mathys, 2009; Mostaert et al., 1989), and depositional features such as intercalated coastal plain and near-coastal marine deposits as has been observed in the Yser Valley (Bogemans et al., 2016).

A series of morphological features of the pre-Quaternary surface on the BCS have been mapped and described in literature (Liu, 1990; Liu et al., 1992, 1993; Mostaert et al., 1989; Mathys, 2009). Recently a new high-resolution DCSM has been constructed for the BCS and connected to the coastal plain (modified from De Clercq et al., 2016: Figure 5.1 B). This high-resolution map not only resulted in a more in-depth description for the already described morphological features but also for several newly discovered ones (De Clercq et al., 2016).

The pre-Quaternary surface can be morphologically subdivided into three major planation surfaces (Marginal, Middle and Offshore Platforms), each bounded by escarpments or slope breaks that cut across different Paleogene strata of varying resistance (De Batist, 1989; De Clercq et al., 2016; Liu, 1990; Mathys, 2009) that were formed under marine to near-coastal conditions (Boulvain and Vandenberghe, 2018; Gibbard and Lewin, 2016). This setting resulted in a lateral changing lithology composed of consolidated clays and sand-silt-clay layers locally containing sandstone layers (De Batist, 1989; Jacobs and De Batist, 1996; Le Bot et al., 2003). Just offshore the modern shoreline, the Marginal Platform divides into three segments (Western, Central and Eastern Marginal Platform) separated by the palaeo-Scheldt River (Ostend and Zeebrugge Valleys), while in the western coastal plain the isolated drainage system of the Yser Valley is situated (Bogemans et al., 2016). The palaeo-Scheldt Valley connects to the inland Flemish Valley and extends into the Coastal Valley, parallel with the present-day coastline, and turns west to enter the BCS at Ostend. There it opens up into the funnel-shaped Ostend Valley. The Ostend Valley and Zeebrugge Valley can be traced over considerable distance across the Inner BCS. The continuation of the Zeebrugge Valley is unclear further downstream but appears to be oriented in the direction of the Thornton Valley, while the Ostend Valley can be traced down to the landward edge of the Middle BCS (Middle Scarp: Liu et al., 1992). In an east-west direction the Middle BCS commences with a high in the pre-Quaternary surface southwest of the Thornton Valley. In a westward direction it widens to a ca. 20 km wide plateau almost devoid of any structures. Near the Belgian-French border a partially infilled palaeovalley is present (Fairy Valley: De Clercq et al., 2016) that cuts through the Offshore Scarp, the boundary between the Middle-Outer BCS (cf. Liu et al., 1992). The Outer BCS (i.e. Offshore Platform) is characterised by north-northwest to south-southwest oriented elongated palaeodepressions located in between ridges.

The study area comprises the combined Coastal-Ostend Valley and its downstream trajectory (Figure 5.1 B), which will from here on be referred to as the palaeo-Scheldt Valley. The sedimentary infill of this valley system shows a stacked sequence of dominantly sandy sediments deposited under different climate conditions (De Breuck et al., 1969; De Moor and Heyse, 1974; Depret, 1981; Devos, 1984; Heyse, 1979; Mathys, 2009). The average thickness of the deposits in the palaeo-Scheldt Valley is ca. 20–30 m, with local much thicker accumulations in deeply scoured pits in the Ostend Valley (cf. Sepia Pits: Liu et al., 1993), up to more than 50 m. In the downstream part of the palaeovalley the sediments completely disappear at the Middle Scarp (Mathys, 2009).

Table 5.1 summarises the results of previous investigations carried out in the palaeo-Scheldt Valley. Upstream, in the eastern coastal plain, the palaeo-Scheldt Valley fill consists of a relative complete sequence of Saalian and Weichselian fluvial sediments separated by Eemian fluvial-estuarine-marine deposits in between (Figure 5.1 C area 1). Further downstream, this sequence deteriorates with the loss of Saalian sediments and an increasing thickness of Eemian deposits dominating the valley fill (Figure 5.1 C areas 2–4). These Eemian deposits largely have a marine origin and interfinger with intertidal sediments at the edges of the valley or fluvial sediments upstream. On top, a thin, sometimes fragmented sequence of Weichselian fluvial sediments remains present. Offshore, on the BCS (Figure 5.1 C area 5), the Weichselian sediments have been removed, possibly during the Holocene transgression, leaving ca. 30 m thick sequence of Eemian estuarine and marine valley fill (Mathys, 2009).

5.3 Data and methods

5.3.1 Seismic reflection dataset

Over the past 40 years the Renard Centre of Marine Geology (Ghent University) collected a large amount of very-high-resolution sub-bottom 2D single-channel seismic data that covers the entire BCS (Figure 5.2 A; for a more complete view see De Clercq et al., 2016). These seismic data are collected using various acoustic sources (in the study area mainly boomer and multi-tip sparker) operating at different frequencies, that penetrate and resolve unconsolidated and consolidated sediments up to ca. 100 m below the seafloor (for a summary see Mathys, 2009 and Zurita Hurtado and Missiaen, 2016). Vertical resolution of the seismic data ranged on average between 0.3 and 0.5 m. A large part of this database consisted of analogue paper data that was converted by Mathys (2009) into a reusable digital format. De Clercq et al. (2016) expanded this database with additional high-resolution seismic reflection data and converted it into a DCSM of the pre-Quaternary surface by using a dynamic depth conversion algorithm that integrates the interpreted horizons from the seismic data (in time) and borehole information (in depth) into a grid-based layered velocity model. This map has been updated with new high-resolution seismic reflection data collected across the BCS. The same process cf. De Clercq et al. (2016) has been applied to these data of which the result is visible in Figure 5.1 B. Interpretation of the seismic reflection data in the near coastal area was often hampered by the presence of shallow gas in the subsurface (translucent orange zone in Figure 5.2 A) or by shallow water depths producing strong multiples. Characterisation and interpretation of seismic facies and stratigraphy was carried out according to the criteria of Mitchum and Vail (1977) and Mitchum et al. (1977a, 1977b, 1977c).

5.3.2 Sediment core dataset

The interpretation of the marine seismic reflection data is supported by sedimentological core descriptions (Figure 5.2 A). The comprehensive offshore core database (Kint and Van Lancker, 2016) is the result from a wide range of academic and commercial-industrial studies and consists of 440 cores and vibrocores. 396 of these cores penetrate the upper 5 m of sediments, 34 cores are between 5 m and 10 m deep while 10 cores capture over 40 m of sediments. Onshore, the lithological characteristics of the palaeo-Scheldt Valley infill is supported by sediment core descriptions collected from the Flanders subsurface database (DOV, 2014). 139 of the 750 cores penetrate the complete Quaternary sequence. Data obtained from the sediment core databases form the basis for lithostratigraphic analysis and characterisation of the sedimentary facies. A summary of all sedimentary facies is presented in Table 5.2 and their distribution in the palaeo-Scheldt River Valley is represented in Figure 5.3 and as geological cross-sections in Figure 5.4. From this dataset core GR1 (part of the RBINS geological core collections) is located in a strategic position at the Coastal-Ostend Valley transition. GR1 captures the most complete and longest infill sequence (ca. 45 m) and was therefore subjected to further analysis (pollen, malacology and OSL dating). Further offshore vibrocore VC-0035, located in the downstream area of the Ostend Valley, also captures a well-preserved sequence of the valley infill and was subjected to pollen analysis. The location of these sediment cores, and other cores mentioned in the text, is presented in Figure 5.2 B.

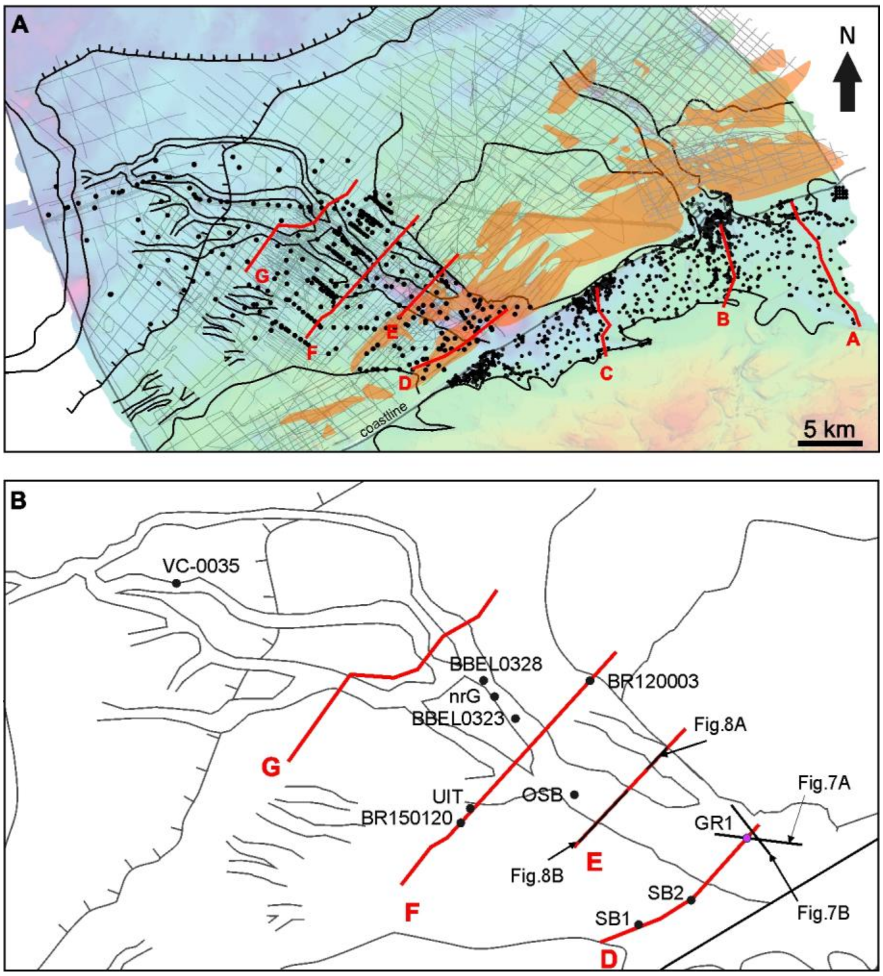


Figure 5.2 A) Overview of the seismic reflection and core data used for this study. The translucent orange area shows where shallow gas is present in the subsurface. B) Detailed view on the Ostend Valley and the Ostend Valley Outflow with indication of core names mentioned in the manuscript. Red lines indicate cross-sections imaged in Figure 5.4.

5.3.3 Malacological analysis

Six samples for malacological analysis were collected from core GR1 and analysed at TNO, the Netherlands. The samples were collected from fine-grained sandy intervals with high shell concentrations to interpret the depositional environment. For every sample sediments and organic material were removed and hydrogen peroxide was added. An overview of the results of the malacological analysis is presented in Supplementary Table 5.1.

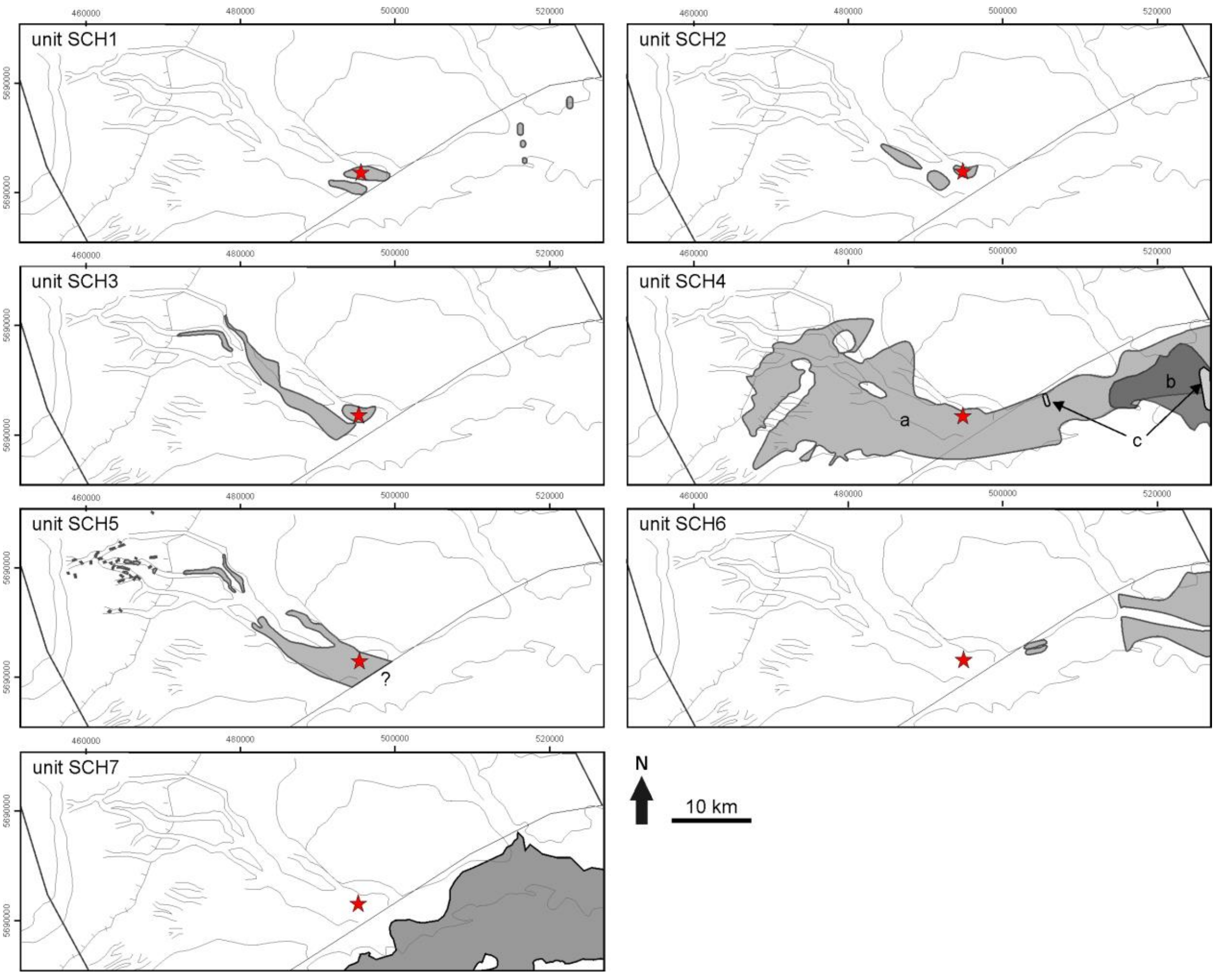


Figure 5.3 Geographical distribution of the identified units in this study. Unit SCH7 is expanded with information from Jacobs et al. (2004). The red star indicates the location of sediment core GR1.

5.3.4 Pollen analysis

In this study, pollen analysis was primarily used as a biostratigraphic correlation tool in order to relate the sedimentary record to the floating timescale for the Eemian as established by Zagwijn (1961, 1996). Pollen analysis was performed at TNO, the Netherlands. 33 samples were taken in series from silty and clayey intervals from core GR1 (Figure 5.9) and clayey and peaty intervals from vibrocore VC-0035 (Figure 5.10). 200–300 pollen grains were counted from each sample. Northwest European regional PAZs were determined based on differing taxa assemblages and known Late Pleistocene vegetation-development and responses (cf. Zagwijn, 1961, 1996).

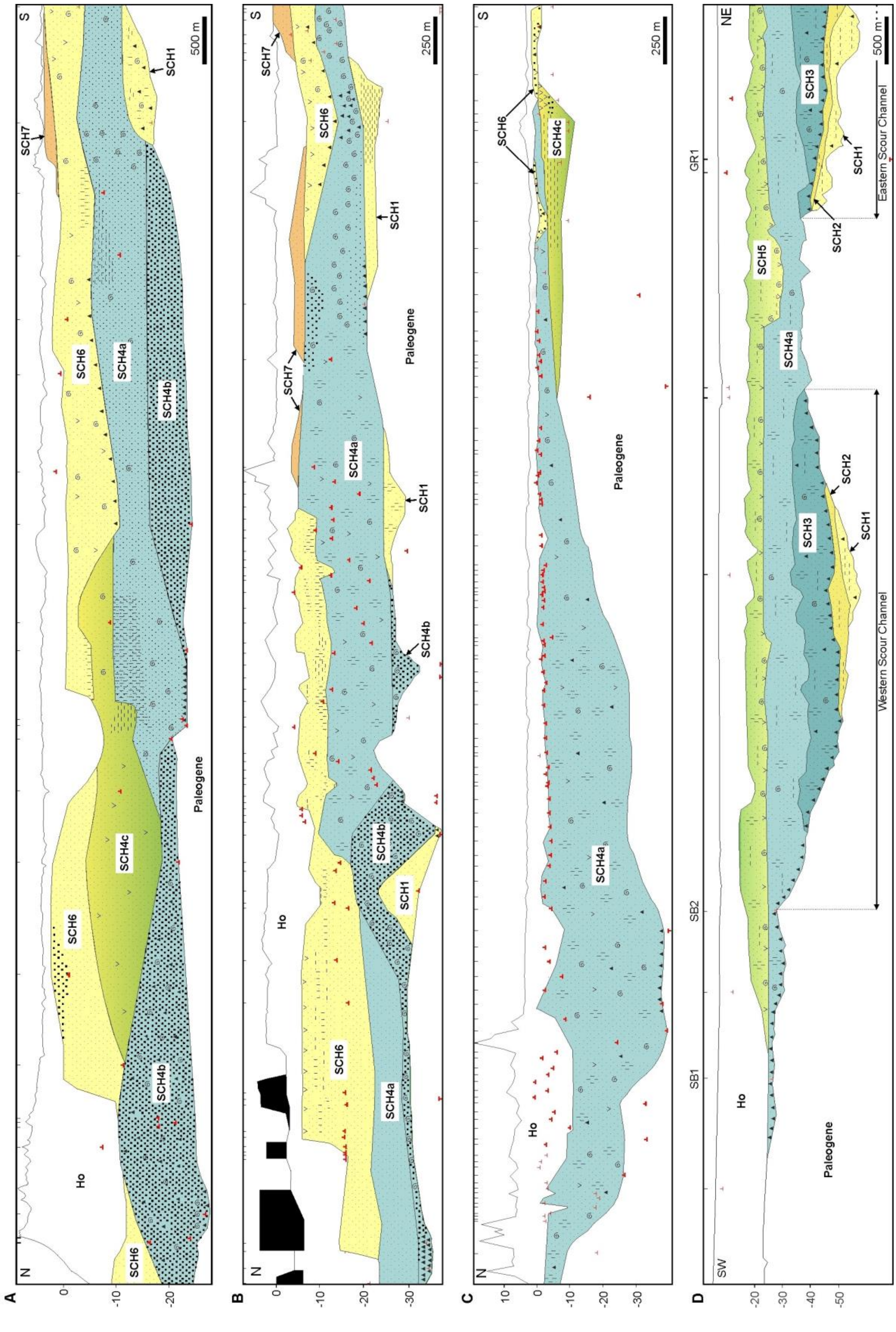


Figure 5.4 (previous page) Geological cross-sections showing the sedimentary architecture and stratigraphical position of the units identified in the palaeo-Scheldt Valley from upstream (A) to downstream (G). Cross-section locations are plotted in Figure 5.2. Horizontal and vertical scales differ between the cross-sections. The colours of the units are matched to the landscape environments in Figure 11–18. Depths are in meter LAT.

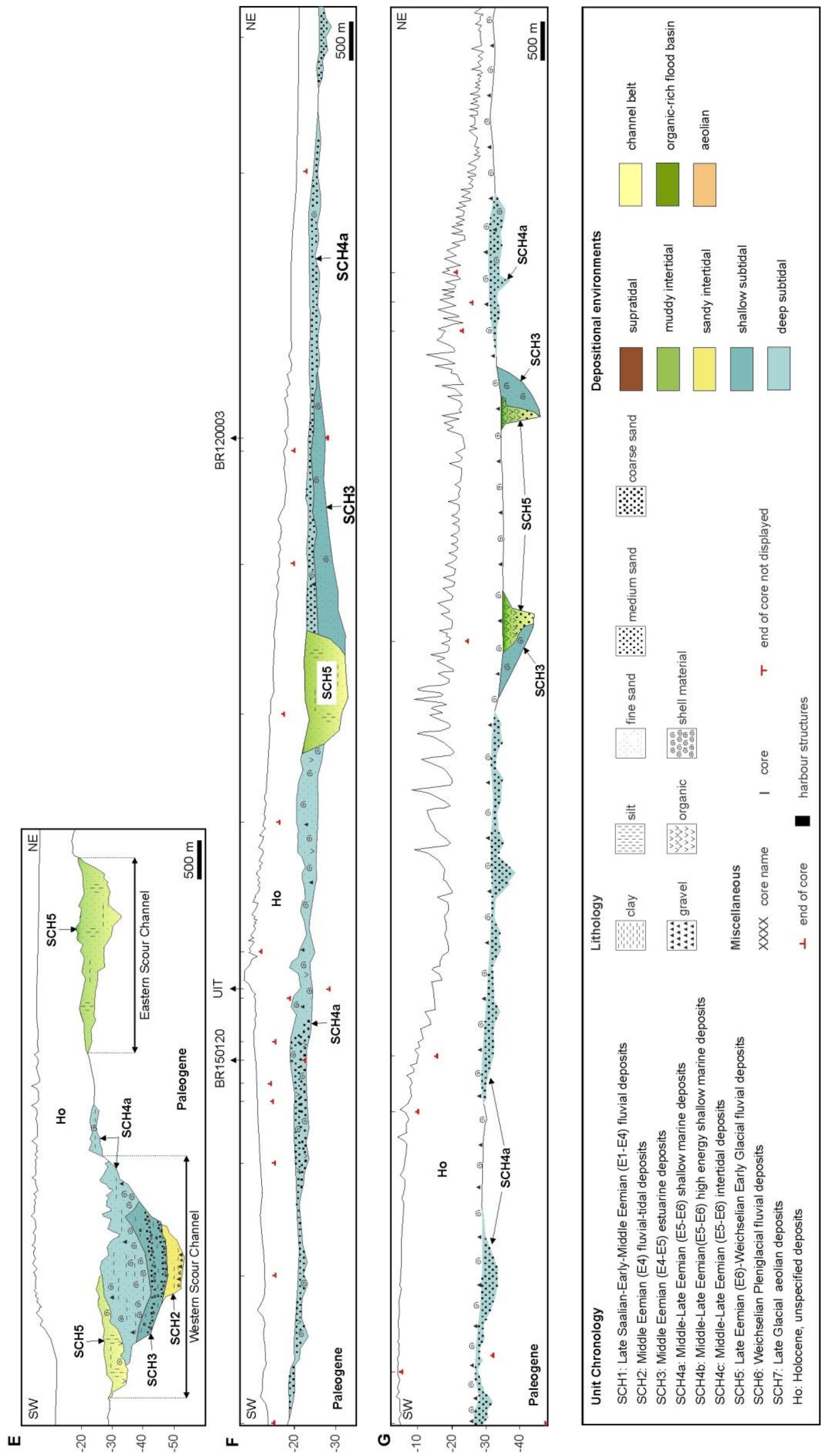


Figure 5.4 Continued.

Table 5.2 Summary of sedimentary and seismic facies of the depositional units identified in the separate sections of the palaeo-Scheldt Valley: Coastal Valley, Ostend Valley and downstream Ostend Valley (Figure 5.1).

Unit	Sedimentary facies	Seismic facies	Old Stratigraphy
Coastal Valley			
SCH1	Silty to clayey very fine- to fine-grained sands	NA	Adegem
SCH4a	Fine- to medium-grained (very silty) sands	NA	Moerkerke, K1, K2, Oostende
SCH4b	Coarse- to very coarse-grained sands	NA	Kaprijke, Oostende, Zeebrugge
SCH4c	Silty-clayey fine- to medium-grained sands and peat	NA	Meetkerke, Houthave
SCH6	Silty very fine to medium-grained sands; clay and peat; organic sandy silts	NA	Eeklo, Damme?
SCH7	Well-sorted (silty) very fine- to fine-grained sands; humic-peaty intercalations	NA	Maldegem
Ostend Valley			
SCH1	Clay-sand alternations; clayey fine-grained sands; silty clay and clayey silt	Sigmoidal to parallel oblique	U1
SCH2	Fine- to medium-grained well-sorted clayey sands	Indiscernible	U1, U2
SCH3	Poorly to well-sorted fine- to coarse-grained clayey sands	Low- to high amplitude; transparent; high frequency	U2, U3
SCH4a	Poorly-sorted very fine-grained silty-clayey sands with sand-silt-clay alternations	Continuous sub-parallel; subchannels with prograding sigmoidal to oblique infill	U3
SCH5	Well-sorted very fine to fine-grained silty sands and clay	Horizontal to slightly wavy; transparent; parallel to sub-parallel; chaotic to oblique/chaotic	Incorrectly lumped with Holocene tidal deposits in Mathys (2009)
Downstream Ostend Valley			
SCH5	Very-fine to medium-grained sands; consolidated peat and clay		NA

5.3.5 OSL dating

Burial ages of eight samples were determined using state-of-the-art quartz OSL dating methods. Samples were obtained from sediment core GR1 under subdued amber light conditions, and prepared and analysed at the Netherlands Centre for Luminescence dating (Wageningen University). Both dose rate estimation and equivalent dose estimation as well as recuperation tests were performed.

For each of the samples, the burial age is determined by dividing the palaeodose by the dose rate. Information pertaining to the accuracy of published ages is provided in Supplementary Section 5.11.1 and Figure S4.

Table 5.3 Summary quartz OSL age results for sediment core GR1. For location of samples see Figure 5.9. See Supplementary Section 5.11.1 and Figure S4 for detailed information.

NCL Code	Depth (m LAT)	Unit	Equivalent dose (Gy)	Dose rate (Gy/ka)	Age (ka)	Validity	Aliquot saturation (%)
NCL-7616111	15.70	SCH4	89.0 ± 5.1	1.01 ± 0.04	88.1 ± 5.9	Likely OK	11
NCL-7616110	18.55	SCH4	100.1 ± 4.6	0.89 ± 0.03	112.5 ± 6.6	Likely OK	21
NCL-7616109	19.84	SCH4	84.5 ± 3.6	0.95 ± 0.03	88.8 ± 4.9	Likely OK	11
NCL-7616108	24.87	SCH4	96.3 ± 5.6	1.01 ± 0.04	95.6 ± 6.6	Likely OK	21
NCL-7616107	32.60	SCH4	104.7 ± 6.0	0.97 ± 0.04	108.0 ± 7.3	Likely OK	20
NCL-7616106	35.79	SCH3	104.1 ± 7.0	0.85 ± 0.03	122.7 ± 9.4	Questionable	39
NCL-7616105	43.71	SCH2	77.6 ± 3.9	0.75 ± 0.03	103.9 ± 6.4	Likely OK	18
NCL-7616104	44.45	SCH1	67.6 ± 3.0	0.68 ± 0.03	99.3 ± 5.8	Likely OK	10

5.3.6 Palaeolandscape map series

For this study a map series of palaeolandscape scenarios was constructed for the BCS and the southern North Sea Basin ranging from the Late Saalian to the Late Pleniglacial. The existence and persistence of a land bridge connecting East Anglia with Belgium and southwest Netherlands strongly influenced the northwest European landscape evolution. Glacio-isostatic adjustments of ice-sheets located to the north of the study area may have influenced the formation of this landscape feature. Indications of glacio-isostatic uplift within the study area can be found in the behaviour of fluvial systems by deflection and valley incision, whereas glacio-isostatic subsidence can be traced in their infill architecture. The deepest incisions achieved can for example be linked to the period of maximum uplift (e.g. Busschers et al., 2008). The effects of glacio-isostasy on the palaeo-Scheldt Valley will be discussed by combining the results of the high-resolution DCSM of the pre-Quaternary surface with the overall infill architecture of the valley system. These results will then be combined with age control (pollen and OSL) to get a better grip on the possible time constraints in which these events occurred.

Within this study we specifically focus on the BCS landscape evolution during the Eemian interglacial sea-level rise and how this affects the regional-scale landscape evolution of the southern North Sea Basin. The map series are the product of projecting eustatic sea-level values on the pre-Quaternary surface (Figure 5.1 B) as this surface is believed to be the base of the Quaternary deposits at the time of these landscape changes (Mathys, 2009). In accordance with Mathys (2009) a 4 m mean tidal range is presumed analogue to the present-day mean tidal range in Ostend, not taking into account modulations within the Eemian palaeo-Scheldt estuary due to valley convergence and friction. Our template depositional model for this area follows Baeteman (2008). It considers tidal channels are in subtidal position, below mean low water, and only fall dry during spring tide. Clayey and sandy intertidal environments typically occur between high water at neap tide, and mean low water level. For this study it is assumed that: 1) sandy intertidal environments occur to a level in the middle of the intertidal environment, halfway between high water at neap tide and mean low water; 2) salt marshes develop in supratidal conditions, above high water at neap tide (3.5 m above the tidal-channel level); and 3) freshwater marshes can develop above the high-water level at spring tide (4.25 m above tidal-channel level). The different palaeolandscape models are presented in Figure 5.12 to Figure 5.19.

5.3.7 Sea-level curve

In this study we compare the Eemian sea-level rise of the palaeo-Scheldt River to that of the Rhine River in the Netherlands (see Peeters et al., 2016). We assume that the Eemian pollen stratigraphy obtained from the palaeo-Scheldt Valley is isochronous with that of the Rhine River. This way a compatible time-transgressive chronology is used between both buried incised-valley systems.

Four malacological samples and two additional diatom samples from Mathys (2009) were used to construct a sea-level curve for the Eemian sea-level rise. The malacological samples were taken from sediment core GR1 (samples M1 and M3-M5; Figure 5.9; Supplementary Table S1). The two diatom samples come from sediment core GR1 and core OSB (10 km downstream of core GR1; see Figure 5.2 B for location, Supplementary Table 5.2). All samples were taken from fine-grained sediments within the Eemian estuarine infill of the palaeo-Scheldt Valley and correlate with specific Eemian PAZs (cf. Zagwijn, 1961, 1996). During episodes of relative sea-level rise, landward migration (transgression) of the shoreface erodes older deposits and may leave a lag of reworked shell material behind (Shennan et al., 2015). The taxonomic analysis of the shell material provides information on the depositional environment, while the preservation state was used to derive information about the transportation mechanism. The shell material and diatoms can be used as sea-level indicators to estimate former sea levels when combined with chronological information from the obtained Eemian PAZs. To do this they must possess a systematic and quantifiable relationship to elevation in the tidal frame (van de Plassche, 1986). The tidal range is derived from the model of Baeteman (2008), inspired on the modern tidal ranges in the Dover Strait and Scheldt system, and is estimated to be ca. 4 m. This vertical range is used as the vertical error bar on the depositional depth of the shell material within the estuarine environment. Considerable care must be taken in using these sea-level indicators as the compaction effects of the sandy and silty sediments are unknown and may have influenced the depth of the samples. Compaction of sediments will result in underestimation of sea levels and depends on several factors, such as initial water content, lithology, particle size and shape, age and the thickness and nature of the overburden (Vink et al., 2007). As these factors vary in their dimensions from sample to sample the samples have not been corrected for depth. The results are expressed in msl to compare them with the Netherlands sea-level curve.

Sedimentary analysis combined with age control (pollen and OSL) provides a stratigraphic framework for the incision and infill history of the palaeo-Scheldt Valley system, while depositional environments and water depths (malacology) provide information on the drowning history during transgression. The palaeogeographic map series show the evolution of the landscape in relation to regional landscape changes such as river deflection and trunk valley connection. A sea-level curve shows the transgression history of the study area compared to eustatic sea levels. This allows effects of glacio-isostatic adjustment within the study area to be analysed. Additional comparison with the Netherlands sea-level curve may help in determining the location of the forebulge crest.

5.4 Results

5.4.1 Geomorphology

5.4.1.1 palaeo-Scheldt Valley

In a downstream direction the width of the palaeovalley changes from ca. 16 km to 20 km while at the Coastal-Ostend Valley transition the width is constrained to ca. 12 km. Incision depths vary greatly within the valley. Upstream in the Coastal Valley depths range between 10 and 30 m, while in the downstream part of the Ostend Valley depths are generally shallower ranging from several meters to 13 m. The shallowest channels in the Ostend Valley are cut ca. 6–9 m in the pre-Quaternary surface while the deepest structures are up to 51 m deep (Figure 5.5). The slopes of the valley are highly variable, while the average downhill slope of the valley floor is 0.01°. Slopes on the southern edge of the Coastal valley range between 0.1° and 3°. The Ostend Valley slopes range between 0.2 and 5°. The longitudinal profile of the palaeo-Scheldt Valley is characterised by several steep slopes. In the Coastal Valley an almost flat valley floor with a slope of ca. 0.01° is measured. At about half way the valley a 'steep' drop of 11 m (a slope 0.06°) is recognised that continues over a distance of ca. 7.5 km. At the transition of the Coastal-Ostend Valleys a 'steep' drop of another ca. 12 m is recognised that continues into the elongated scour features. The drop to the central ridge between the scour features is ca. 5 m to its deepest point. This slope rises again for ca. 18 m (slope of 0.2°) where the elongate scour features terminate. From that point on until the Middle Scarp an elevation difference of ca. 4 m is measured.

5.4.1.2 Elongate scour features WSC and ESC

Recent mapping of the pre-Quaternary surface (Figure 5.1 B) revealed that the features formerly known as the Sepia Pits (Liu, 1990; Liu et al., 1993; Mostaert et al., 1989) located within the Ostend Valley are in fact interconnected features composed of two parallel channel incisions: the Western and Eastern Scour Channels (WSC and ESC respectively; Figure 5.5). These channels are located just northwest of a bend in the palaeo-Scheldt Valley. The scoured features are incised in homogeneous consolidated Paleogene clays that reach up to 150 m in thickness (De Batist, 1989; Henriet and Moor, 1989).

The WSC is 12 km long and 1–2 km wide and is incised up to 30 m below the pre-Quaternary surface. Its morphology is quite irregular resembling terraces (Figure 5.5 A–F). A striking feature is the changing location of the steepest edge of the channel (between the northeast and southwest flank).

The ESC is over 9 km long and 1–3.5 km wide and is incised up to 51 m below the surrounding pre-Quaternary surface. Downstream the depth shallows to 20 m. The narrow downstream section of the ESC has a steeper northeast flank (max. 6.5°) and a gentler southwest flank often less than 1° (max. 2.5°) and is locally marked by a terrace structure (Figure 5.5 H–L). The southeast part of the ESC is unclear and has a steep northeast flank (max. 3.5°, see Figure 5 H).

Both the WSC and ESC continue downstream in channels that increasingly shallow and bifurcate. Across the Middle Scarp, on the Middle BCS, these channels are referred to as the Ostend Valley Outflow (OVO; Figure 5.6).

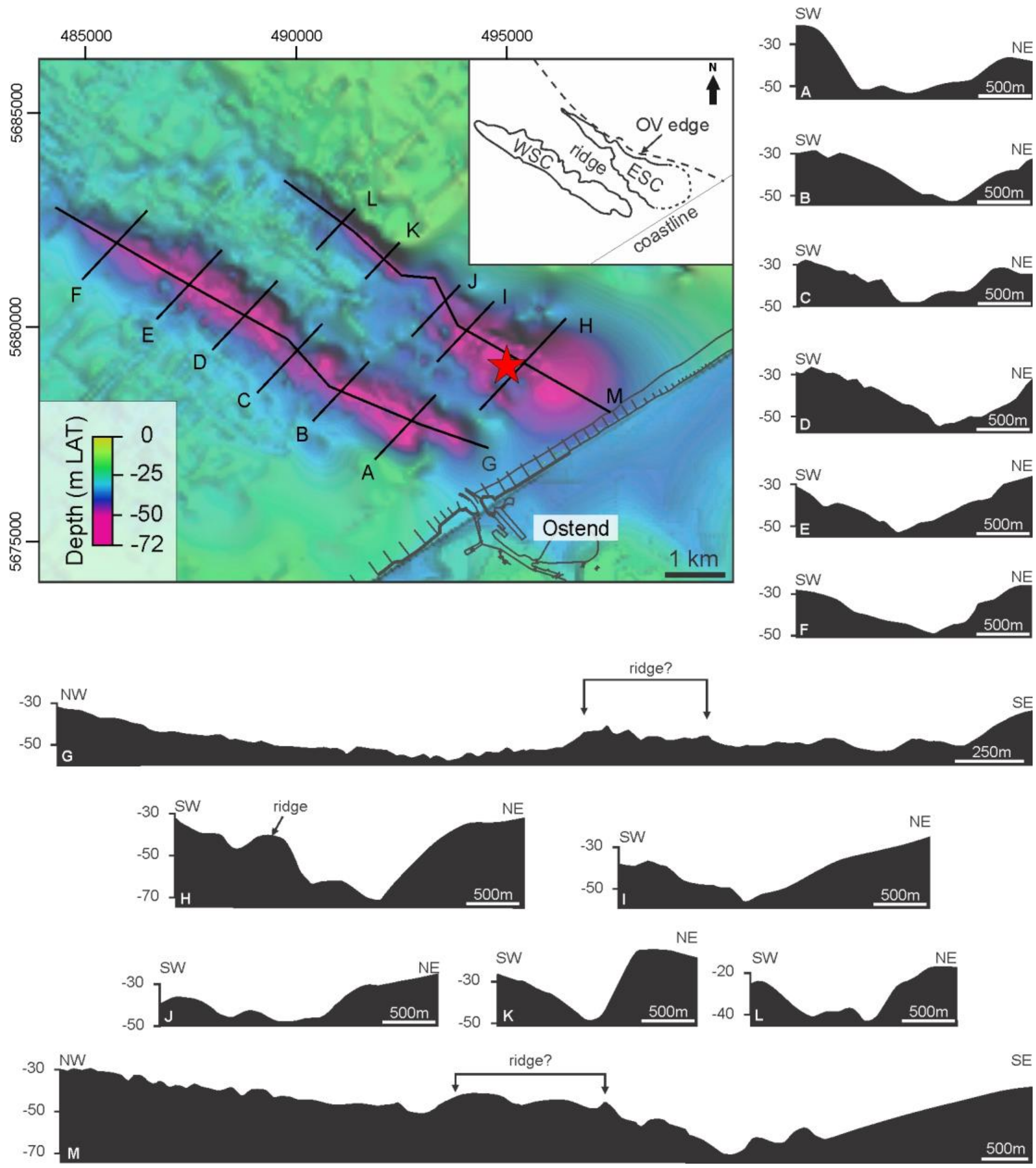


Figure 5.5 Ostend Valley scour channels (WSC and ESC) incised in the pre-Quaternary surface. The red star indicates the location of sediment core GR1. Depths are in meter LAT. Abbreviations mentioned in the figure can be found in Figure 5.1.

5.4.1.3 Downstream Ostend Valley

The downstream part of the Ostend Valley area reveals several exposed channel incisions in the high-resolution bathymetric data (Figure 5.6 A). The seismic data suggest that these incisions are part of an interconnected network of channels with a southeast-northwest orientation extending from the shallow channels observed in the Ostend Valley (Figure 5.6 B). The widths of the channels decrease in a downstream direction but generally vary from 0.1–0.6 km with a depth ranging from 1 m to locally 6 m below the pre-Quaternary surface. The longitudinal gradient of these channels show an elevation difference of 4 m over a distance of ca. 25 km (slope of < 0.01°), measured between the Middle and Offshore Scarps.

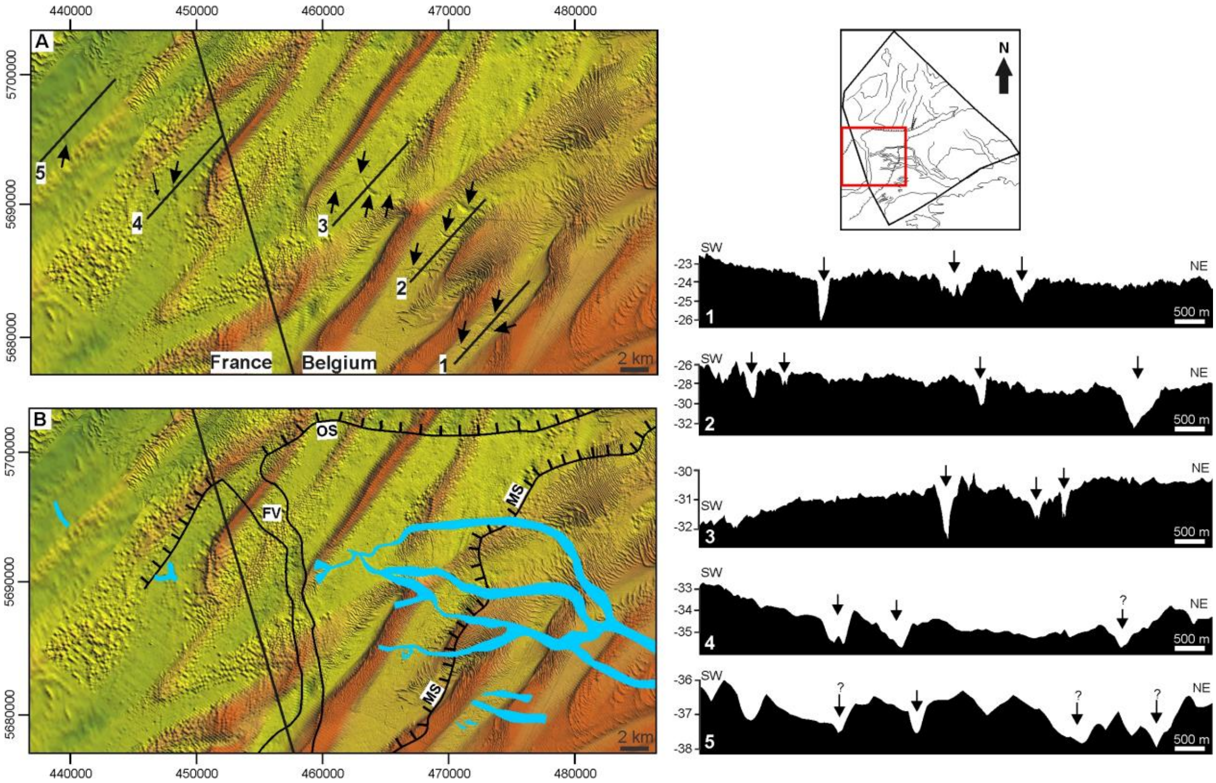


Figure 5.6 A) High-resolution bathymetry of the western BCS and French waters showing exposed channels in the sandbank swales (black arrows). Cross-sections 1 to 5 visualise these exposed channel sections. B) Same image as in A showing known channel geometry based on buried, derived from seismic reflection data, and exposed channel sections. Other morphological elements are indicated in black (see Figure 5.1). All depths are in meter LAT.

5.4.2 Valley infill stratigraphy

In the study area, a valley infill sequence of seven stacked seismic-sedimentary units was identified incised into older Paleogene deposits. The lower part of the fill consists of very fine to fine-grained fluvial deposits (unit SCH1). On top of these fluvial sediments lie three units of fine- to very coarse-grained sediments characterised by a Lusitanian fauna that are attributed an estuarine to marine origin (units SCH2 to SCH4). These interglacial deposits are truncated and buried by two fine-to medium-grained fluvial sedimentary units (units SCH5 and SCH6), which are subsequently overlain by Holocene or local well-sorted fine-grained aeolian deposits (unit SCH7). In the Ostend Valley the transition to Holocene sediments is marked by a gravel lag or coarse-grained sand, while in the Coastal Valley this transition is characterised by peat layers, incised-channels filled with sands or silt and clay in between these channels. All fluvial and estuarine deposits were attributed a palaeo-Scheldt River origin based on their tracing contacts and lithostratigraphic correlations to previous studies within the study area (De Moor and Heyse, 1974; Depret, 1981; Devos, 1984; Heyse, 1979; Mathys, 2009; see Table 5.2 for previous naming conventions that are referred to in the text). The sedimentary architecture and position of the units is illustrated in seven geological transects A–G (Figure 5.4; see Figure 5.2 for their locations). The sedimentary and seismic facies characteristics of each unit are described below together with their interpretation and are summarised in Table 5.2 that provides correlations to stratigraphic schemes used in past publications. Figure 5.3 shows the geographical distribution of all identified units.

5.4.2.1 Unit SCH1

Description – The lowermost unit SCH1 is a discontinuous deposit immediately overlying the pre-Quaternary surface and filling the deepest scour channels WSC and ESC of the palaeo-Scheldt Valley (Figure 5.5). Upstream in the Coastal Valley SCH1 consists of silty-clayey very fine- to fine-grained sand rich in calcium carbonate (Figure 5.4 A–B). Shell material and sandstone fragments occur throughout the unit and are interpreted as reworked from the underlying Paleogene strata. Downstream, in the Ostend Valley, SCH1 is present in the scour channels WSC and ESC (Figure 5.4 D–E) where it consists of sand-silt-clay alternations that change upwards into clayey fine-grained sand with low concentrations of gravel and shell material (e.g. sediment core GR1: Mathys, 2009). Seismic facies shows sigmoidal to parallel oblique reflections onlapping on the Paleogene substratum (Figure 5.7). In the Coastal Valley, units SCH4a and SCH4b truncate SCH1, while in the Ostend Valley the unit is truncated by SCH2.

Interpretation – SCH1 is interpreted as a fluvial deposit. Lateral accretion of point bars, with upward-fining cycles (Figure 5.7) indicate floodplain deposits (Miall, 2006), while pockets with a dominant sandy, clayey or silty facies upstream in the Coastal Valley suggest variable energy conditions within the fluvial environment. The presence of reworked Paleogene material is derived directly from the pre-Quaternary strata with a marine origin (De Batist, 1989; Jacobs and De Batist, 1996) certainly not possibly through erosion. This supports sedimentological and palaeoenvironmental interpretations by Heyse (1979) who interprets a subaquatic origin for this unit (also known as the periglacial Adegem Formation, Table 5.2) that largely comprises reworked marine sands of Paleogene origin. At the time of formation the Paleogene landscape must have been subject to strong erosion by water runoff (Heyse, 1979).

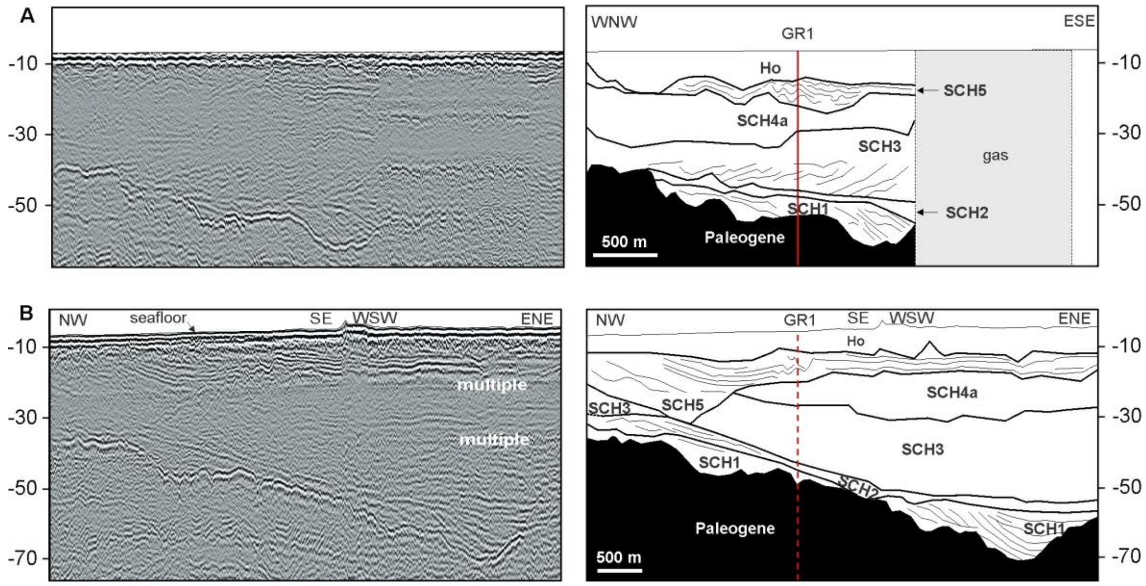


Figure 5.7 Interpreted seismic reflection profiles of the ESC channel in the Ostend Valley. Locations of the cross-sections are plotted in Figure 5.2 B. Please note that the thickness of unit SCH1 in panel A and B differs. This is caused by an offset of several tens of meters of sediment core GR1 to the seismic line in B (striped line). Depths are in meter LAT.

5.4.2.2 Unit SCH2

Description – Unit SCH2 is a 2 to 10 m thick deposit filling in the scour channels WSC and ESC of the Ostend Valley (Figure 5.2, Figure 5.4 D–E, Figure 5.7 and Figure 5.8). The thickness of the unit reaches a maximum of 10 m downstream scour channel WSC. SCH2 consists of well-sorted clayey fine- to medium-grained sand with some shell material that changes downstream into a sequence of well-sorted fine-grained sand. Furthermore oxidised clay laminae, lenses, chunks and flasers with a broken or wavy appearance have been identified. In scour channel WSC, a heterogeneous lag, consisting of coarse-grained sand, rolled clay pebbles, silex and shell fragments, is present at the downstream end of the channel. Seismic reflection data did not reveal any discernible reflections.

Interpretation – The presence of clay lenses/laminae alternated by sand layers likely results from periods of slack-water conditions in a relatively protected tidal environment (Baeteman, 2008; Dalrymple and Choi, 2007; Martinius and van den Berg, 2011; Wells, 1995). This is corroborated by the presence of the diatom species *Paralia sulcata* encountered in the clay laminae (Supplementary Table 5.2), which is typical for a tidal flat environment. A tidal flat environment also explains the discontinuous and wavy expression of clay lenses/laminae suggesting bioturbation occurred (Dalrymple and Choi, 2007). The occasional basal gravel lag may indicate lateral-accretion occurred through migration of tidal channels (Dalrymple and Choi, 2007).

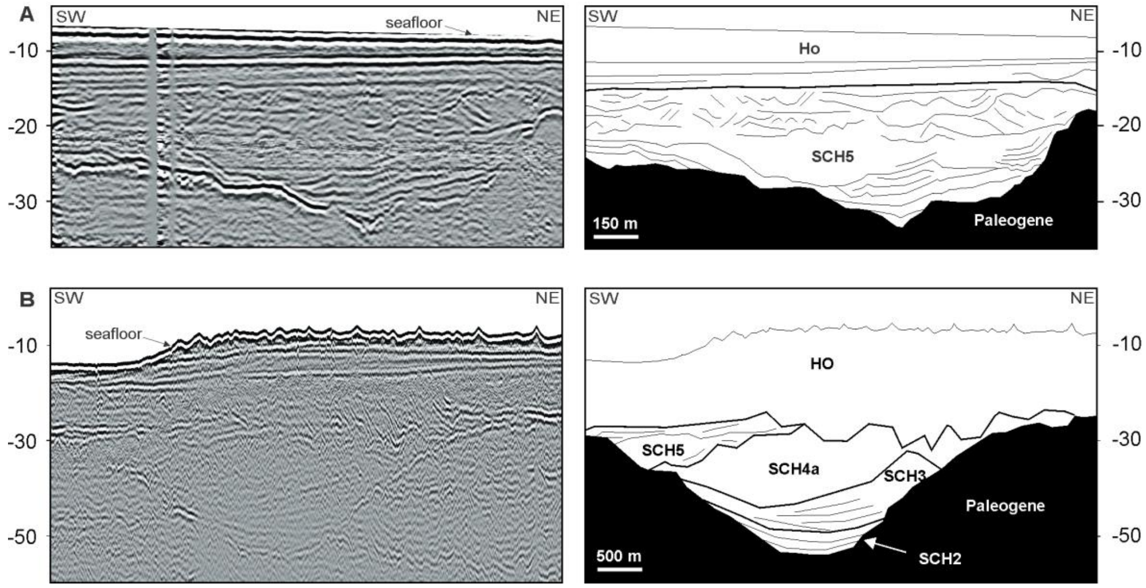


Figure 5.8 Interpreted seismic reflection profiles of the ESC (A) and WSC (B) channels. Locations of cross-sections are plotted in Figure 5.2 B. Depths are in meter LAT.

5.4.2.3 Unit SCH3

Description – Unit SCH3 is a 1.5–13 m thick deposit located in the scour channels WSC and ESC of the Ostend Valley and their downstream bifurcations (Figure 5.2). SCH3 is composed of upward thickening fining-upward sequences each consisting of poorly to well-sorted fine- to coarse-grained sand with clay lenses/laminae (core GR1). The base of SCH3 is characterised by a gravel layer that truncates unit SCH2. Shell material concentrations increase upward and locally accumulate in mm-thick layers. Downstream, SCH3 consists of two 4 m-thick fining-upward sequences composed of poorly sorted clayey fine-grained sand with clay lenses (core OSB). Each sequence has a coarse basal lag composed of broken shell material and gravel. Between the first and second bifurcation of scour channel WSC the unit changes into alternating layers of silty-clayey fine- to coarse-grained sand, with alternating high and low concentrations of gravel and shell material, occasionally mixed with sandy/silty clay mixed with organic material (core BBEL0323 and BBEL0328). Downstream of scour channel ESC, SCH3 consists of fine- to medium-grained sand with lots of shell material and clay fragments (core BR0120003).

In a downstream direction the seismic facies of SCH3 changes from sigmoid-oblique, chaotic reflections (Figure 5.7) or even transparent into low- to high-amplitude reflections alternated with transparent seismic facies (i.e. sand-clay alternations) and zones with high-frequency reflections (thick bedding).

Three shell samples were analysed from unit SCH3 at scour channel ESC (sediment core GR1, Figure 5.9 and Supplementary Table 5.1). Sample M1 is situated just above unit SCH2 and shows marine conditions between fair-weather and storm wave base below 10 m water depth. Sample M2 consists of marine shell fragments deposited in a very-high-energy marine environment. Sample M3 shows a fauna indicating conditions between fair-weather and storm wave base (below 10 m water depth).

Interpretation – Unit SCH3 is interpreted as a landward migrating estuary, i.e. a transgressive system, and is largely equivalent to unit U2 of Mathys (2009). The fining-upward sequences indicate autocyclic cycles of decreasing energy conditions. Similar cycles are observed downstream but are thicker. The abundantly observed sand-clay cycles in the downstream parts suggest daily alternations of high- and low-energy conditions. This is supported by the shell layers in the sand and organic material in the clay. Furthermore, malacology indicates a strong marine influence with water depths up to 10 m and with deposition between fair-weather and storm wave base under high-energy conditions.

5.4.2.4 Unit SCH4a

Description – Unit SCH4a is the most widespread deposit in the palaeo-Scheldt Valley, and the only unit that was directly correlatable both offshore and onshore (Figure 5.6). Upstream in the Coastal Valley the unit reaches 35 m, while downstream in the Ostend Valley the unit thins significantly and disappears at the Middle Scarp. In the Coastal Valley, SCH4a is a glauconitic-bearing deposit composed of fine- to medium-grained sand (Figure 5.3 A) that changes into very silty fine-grained sand (Figure 5.3 C) with a local basal gravel lag. Clay fragments are rare while organic material is ubiquitous and can appear as zones with organic black spots. Shell material is abundant and composed of *Cardium edule*, *Mytilus edulis* and *Ostrea edulis* mixed with Paleogene reworked material. *Hydrobidae* and *Scrobicularia* shells, indicative for a tidal flat environment (Baeteman, 2005b; Catuneanu, 2006; De Moor and Heyse, 1974), are also encountered.

In the Ostend Valley SCH4a is largely equivalent to unit U3 of Mathys (2009). In the scour channels SCH4a shows a transition from very fine-grained silty sand (sediment core GR1) to poorly sorted clayey fine-grained sand (sediment core OSB) to alternations of sand-silt-clay affected by bioturbation and cross-bedding (sediment core nrG). Diatom analysis (Supplementary Table 5.2) shows deposition in a brackish to saline palaeoenvironment with deposition above fair-weather and storm wave base to a full-marine environment (Supplementary Table 5.1).

On the gentle southwest slopes of the Ostend Valley SCH4a consists of well-sorted fine-grained sand with clay laminae (sediment cores SB1 and SB2) that change into fining-upward sequences of coarse- to fine-grained sand (sediment cores UIT, BR150120). The latter contain high concentrations of clay, organic debris and shell fragments. Generally the concentrations of clay laminae increase upwards while shell concentrations decrease.

Seismic reflection data show that prior to the deposition of SCH4a large-scale erosional processes affected the underlying units. The reflections range from transparent (Figure 5.7 and Figure 5.8) to continuous sub-parallel. On the gentle western slopes of the Ostend Valley lots of small channel incisions are observed showing a prograding sigmoidal to oblique infill.

Interpretation – We interpret the fine- to coarse-grained sands and sand-silt-clay alternations to be deposited under marine conditions. The presence of organic material is attributed to the erosion of older or syn-genetic intertidal sediments, while a lag deposit suggests widespread erosion prior to deposition. A brackish to saline palaeoenvironment interpretation is also supported by the diatom and mollusc assemblages as well as by the observed bioturbation (Dalrymple and Choi, 2007).

5.4.2.5 Unit SCH4b

Description – Unit SCH4b is a 1 to 15 m thick deposit located upstream in the Coastal Valley, just south of the Zeebrugge Valley bifurcation (Figure 5.6). Here, SCH4b lies directly on top of the pre-Quaternary surface or is locally overlying unit SCH1 (Figure 5.3 A-B). SCH4b is composed of a calcium carbonate-rich coarse- to very coarse-grained sand with lots of gravel and a mixture of broken Eemian Lusitanian shells, from a shallow marine to intertidal environment, and Paleogene shells. This shell material and gravel can accumulate into massive shell and gravel layers (Depret, 1981).

Interpretation – The coarse-grained sand, the abundance of Paleogene and Eemian Lusitanian broken shell material in combination with the widespread occurrence of gravel suggests a very high-energy environment. This is supported by the widespread erosion observed in geological transects A and B (Figure 5.3). SCH4b shows very similar sedimentary characteristics to the

Zeebrugge, Kaprijke and Oostende Formations (De Breuck et al., 1969; De Moor and Heyse, 1974; Depret, 1981). These formations are interpreted as marine sediments deposited in a very-high-energy environment, possibly an estuary (Heyse, 1979).

5.4.2.5.1 Unit SCH4c

Description – Unit SCH4c is a 1 to 10 m fragmented deposit mostly located on the edges of the Coastal Valley (Figure 5.3). The unit consists of grey silty-clayey fine- to medium-grained sand. The sand locally alternates with mm- to cm-thick weak clay and peat layers. Shell material primarily consists of *Hydrobidae* and *Scrobicularia*, while an occasional gravel fragment may be present.

Interpretation – The clay and peat layers in combination with the shell material indicate deposition in an intertidal environment, an interpretation that was also followed in previous publications (Meetkerke and Houthave Formations: De Breuck et al., 1969; De Moor and Heyse, 1974; Depret, 1981; Devos, 1984; Heyse, 1979).

5.4.2.6 Unit SCH5

Description – Unit SCH5 extends offshore from the Ostend Valley to the Fairy Valley (Figure 5.3). In a downstream direction the width of the infill decreases from ca. 9 km to several hundred meters. Upstream SCH5 consists of well-sorted silty very fine- to fine-grained sand alternated by clay layers, with some shell and organic material (sediment cores GR1 and SB2), changing downstream into a two layered channel fill consisting of a medium-grained sand overlain by a sequence of consolidated clay and peat layers (vibrocore VC-0035). Within the clay accumulations layers of *Bithynia* are present. Malacological evidence (Supplementary Table 5.1) shows a Lusitanian marine fauna occurs at the base (core GR1). Pollen analyses also revealed the presence of freshwater plants (*Pediastrum*, *Ceratophyllum*, *Nymphaeaceae*, *Sparganium*, *Typha atifola*: Figure 5.9 and Figure 5.10).

Seismic reflection patterns in the scour channel WSC show continuous sub-horizontal to slightly wavy reflections, a transparent channel infill, divided by a high-amplitude reflector (Figure 5.7 and Figure 5.8), while in the narrow downstream part of the scour channel ESC parallel to sub-parallel reflections are present followed by a transparent zone, chaotic to oblique/chaotic reflections and on top multiple smaller channel incisions (Figure 5.8).

Interpretation – The silty fine-grained sands alternated with clay and peat layers are interpreted as fluvial sediments in an overall low and stable energy regime locally interrupted by intermittent higher-energy conditions (medium-grained sand). The freshwater flora and the mollusc *Bithynia* corroborate this. The marine molluscs at the base of the unit are interpreted as reworked from the underlying unit SCH4a. The seismic facies characteristics of scour channel WSC are interpreted as a stable and low-energy environment (parallel to sub-parallel draped reflections) followed by more variable higher energy conditions (transparent zone followed by a chaotic to oblique/chaotic reflections). The smaller channel incisions at the top represent a decline in water supply to the fluvial system.

5.4.2.7 Unit SCH6

Description – Unit SCH6 is a 1–10 m thick and 7.5–14 km wide deposit located in the upstream Coastal Valley (Figure 5.2). The unit is marked by a variable lithology: very fine- to medium-grained sand that can become silty mixed with clay lenses, shell material (*Bithynia*) and organic debris; clayey to silty fine-grained sand with low glauconitic concentrations, vegetation horizons and fine shell material; at the top alternating organic sandy silt and coarse-grained sands with shell material and gravel may occur. The base of SCH6 is very sharp and consists of a coarse-grained lag of shell material, gravel and fragments of Paleogene sandstones.

Interpretation – The presence of fine-grained sediments mixed with silt-clay layers and vegetative horizons points towards deposition in a floodplain, abandoned channels or abandoned areas of active channels within a fluvial environment (Miall, 2006). This is corroborated by the presence of *Bithynia* in some locations. The coarse-grained lag at the base of SCH6 suggests that prior to deposition erosional incision occurred. This may explain the presence of glauconitic and sandstone fragments locally derived from the underlying Paleogene strata (Le Bot et al., 2003). The presence of coarse-grained sand alternating with organic sandy silt at the top of SCH6 may be explained by changes in flow regime caused by seasonal fluctuations (Miall, 2006). This facies is comparable to that of the formations of Eeklo and Damme, found in the same area by De Moor and Heyse (1974) and Depret (1981).

5.4.2.8 Unit SCH7

Description – Unit SCH7 is a 1–4 m thick fragmented unit located in the onshore study area that extends further inland (Figure 5.2). SCH7 locally outcrops but is generally covered by Holocene sediments. It consists of well-sorted grey-brown (silty) very fine- to fine-grained sand deprived of calcium carbonate with occasional humic/peaty intercalations. At some locations a gravel lag may occur at the base.

Interpretation – The homogeneous lithology likely originates from local aeolian processes, while the coarse-grained gravel lag at its base may represent a deflation layer, similar as the Middelburg Gravel (Derese et al., 2010; Heyse, 1979). The humic to peaty intercalations on the other hand are formed by swampy conditions that developed in nearby depressions (Heyse, 1979). This facies is comparable to that of the Maldegem Formation, identified by De Moor and Heyse (1974).

5.5 Chronology

Age control on the Eemian interglacial sequence in the Ostend Valley was obtained by pollen stratigraphy complemented by eight OSL dates (core GR1: Table 5.3). In the Ostend Valley Outflow system, chronological control was based on 11 pollen samples (core VC-0035).

5.5.1 Pollen stratigraphy

Pollen analysis for core GR1 shows a dominance of three superimposed vegetation zones that reflect the succession of a (warm) temperate forest. The Holocene-Pleistocene boundary is located at ca. 10 m depth. No pollen samples were taken from fluvial unit SCH1. Pollen concentrations from the tidal-fluvial unit SCH2 are low but yielded sufficient pollen for interpretation. The transgressive estuarine sediments of unit SCH3 were sterile for pollen and is attributed to severe reworking before deposition. The marine unit SCH4a and fluvial unit SCH5 show a wide range in pollen counts but generally they were sufficient for environmental interpretation. Pollen counts for the Holocene sediments were very good, but were not further used in this study. Pollen analysis for vibrocore VC-0035 shows a dominance of a (warm) temperate forest that can be correlated upstream to core GR1 as fluvial unit SCH5. Reasonable pollen concentrations were present in the medium-grained basal section of vibrocore VC-0035 and show indication of reworking. The upper clay and peat layers are rich in pollen.

5.5.1.1 Ostend Valley: sediment core GR1

From sediment core GR1 17 samples between 0 and 45 m were processed for pollen analysis (Figure 5.9). Unit SCH2 is dominated by *Corylus*, *Quercus* and *Pinus* pollen while *Carpinus*, *Ulmus*, *Taxus*, *Picea* and *Dryopteris* are present in lower concentrations. Minimal percentages of herbs and Poaceae are also observed. Based on the dominance of *Corylus*, and the record of *Picea* and *Taxus*, an Eemian correlation is likely. These spectra indicate a Middle Eemian PAZ E4 age and suggest mixed temperate woodland existed in the upstream regions during deposition. *Picea* and *Taxus* first appear in Eemian spectra in Belgium and the Netherlands during PAZ E4 (Bogemans et al., 2016; Mostaert and De Moor, 1984, 1989; Zagwijn, 1996). No pollen samples for unit SCH1 were taken indicating a youngest possible age for this unit of PAZ E4. Unit SCH3 was sterile for pollen, diatoms, dinoflagellates and foraminifera. Bracketing of this unit by pollen bearing units SCH2 and SCH4a, however suggests it was deposited in the Middle Eemian PAZ E4 and/or early PAZ E5. For unit SCH4a pollen count is reasonable to good except for the interval between 25 and 26 m. Up to 18 m, SCH4a is dominated by *Carpinus*. In between peaks of *Carpinus Pinus* is the dominant tree species. *Quercus* is present at persistent intermediate concentrations, while *Corylus* is present at low concentrations. The first two acmes in *Pinus* show a contemporaneous increase in *Picea*, herbs and heathland along with the local *Dryopteris* and the aquatic *Pediastrum*. *Pediastrum* concentrations show a continuous increase in the second half of the biozone. The dominance of *Carpinus*, in the absence of reworking, suggests it was deposited in the Middle Eemian PAZ E5 (Zagwijn, 1996). In the upper part of unit SCH4a, consisting of very fine-grained silty sands, *Carpinus* disappears while *Quercus* and *Pinus* pollen concentrations increase. *Corylus*, *Ulmus*, *Alnus* and *Betula* are present at persistent, yet low percentages. This persistent decline in tree species is compensated with an increase in herbs. Poaceae and Chenopodiaceae are dominant herbal species, while *Armeria*, Cyperaceae and *Polypodium* are present at lower concentrations. *Dryopteris* remains present at the same to slightly increasing concentrations compared to PAZ E5. *Pediastrum* concentrations show a marked increase in the second half of the biozone. The assemblage suggests the upper part of SCH4a was deposited in Middle-Late Eemian PAZ E5 and early PAZ E6. The pollen assemblages from the upper unit Ho contain arboreal elements (particularly high concentrations of *Pinus*), which indicate a Preboreal to Subboreal age for this unit. However, in the uppermost levels *Secale cereale* indicates a medieval or younger age for these sediments (Verbruggen et al., 1996).

In unit SCH5, *Pinus* is the dominant tree species and becomes even more dominant upward. Other tree species, such as *Quercus*, *Corylus* and *Ulmus*, are on the decline similar to herbs, which reached a maximum in the lower part of unit SCH5. Poaceae, Chenopodiaceae and *Armeria* are present in highest concentrations, while *Dryopteris* is present in the same persistent concentrations as in unit SCH4a. *Pediastrum* concentrations increase and show a peak halfway. At this same acme *Botryococcus* shows a peak in concentration. The assemblage suggests that unit SCH5 was deposited in the Late Eemian PAZ E6.

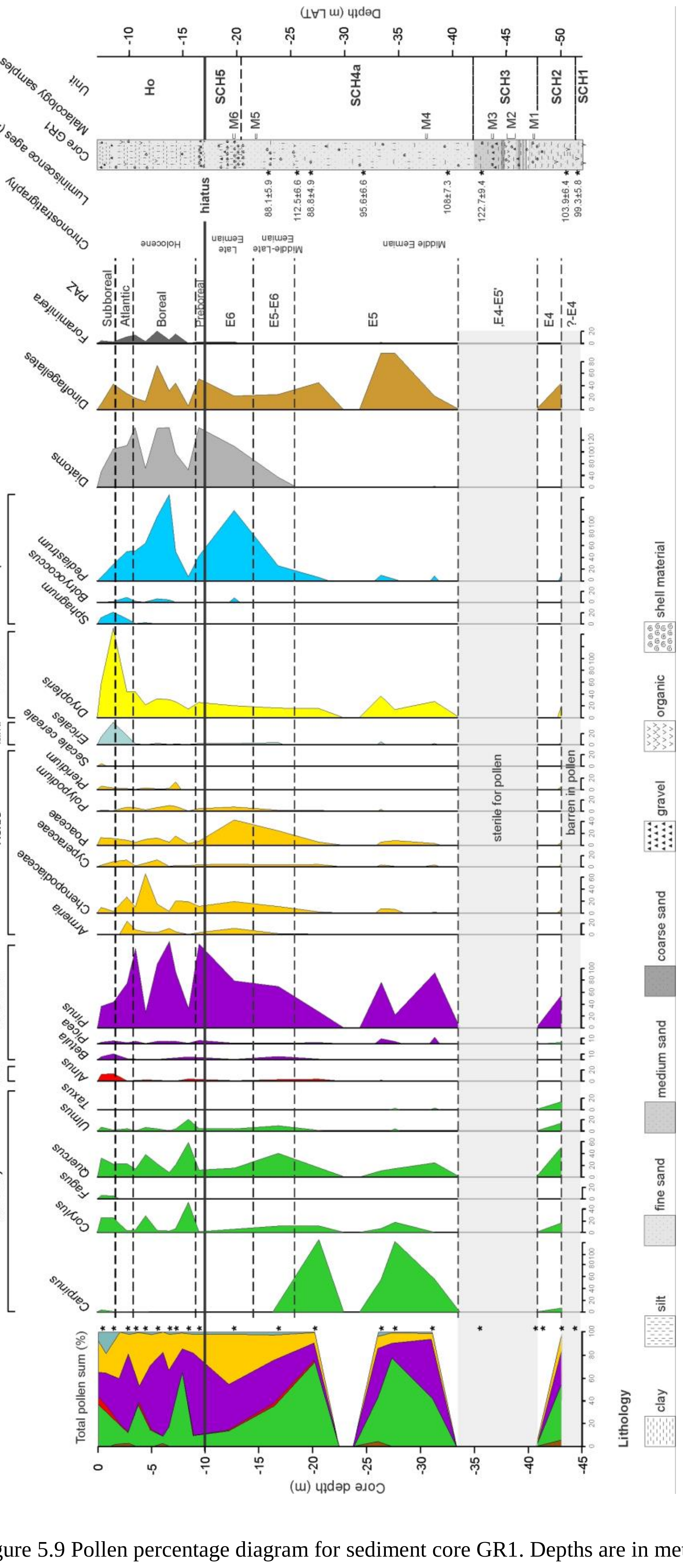


Figure 5.9 Pollen percentage diagram for sediment core GR1. Depths are in meter LAT.

5.5.1.2 Downstream part of the Ostend Valley: vibrocore VC-0035

From vibrocore VC-0035, 11 samples between 3 and 5.3 m depth were processed for pollen analysis (Figure 5.10). *Pinus* and *Betula*, which show inversed peaks, dominate the pollen assemblage for unit SCH5. Herbs, such as Poaceae, *Artemisia* and Cyperaceae, as well as *Dryopteris* are present at relative stable concentrations. From the freshwater species a dominance of *Ceratophyllum* is observed. Between 3.5 and 4.4 m depth *Pinus*, *Dryopteris* and *Ceratophyllum* show a significant drop with a contemporaneous increase in *Betula*, Poaceae, *Nymphaeaceae* and *Sparganium*. The dominance of *Pinus* suggests SCH5 was deposited in Late Eemian PAZ E6 (Zagwijn, 1961, 1996), while the intermittent peaks in *Betula* and Poaceae would represent short (cold) events and/or a local dry phase, within the Late Eemian PAZ E6. The manifestation of such events is in agreement with the general notion of the regional climate and vegetation developments over the transition of the Eemian to the Early Weichselian in Europe (Sirocko et al., 2007; Helmens, 2014). No pollen for Unit Ho were analysed due to a transgressive lag and evidence from seismic reflection data suggesting Holocene transgressive marine sediments are located on top of the channel infill.

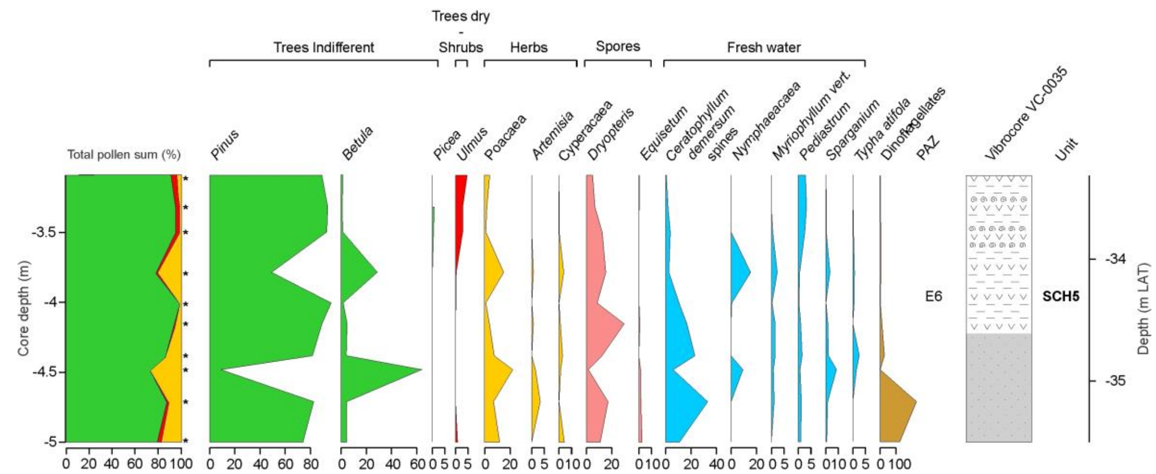


Figure 5.10 Pollen percentage diagram for vibrocore VC-0035. The lithological legend can be found in Figure 5.9. Depths are in meter LAT.

5.5.2 OSL dating

OSL dating results from core GR1 indicate that units SCH1 to SCH4a in the Ostend Valley were deposited between 88 and 112 ka (Table 5.3) suggesting a Weichselian Early Glacial age (MIS 5d–b). For the samples of units SCH1 to SCH3, there is no trend of age with depth, indicating that all samples are of similar age, estimated at 101 ± 5 ka. The sample above (NCL-7616111) returns an age in the similar range, and if this sample is also taken into account a total mean age for units SCH1 to SCH5 of 99 ± 5 ka is obtained. However, these results should be treated with caution, as palaeodoses are near quartz OSL saturation, and previous research has indicated that equivalent doses in this range may be underestimated (e.g. Timar-Gabor and Wintle, 2013). Interestingly, there seems to be a slight dependency of age on dose rate, with lower ages obtained for samples with high dose rate. This could hint at an age underestimation related to near-saturation of the OSL signals. Although the quartz OSL dating results suggest a slightly younger age than the expected Eemian pollen zones E4-E6, the results lack robustness to draw firm conclusions. The dose recovery ratio and slight dependency of age on dose rate provide reason to believe that the OSL ages may slightly underestimate the true burial age (see Supplementary Information). The OSL dating results did not provide an absolute chronology for the identified units.

5.5.3 Eemian sea-level curve

Comparing the Belgian (this study) and the Netherlands (Peeters et al., 2016) Eemian sea-level curves shows that the sea-level rise in Belgium lagged behind the Netherlands (a thousand years or less) and that the initial rise was faster in the Netherlands compared to Belgium (Figure 5.11). No direct evidence of a sea-level rise is observed within the small dataset here. High-stand conditions were achieved in the Netherlands during PAZ E4–E5 while in Belgium sea level continued to rise and reached high-stand during the PAZ E5–E6 transition or even during E6. The duration of the high-stand in Belgium is not known but may be comparable to that observed in the Netherlands. The highest Eemian sediments were found upstream the palaeo-Scheldt Valley (unit SCH4a) at a depth of ca. 2 m below LAT (or 0.8 m msl). Two additional values from other studies were added to the sea-level curve: high-stand peat deposits from Heyse (1979) upstream the palaeo-Scheldt Valley (-2 m msl) dated at PAZ E4a; and estuarine sediments from Bogemans et al. (2016) from the Yser Valley (1 msl) dated at PAZ E4. This age is likely not the true age for the whole estuarine sequence as only the lower part of this section is correlated to PAZ E4. The upper estuarine sequence likely represents PAZ E5 and possibly also E6. Overlying Weichselian fluvial channel sediments may have removed part of the upper Eemian sequence. For this reason a continuation was made to younger ages on the sea-level curve in Figure 5.11 (striped line). From this the result of the Yser Valley and the result of Heyse (1979) appear not to fit with the presented sea-level curve of the palaeo-Scheldt Valley as they reach higher sea-level values much earlier in time, This discrepancy may be the result of not taking into account compaction resulting in an underestimation of sea levels. On the other hand, the top of unit SCH4a appears to fit the maximum observed sea-level positions observed by Bogemans et al. (2016) and Heyse (1979). Finally, the sea-level rise in Belgium achieved higher conditions than observed in the Netherlands.

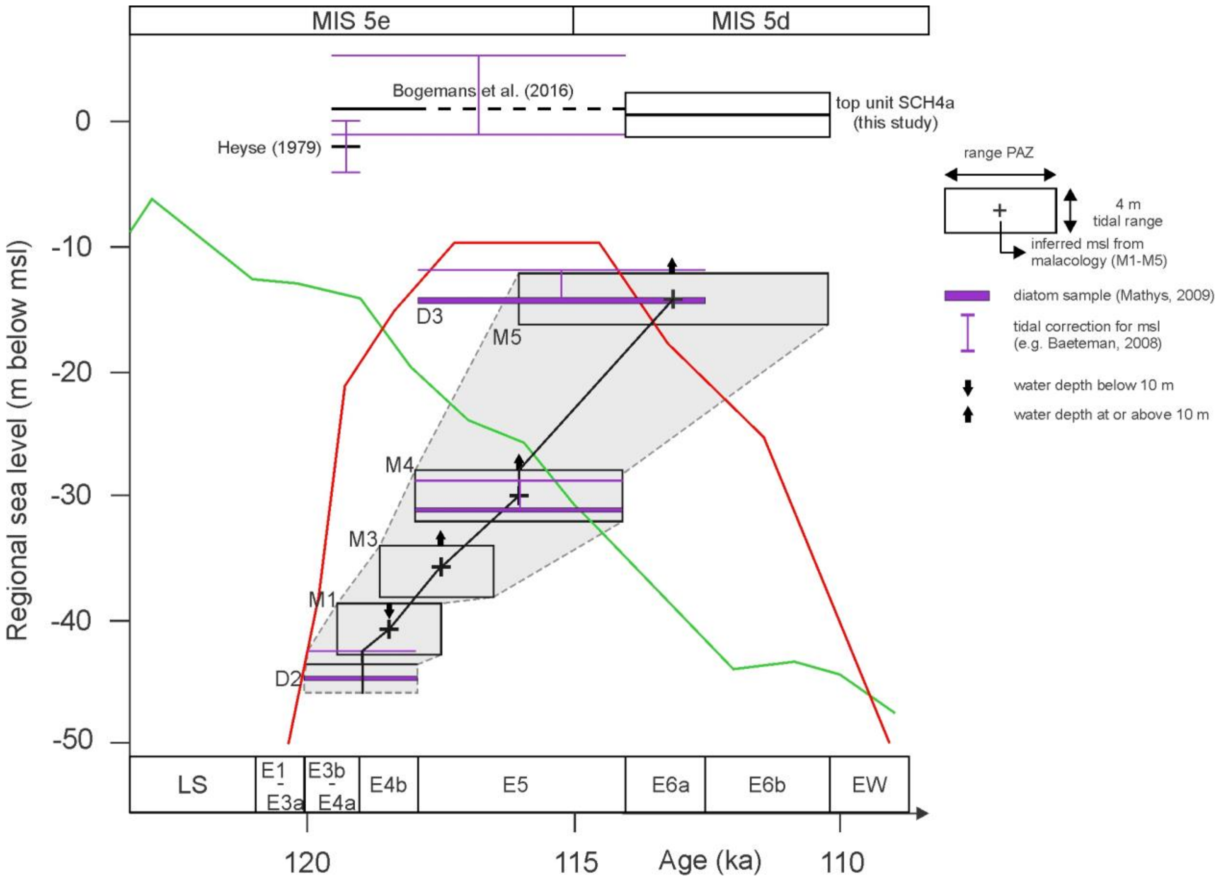


Figure 5.11 Diagram showing the Eemian interglacial sea-level envelope of Belgium (black) compared to the sea-level curve of the Netherlands (the sea-level curve of Zagwijn, 1983, 1996 modified by Peeters et al., 2016) for the southern North Sea. The green sea-level curve is the global background sea-level curve of Lisiecki and Raymo (2005, 2009) based on the LR04 $\delta^{18}O$ record.

5.6 Palaeogeographic evolution

5.6.1 Late Saalian

The dynamic oscillation and unknown thickness histories of the main masses of the Saalian ice-sheets covering British and Scandinavian bedrock further north resulted in complicated glacio-isostatic movements in northwest Europe (Ehlers et al., 2011; Lambeck et al., 2006; Lang et al., 2018). The most southern extension of the Drenthe Stage glaciation, ca. 160–155 ka ago (Busschers et al., 2008; Lang et al., 2018; Toucanne et al., 2009b), occurred ca. 30 ka before the onset of the Eemian interglacial (e.g. Peeters et al., 2015, 2016). The following Warthe substage of the glaciation, ca. 140 ka ago (Lambeck et al., 2006), had a similar configuration as the Weichselian Last Glacial Maximum across North Germany and Denmark. These ice-sheet configurations in combination with glacio-isostatic modelling (Lambeck et al., 2006) position the Netherlands in the vicinity of the forebulge periphery during the onset of the Eemian interglacial. The inferred presence of a supposed land bridge between East Anglia and Belgium (Busschers et al., 2008; Gibbard and Cohen, 2015; Hijma et al., 2012) combined with the above ice-sheet extent observations and geophysical modelling suggest Belgium to have been positioned on this peripheral forebulge, which may in turn explain the depths and thickness of the incised-valleys of this study. It is well possible (given uncertainty in exact ice masses and regional variance in crust-mantel rheological properties), that the study area was at the highest point of forebulge upwarping. This is supported by the sea-level curve (Figure 5.11) that shows that the study area experienced a longer sea-level rise than the Netherlands (when not considering the compaction problem). The forebulge is likely a combination of a structural topographic ridge related to Eocene-Oligocene substrate at the rim of the Neogene-Quaternary North Sea depocentre and erosional phases prior to the Saalian cuesta formation in response to Elsterian events (e.g. Busschers et al., 2008; Gibbard and Cohen, 2015; Hijma et al., 2012) that gained extra elevation owing to forebulge glacio-isostatic adjustment. It is this ridge, or at the time in question, a land bridge, that blocked the Late Saalian drainage in the direction of the Dover Strait resulting in the formation of a proglacial lake.

Busschers et al. (2007) demonstrate that the Weichselian Late Pleniglacial Rhine-Meuse system strongly responded to glacio-isostatic crustal movements, first by deflection and incision during glacio-isostatic uplift, i.e. the formation of the peripheral upwarped zone, followed immediately by infill of the incised structure and deflection in the opposite direction. As the palaeo-Scheldt flows in a southwest-northeast direction across the peripheral upwarped zone (Figure 5.12 A) a similar response as the Weichselian Late Pleniglacial Rhine-Meuse system is to be expected. In the study area, such response could even be stronger expressed, for reasons of the nature of the substratum (semi-consolidated Paleogene clay and sand formations; instead of Middle Pleistocene sands), magnitudes of discharge passing the study area (Late Saalian Axial Valley incision at downstream end of studied incised valleys), and magnitude of the Drenthe-to-Warthe stage forebulge upwarping, compared to that of the Last Glacial Maximum Rhine. The glacio-isostatic upwarping of the study area induces tilting of the river valley, as the northern section was updoming more than the southern section. As rivers naturally tend to gravitate toward the subsided zone (e.g. Holbrook and Schumm, 1999) the palaeo-Scheldt River north of Ghent deflected to the west in the direction of Zeebrugge and Ostend where it enters the present-day

BCS (Figure 5.12 A and B). Shortly after the palaeo-Scheldt had repositioned to a more western trajectory a distinct incised-valley system developed with incisions up to ca. 20 m (Figure 5.4). Thus, rather than a response to climate change the incision establishing the main studied incised-valley can be interpreted as compensating response to syn-depositional glacio-isostatic updoming. Following mechanisms listed by Holbrook and Schumm (1999), updoming of the area between East Anglia and Belgium disturbed the longitudinal profile of rivers and triggered downcutting to maintain equilibrium (Busschers et al., 2007; Cohen, 2003). In the study area, the deflected palaeo-Scheldt River crosses formations of variable consolidated lithologies ranging from consolidated sand to clay (Le Bot et al., 2003). These variable lithologies likely influenced the speed of the downcutting process. For example the presence of cm-thick sandstone layers near Zeebrugge may have slowed down the downcutting process, while downstream near Ostend this downcutting process may have been accelerated due to the more homogeneous clay composition of the Paleogene strata. Similarly, the orientation of the palaeo-Scheldt may be influenced by the variable lithology of the Paleogene layers. The Coastal-Ostend Valley transition is characterised by a change in orientation from almost normal to the Paleogene strata to almost parallel in the Ostend Valley. Finally, the palaeo-Scheldt River responded in such a way that it captures the Lys and Waardamme Rivers and formed the Coastal–Ostend–Zeebrugge Valley system. The Waardamme and Zeebrugge Valleys merge downstream in to what is known as the Thornton Valley (see Mathys, 2009).

The stepwise downcutting process in the Coastal Valley across Paleogene layers of lithological variable compositions achieves depths of ca. 20 m below the Middle Platform. This incision depth is about the same as reached by the Axial Channel in our study area. Downstream in the Ostend Valley the two relatively straight channel incisions WSC and ESC, cut in a homogeneous consolidated clay lithology, achieve depths up to 10 m deeper than that of the upstream Coastal Valley and other downstream sections of the Ostend Valley. The deep incisions of the WSC and ESC channels may also be stimulated by the slope from the Marginal Platform (that is now being divided into three segments by the palaeo-Scheldt River) in conjunction with the updoming forebulge where the highest slopes are achieved. Remarkably, no strong incisions are observed beyond the Ostend Valley except for the continued bifurcations downstream of the WSC and ESC channels. These channels may also have been formed during the Late Saalian based on their infill by younger Middle Eemian PAZ E4–E5 sediments (unit SCH3: Figure 5.4 G) and were later used as tidal channels in the Eemian estuary. Presumably, waves and tidal currents have removed further downstream incisions during the Eemian marine transgression, similar with Holocene-known erosion of this type. These are the type of processes that would have deposited unit SCH1, in the west of the study area starting in the Early Eemian already.

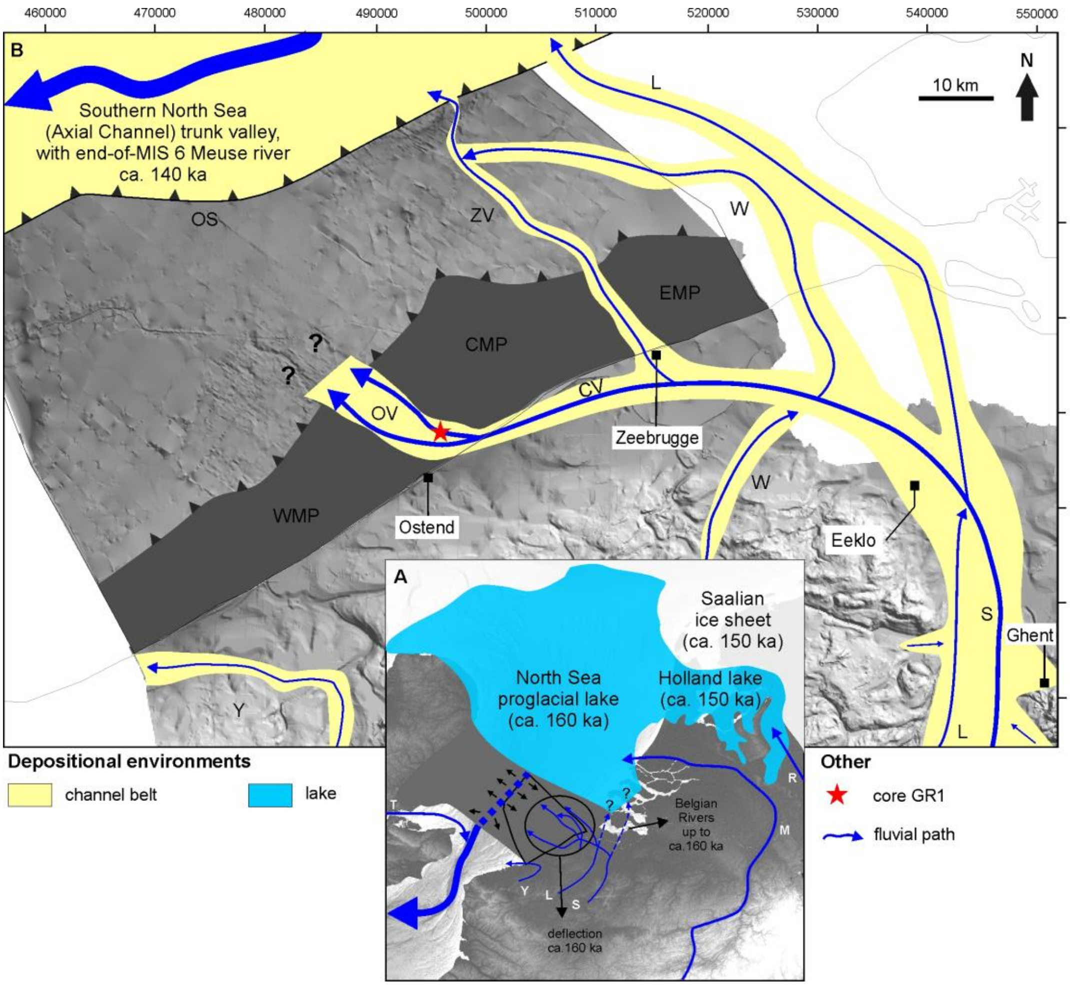


Figure 5.12 Regional (A) and local (B) scenario for the end of the Saalian (latest Saalian; MIS 6-5e transition). The regional scenario shows the northwest European landscape ca. 160–150 ka ago, while the local scenario shows the local landscape after a North Sea proglacial lake (Gibbard and Cohen, 2015) has breached a land bridge forming the Southern North Sea 'Axial Channel' trunk valley ca. 140–130 ka ago. Saalian ice sheet and proglacial lake from Busschers et al. (2008), Hijma et al. (2012) and Peeters et al. (2015, 2016). The Holland Lake in the IJssel Basin is formed after the maximum Saalian glaciation, i.e. when the ice sheet started to retreat (Busschers et al., 2008). The Rhine River flows into this basin, while the separate Meuse River follows a path south towards the land bridge: Busschers et al. (2008), Hijma et al., (2012), Peeters et al. (2015, 2016). The Belgian Rivers prior to deflection: De Moor and Pissart (1992). The offshore Lys and Waardamme: Ebbing and Laban (1996). Meuse River on BCS: Mathys (2009) and Hijma et al. (2012). Yser River: Liu (1990) and Bogemans et al. (2016). The reader is referred to Figure 5.1 for abbreviations used.

When the land bridge between East Anglia and Belgium was overtopped and breached (following scenarios in Busschers et al., 2008; Gibbard and Cohen, 2015; Gupta et al., 2017) additional downcutting of the Axial Channel trunk valley took place. This process likely initiated additional vertical incision in the already established palaeo-Scheldt Valley. By consequence, the palaeo-Scheldt Valley was already established when the downcutting of the Axial Channel trunk valley, following the overtopping of the land bridge in the Late Saalian, takes place. Additional vertical incision is similarly observed in the Meuse Valley downstream as it erodes up to 10 m of Paleogene sediments (Hijma et al., 2012) and forms the Offshore Scarp on the BCS (Figure 5.12 B; Mathys, 2009). Finally, when the Saalian glaciation transitions into the Early Eemian (PAZ E1-E3), the palaeo-Scheldt, Zeebrugge, Waardamme and Lys Rivers, retained their former positions and consolidated their course (Figure 5.13 B).

5.6.2 Eemian interglacial

At the very end of the Saalian glaciation, climate in northwest Europe ameliorated, defining the onset of the Eemian interglacial. The Eemian interglacial is characterised by rapidly rising temperatures (e.g. Brewer et al., 2008; Kaspar et al., 2005; Kukla et al., 2002). During high-stand, eustatic sea-level exceeded modern levels by at least 5.5 m (Dutton and Lambeck, 2012; Kopp et al., 2009; Medina-Elizalde, 2013). These climate changes and the sea-level change have been recognised in successive pollen zones (E1–E6) across northwest Europe (Zagwijn, 1961, 1983, 1996).

The Eemian transgression entered the study area via the Dover Strait and from there extended northward into the southern North Sea and its valleys in the lowlands between present-day Belgium and England (Figure 5.13 A). The resulting Eemian deposits in the southern North Sea are quite thin and fragmented and are mostly present north-northeast of the BCS (Balson et al., 1992; Hijma et al., 2012; Rijsdijk et al., 2005). Recent studies in the Netherlands Rhine delta/estuary have made the case that the Eemian sea-level high-stand in the southern North Sea has a *regional* character as it lagged the global eustatic high-stand by several thousand years (Peeters et al., 2016; Sier et al., 2015). The observed lag in regional high-stand is attributed to glacio-isostatic adjustment in the periphery of the former Saalian ice-sheet and associated mechanisms extending to the southern North Sea near-field setting (Cohen et al., 2012, 2017; Lambeck et al., 2012; Long et al., 2015). Based on the short distance to the Rhine delta/estuary (ca. 250 km) from the study area and a similar position on the, at the time subsiding forebulge, it is to be expected that a similar time lag is present in the Eemian transgression of the study area. In other words: the palaeo-Scheldt infill could be of use to test if the lagged timing and the magnitudes of sea-level rise occurring within the Eemian reproduce.

5.6.2.1 Early Eemian: PAZ E1–E3

Pollen analysis of the Adegem Formation in the Coastal Valley section (De Moor and Heyse, 1974) shows the presence of a broad-leaved forest, dominated by *Alnus*, *Corylus* and *Quercus* (Heyse, 1979) typical for an Eemian setting (Zagwijn, 1961, 1996). This is supported by the lateral accretion of point bars and upward-fining cycles in unit SCH1 suggesting a stable river system, probably representing the infilling of a channel on an actively aggrading floodplain.

At the same time, the Eemian North Sea was already situated at the Offshore Scarp (Figure 5.13 B). This scarp acted as a 'cliff line' and prevented a rapid marine inundation of the Inner-Middle BCS. Since the western edge of this scarp, near the Belgian-French border, is located ca. 10 meters lower than its eastern counterpart near the Belgian-Dutch border, this is likely the first location where the Eemian North Sea transgresses the Middle BCS. This is further supported by the downstream tributaries of the palaeo-Scheldt River, filled with the estuarine unit SCH3, that flow to the distal southwest area of the peripheral upwarped zone (Figure 5.13 B). As a result a shallow estuary is formed in the western approaches of the Middle BCS where tidal-estuarine-marine sedimentation takes place (units SCH2-SCH3-SCH4).

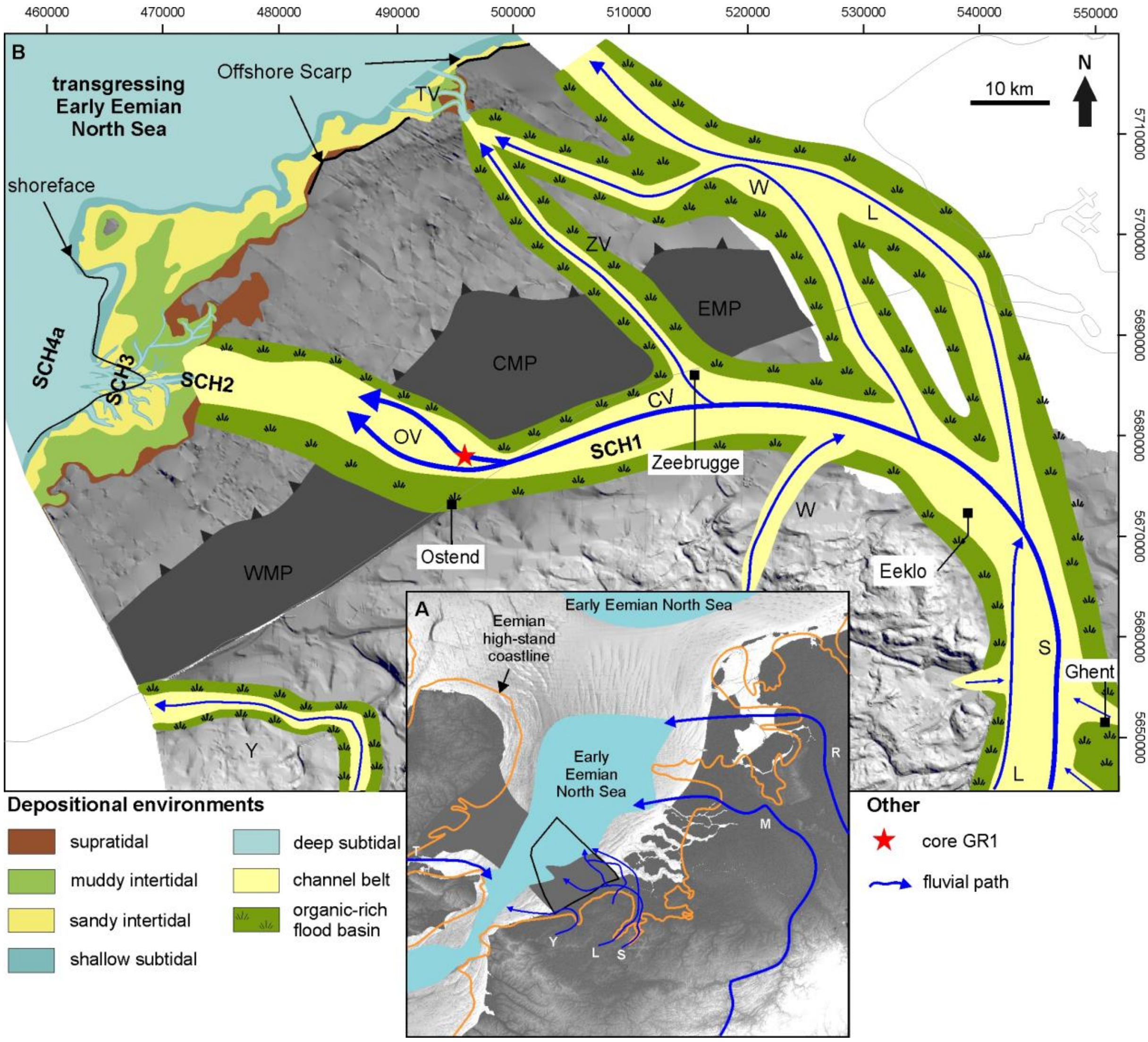


Figure 5.13 Regional (A) and local (B) scenario for the Early Eemian (MIS 5e, inferred PAZ E1–E3). The regional scenario shows the northwest European landscape during the Early Eemian transgression. Eemian high-stand coastline: Streif (2004), Mathys (2009) and Hijma et al. (2012). Rhine and Meuse Rivers: Busschers et al. (2008), Hijma et al., (2012), Peeters et al. (2015, 2016); Belgian Rivers: this study. Offshore Lys and Waardamme Rivers: Ebbing and Laban (1996). Yser River: Liu (1990) and Bogemans et al. (2016). The reader is referred to Figure 5.1 for abbreviations used.

5.6.2.2 Middle Eemian: PAZ E4–E5

During the climatic optimum of the Eemian interglacial, the palaeo-Scheldt Valley completely transformed into a tidal estuary with a funnel-shaped morphology (Figure 5.14 B). The fluvial environment (unit SCH1) migrates upstream and is now located in the southwest Netherlands. At the same time the fluvial-tidal transition zone as well as the inner part of the middle estuary (unit SCH2) are located near the Ostend-Coastal Valley transition. These highly variable environmental conditions can well explain the great variability in the degree of bioturbation (e.g. Dalrymple and Zaitlin, 2006). At the same time the marine-dominated middle-outer estuary (unit SCH3) was situated downstream in the Ostend Valley and rapidly transgresses landward. This process is accompanied by shoreface erosion removing older sediments, such as unit SCH2, from channels and levelling the pre-Quaternary surface (Figure 5.4 D–E). The erosional shelf or the landward migrating shoreface marks the transition of the outer estuary towards the open sea (unit SCH4a).

The Zeebrugge Valley had now detached from the direct fluvial influence of the palaeo-Scheldt River and has consequently transformed into a tidal-influenced river. For the Waardamme and Lys Rivers of the time, we cannot verify the tidal situation with the current data. It is however safe to assume that these rivers start to experience increased tidal forces by the transgressing Eemian North Sea.

It is generally assumed that the Middle Scarp marks the downstream boundary of the Ostend Valley (De Clercq et al., 2016; Liu et al., 1993; Mathys, 2009). However, its relationship to the valley has never been explored. Seismic reflection data show that the shallow incised channels originating from the Ostend Valley cross and erode the Middle Scarp. Indeed, in the Ostend Valley these incised channels are filled with the Middle Eemian unit SCH3 followed by the younger Late Eemian unit SCH5 (Figure 5.4 F). At the same time unit SCH3 is absent downstream of the Middle Scarp (Figure 5.4 G) suggesting the latter has a youngest possible age dating to the Middle Eemian. According to Hijma et al. (2012) the Middle Platform, and consequently also its inland scarp, probably represents an older pre-Eemian terrace than the Offshore Platform, as it is located at a higher position. If the Middle Platform is older than the Offshore Platform and it is indeed the result of fluvial erosion than it has to pre-date the Late Saalian proglacial lake spillway erosion stage.

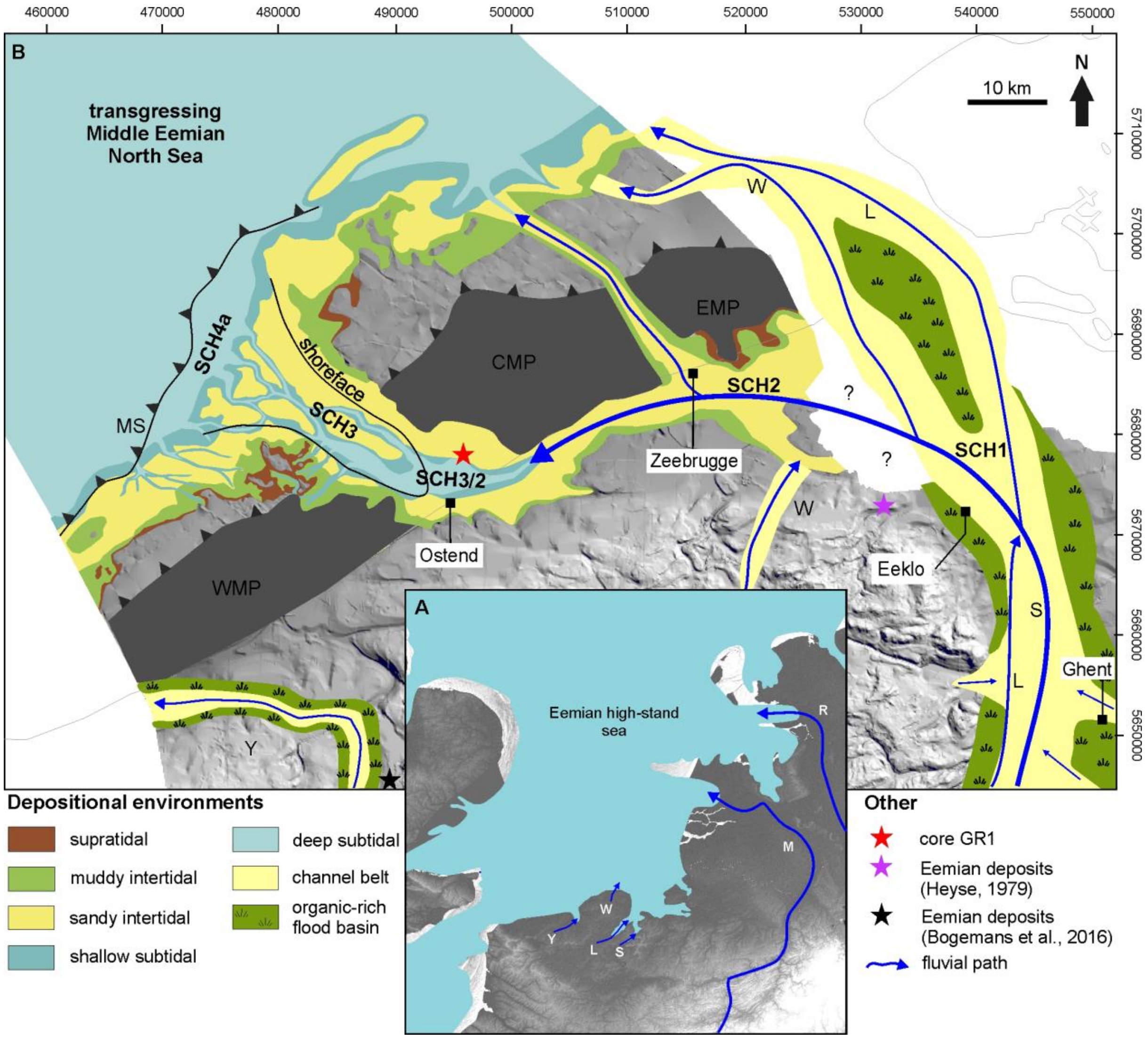


Figure 5.14 Regional (A) and local (B) scenario for the Early-Middle Eemian (MIS 5e, PAZ E4–E5). The regional scenario shows the northwest European landscape during the Eemian high-stand as a reference to the local scenario. The local scenario shows the landscape when eustatic sea level is at -25 m. Rhine and Meuse Rivers: Busschers et al. (2008), Hijma et al., (2012), Peeters et al. (2015, 2016); Offshore Lys and Waardamme Rivers: Ebbing and Laban (1996). Yser River: Bogemans et al. (2016). The reader is referred to Figure 5.1 for abbreviations used.

5.6.2.3 Late Eemian: PAZ E5

During the Late Eemian (PAZ E5), the palaeo-Scheldt Valley is completely drowned and subject to full (shallow) marine conditions (unit SCH4a) through the Ostend Valley (Figure 5.15 A). The remaining segments of the Marginal Platform now provide protection to the palaeo-Scheldt and Yser Valleys from direct wave action. Tidal forces entering the palaeo-Scheldt Valley were up to now directed through the Ostend Valley. However, with rising sea level the former Zeebrugge Valley transformed into an estuary enabling tidal forces to penetrate the palaeo-Scheldt Valley through this secondary opening. This resulted in widespread erosion of older sediments that are redeposited as coarse- to very coarse-grained sands with gravel and shell material (unit SCH4b). These sediments are observed up to 20 km upstream of the Zeebrugge Valley (De Moor and Heyse, 1974) and suggest these newly introduced tidal forces had a widespread impact on the palaeo-Scheldt Valley infill.

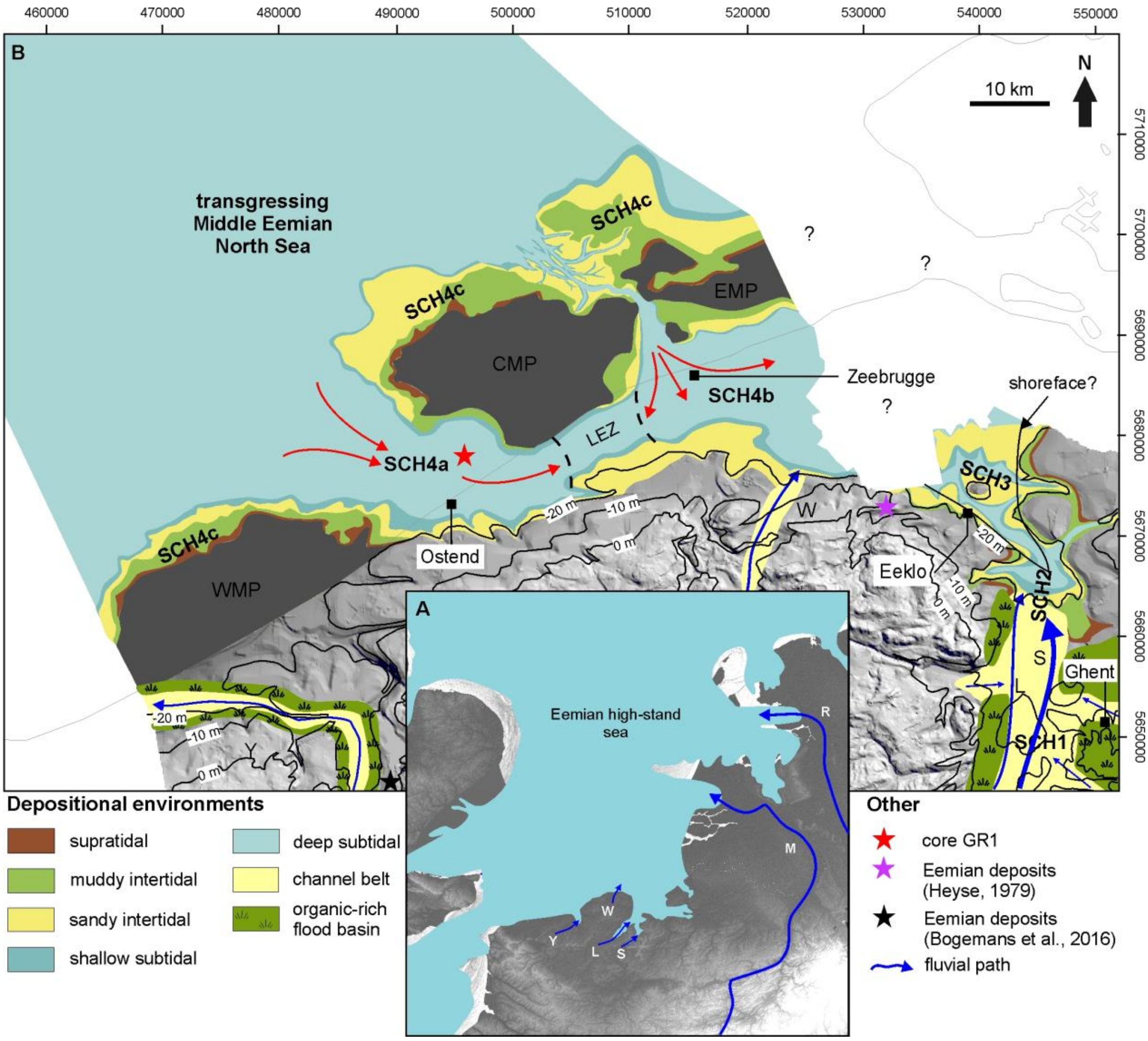


Figure 5.15 Regional (A) and local (B) scenario for the Middle Eemian (MIS 5e, PAZ E5). The regional scenario shows the northwest European landscape during the Eemian high-stand as a reference to the local scenario. The local scenario shows the landscape when eustatic sea level is at -20 m. Rhine and Meuse Rivers: Busschers et al. (2008), Hijma et al., (2012), Peeters et al. (2015, 2016); Offshore Lys and Waardamme Rivers: Ebbing and Laban (1996). Yser River: Liu (1990) and Bogemans et al. (2016). The reader is referred to Figure 5.1 for abbreviations used.

The coarse-grained sands however, are not encountered downstream of the Zeebrugge Valley. Such can be explained as a result of opposing tidal forces entering the Coastal Valley, i.e. from the Ostend Valley in an upstream direction and from the Zeebrugge Valley in a downstream direction (red arrows Figure 5.15 A). Where both opposing tidal forces meet a low-energy zone (LEZ) is formed. This LEZ can be traced in unit SCH4a as increased silt and organic concentrations (Figure 5.4 B–C) and is comparable to the mixed energy zone where fluvial and estuarine tidal forces meet (Dalrymple and Choi, 2007), i.e. the boundary between the middle estuary and the fluvial-tidal transition zone (unit SCH2). This LEZ explains the presence of the badly-sorted medium-grained sands of the K1 Formation followed by the silty very fine-grained sands of the K2 Formation in area 4 of Devos (1984) (Figure 5.1 C; Table 5.2).

In the Ostend Valley the sediments of unit SCH4a consist of silty fine-grained sand with clay layers and cross-bedding features that may indicate hummocky cross-stratification (Trentesaux et al., 1999). According to Davis and Dalrymple (2012) this may be related to the stormy winter season while the presence of bioturbation may be related to the less-energetic summer period. The widespread occurrence of organic material can then be attributed to the erosion of intertidal flats and supratidal marshes along the edges of the estuary during stormy winter events.

5.6.2.4 Late Eemian: PAZ E6

During the final stages of the Eemian, sea-level rise is slowing down in the study area nearing almost a high-stand (Figure 5.16 A). The high-stand coastline reached far inland compared to the present-day coastline and its deposits are found at different elevations across northwest Europe (Streif, 2004b). In the Thames estuary and the Wash, high-stand sediments are found up to +7 m present-day sea level (Bridgland, 1994) and up to +2 m along the East Anglian coastline (Jones and Keen, 1993). Along the northwest European coastline they are found at ca. 1 m in the Yser Valley in southwest Belgium (Bogemans et al., 2016) ca. -2 m in northwest Belgium (Heyse, 1979), -8 m at Amersfoort and -9 to -10 m in the northeast area of the Rhine estuary in the Netherlands (Peeters et al., 2016; Zagwijn, 1983) and in Germany as low as -9 m in Lower Saxony (Streif, 2004b). In the study area, Eemian high-stand sediments are found upstream in the palaeo-Scheldt Valley (Figure 5.4 A) at -2 m LAT (or 0.8 m msl) and consist of the marine sediments of unit SCH4a. These sediments have likely been partially removed by later fluvial action (unit SCH5).

The approaching high-stand position in the study area is demonstrated by maximum water depths of ca. 10 m inferred from core GR1 (Supplementary Table 5.1) indicating accommodation space decreased above this level as sedimentation outstripped the slowing rate of sea-level rise. This is likely caused by the terminal phase of forebulge relaxation within the study area. Meanwhile the estuary in the Zeebrugge Valley has transformed into a shallow seaway surrounded by large intertidal to supratidal islands.

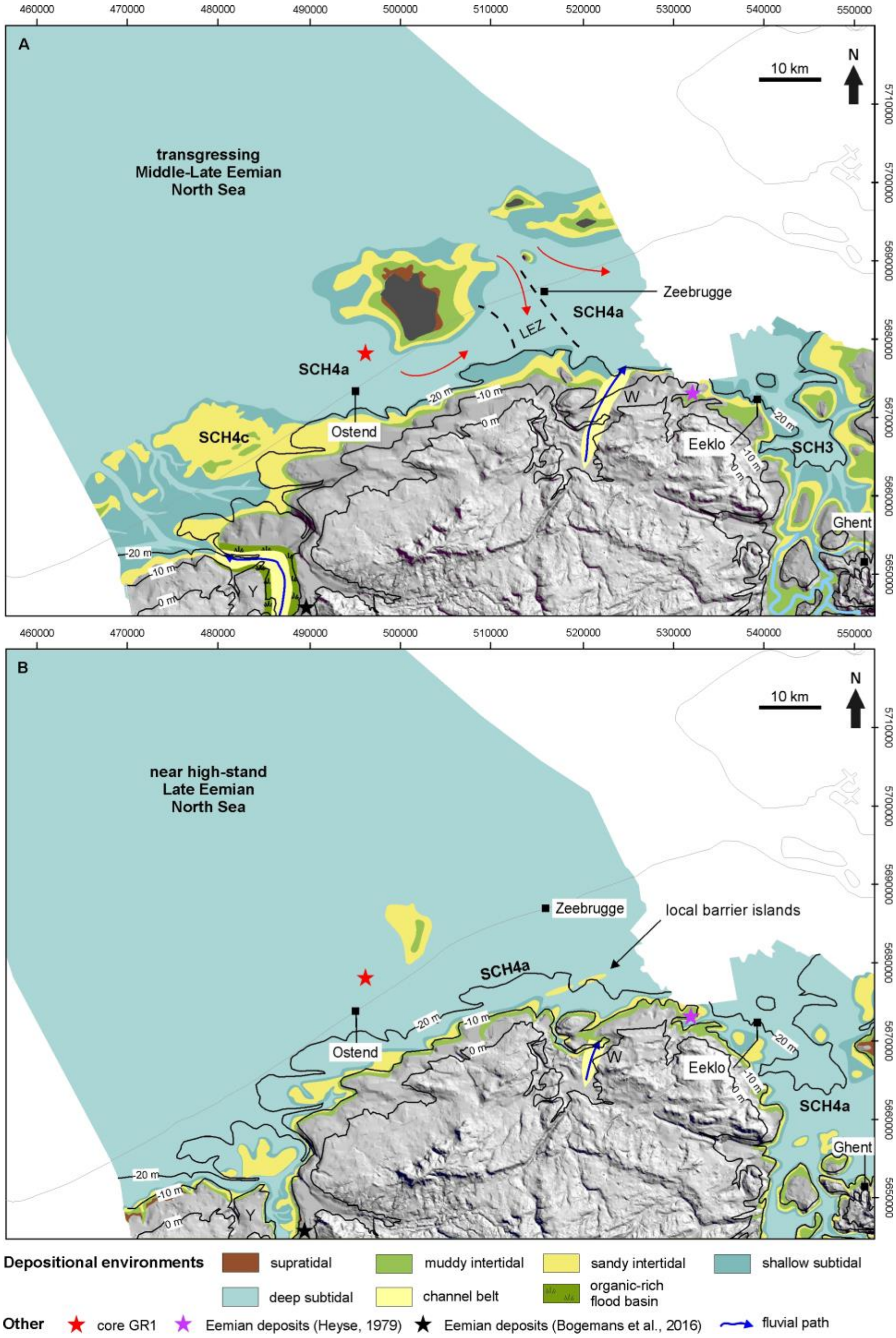


Figure 5.16 Local scenarios for the Late-Middle (MIS 5e, PAZ E5–E6: A) and Late Eemian (MIS 5e, PAZ E6: B). Scenario A shows the landscape at eustatic sea level at -15 m and scenario B at -10 m. The reader is referred to Figure 5.1 for abbreviations used.

Tidal forces within the palaeo-Scheldt Valley have subsided to a more stable level allowing for valley-wide deposition of the marine unit SCH4a (Figure 5.4 A–C). The presence of

Scrobicularia and *Hydrobia* shells indicate active erosion of intertidal flats along the edges of the valley (unit SCH4c). At the same time the estuary has transgressed almost 50 km upstream, near Ghent, allowing tidal sediments to be deposited this deep inland (De Moor and van de Velde, 1994).

In the study area, sea level during PAZ E6 appears to have reached a point where the entire present-day coastal plain is submerged (Figure 5.16 B). Thin stretches of intertidal flats, located alongside the steeper inclined pre-Quaternary surface, remain present along the entire coastline. At the same time the estuary inland transformed into a tidal bay (Mathys, 2009). During PAZ E6, a period that spans almost 5 ka (Müller, 1974), sea level gradually falls.

5.6.3 Weichselian Early Glacial

During the Late Eemian-Weichselian Early Glacial transition, climate deteriorated and got increasingly wetter (Verbruggen et al., 1991). Combined with a gradual sea-level drop to ca. -40 m (Medina-Elizalde, 2013; Waelbroeck et al., 2002; Figure 5.17 A) this entailed renewed incision up to 15–20 m in the inland Flemish Valley (Gullentops and Wouters, 1996; Heyse, 1979; Heyse and Demoulin, 2018 Tavernier and De Moor, 1974; Verbruggen et al., 1991). Also in the downstream palaeo-Scheldt Valley (Mathys et al., 2009 and this study) this time period caused substantial reworking and removal of much of the former sandy Eemian infill (Figure 5.4). Incisions of similar depth and age are further observed in the Thames Valley (Bridgland, 2000).

The presence of reworked marine shells in unit SCH5 shows that as a response to a sea level lowering these channels started to incise into the Eemian interglacial estuarine-marine infill and reworked or eroded these (Figure 5.4 F–G). In the Ostend Valley this incision mounts up to 15 m, assuming unit SCH4a is deposited up to -20 m LAT, between the WMP and CMP, equal to the incised-valley edge (Figure 5.4 E), and continues downstream beyond the Ostend Valley possibly in to French waters with incisions up to 5 m deep in pre-Quaternary strata (Figure 5.6). This results in a thalweg at ca. 30–35 m below LAT at the central section of the Ostend Valley. Similar incisions up to 15 m deep (comparing the shallowest Eemian marine sediments to the thalweg incisions of unit SCH5) are observed in the downstream section of the thalwegs filled with unit SCH5.

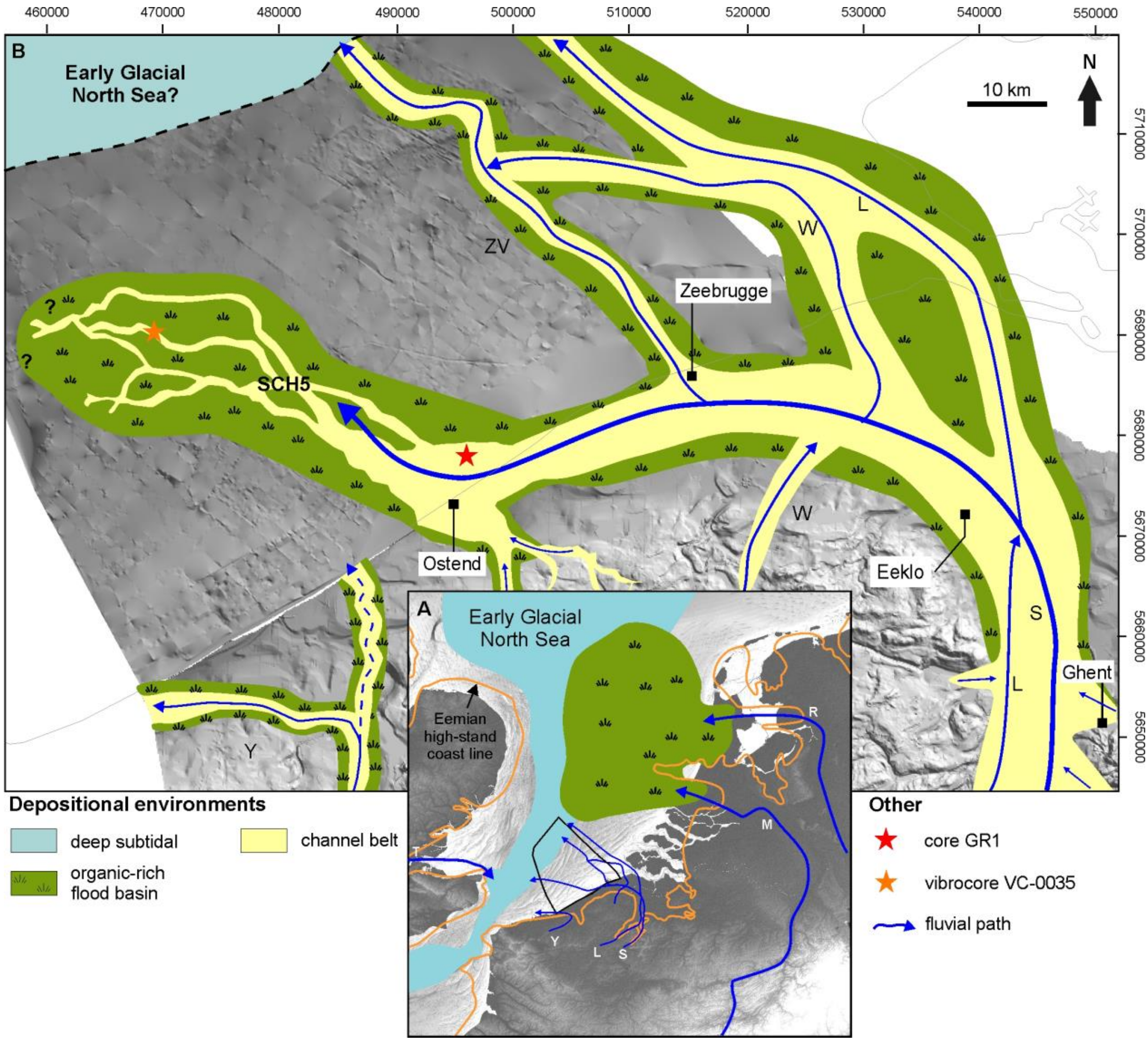


Figure 5.17 Regional (A) and local (B) scenario for the Weichselian Early Glacial (MIS 5d–a) immediately after the Eemian sea level drops. Eemian high-stand coastline: Streif (2004), Mathys (2009) and Hijma et al. (2012). Rhine and Meuse Rivers: Busschers et al. (2008), Hijma et al., (2012), Peeters et al. (2015, 2016); Belgian Rivers: this study. Offshore Lys and Waardamme Rivers: Ebbing and Laban (1996). Yser River: Bogemans et al. (2016). The reader is referred to Figure 5.1 for abbreviations used.

The incisions along the palaeo-Scheldt and Thames Rivers compared to the Rhine and Meuse Rivers show that different settings (e.g. subsidence regime, substrate composition) controlled these river systems and that their connection to the Axial Channel trunk valley may have differed. Indeed, the Rhine-Meuse River system was connected to the Axial Channel trunk valley through a brackish delta system that it had developed as a result of upstream decreased accommodation space during the gradual sea-level fall of the Early Glacial. This delta system is composed of organic clays and peat layers (Hijma et al., 2012; Peeters et al., 2016). The palaeo-Scheldt and Thames Valleys, on the other hand did not develop a delta system during the same sea-level fall. As a result even minor sea-level fluctuations have a direct influence on the incision behaviour of rivers such as the palaeo-Scheldt and Thames whereas the Rhine and Meuse Rivers are only influenced indirectly.

5.6.4 Weichselian Early and Middle Pleniglacial (74–36 ka)

Deteriorating climate conditions of the Early Pleniglacial (74–59 ka: Huijzer and Vandenberghe, 1998) saw the continued degradation of a mixed boreal forest into a tundra landscape with permafrost development in northwest Europe (Huijzer and Vandenberghe, 1998) while in Belgium wet conditions still prevailed (Verbruggen et al., 1991). This caused a lot of hillslope material to be delivered by periglacial slope processes to the lower valleys, where braided rivers were often incapable to cope with such a load (Heyse and Demoulin, 2018). Lateral erosion may have been efficient due to the easily erodible Eemian sandy deposits. The vegetation horizons, clay lenses and silt layers observed in unit SCH6 may originate from channel abandonment and the formation of shallow pools within the braided river system (e.g. De Moor and Heyse, 1974; Heyse, 1979).

During colder intervals the fluvio-hydraulic conditions diminished considerable in drainage systems in northwest Europe (Vandenberghe and Pissart, 1993) focussing river discharge to several channels of the river system (Heyse and De Moor, 1979). The northern SCH6-filled channel directed to the Zeebrugge Valley (Figure 5.3) may be such a drainage-focussing channel. This implies that the downstream Ostend Valley became an ‘inactive’ branch of the palaeo-Scheldt River that is possibly only active during high fluvial discharge such as during spring when greater volumes of meltwater enter the river system. This suggests that by the Middle to Late Pleniglacial transition (ca. 27 ka: Huijzer and Vandenberghe, 1998) the palaeo-Scheldt River most likely drained down the Zeebrugge Valley, and likely also the Waardamme and Lys Rivers, towards the Dover Strait (Figure 5.18).

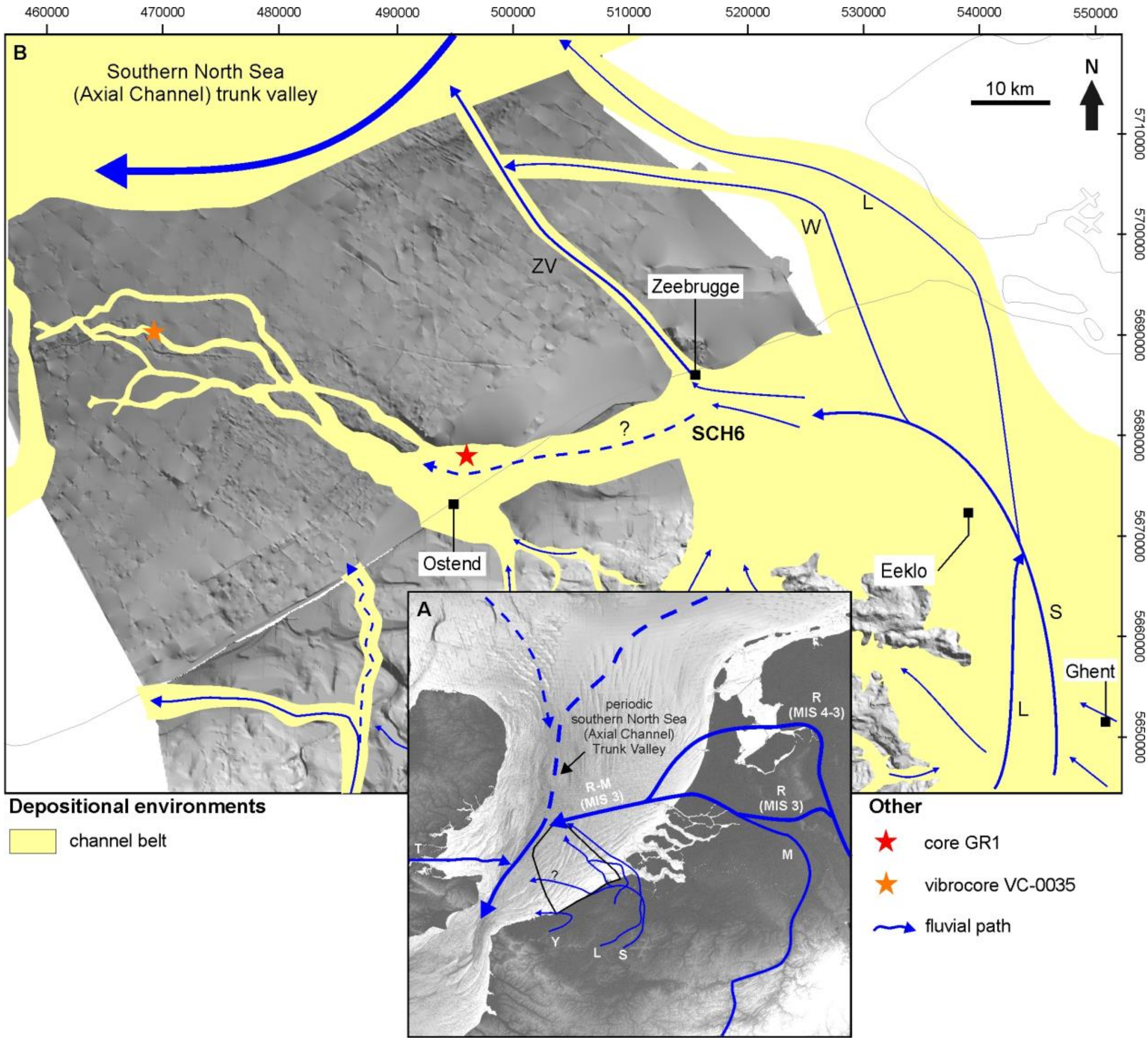


Figure 5.18 Regional (A) and local (B) scenario for the Weichselian Early-Middle Pleniglacial (MIS 4–3). Rhine and Meuse Rivers: Busschers et al. (2008), Hijma et al., (2012), Peeters et al. (2015, 2016); Belgian Rivers: this study. Offshore Lys and Waardamme Rivers: Ebbing and Laban (1996). Yser River: Bogemans et al. (2016). The reader is referred to Figure 5.1 for abbreviations used.

5.6.5 Weichselian late Middle and Late Pleniglacial (ca. 40–15 ka)

At the start of the late Middle Pleniglacial (ca. 37 ka: Huijzer and Vandenberghe, 1998), climate turned very cold and extremely dry (Verbruggen et al., 1991). In the inland Flemish Valley aggradation resulted in ca. 20 m thick sediment sequences (Heyse and De Moor, 1979), while in the downstream palaeo-Scheldt Valley, just south of the Zeebrugge-Coastal Valley bifurcation, ca. 15 m of sediments are formed (see also Depret, 1981). South of this bifurcation shallow channels, filled with braided sediments, continue downstream the Coastal Valley but appear to wedge out (Figure 5.4 B–C). They were either eroded from the geological record by Holocene erosion processes or they are the result of channel abandonment (see above).

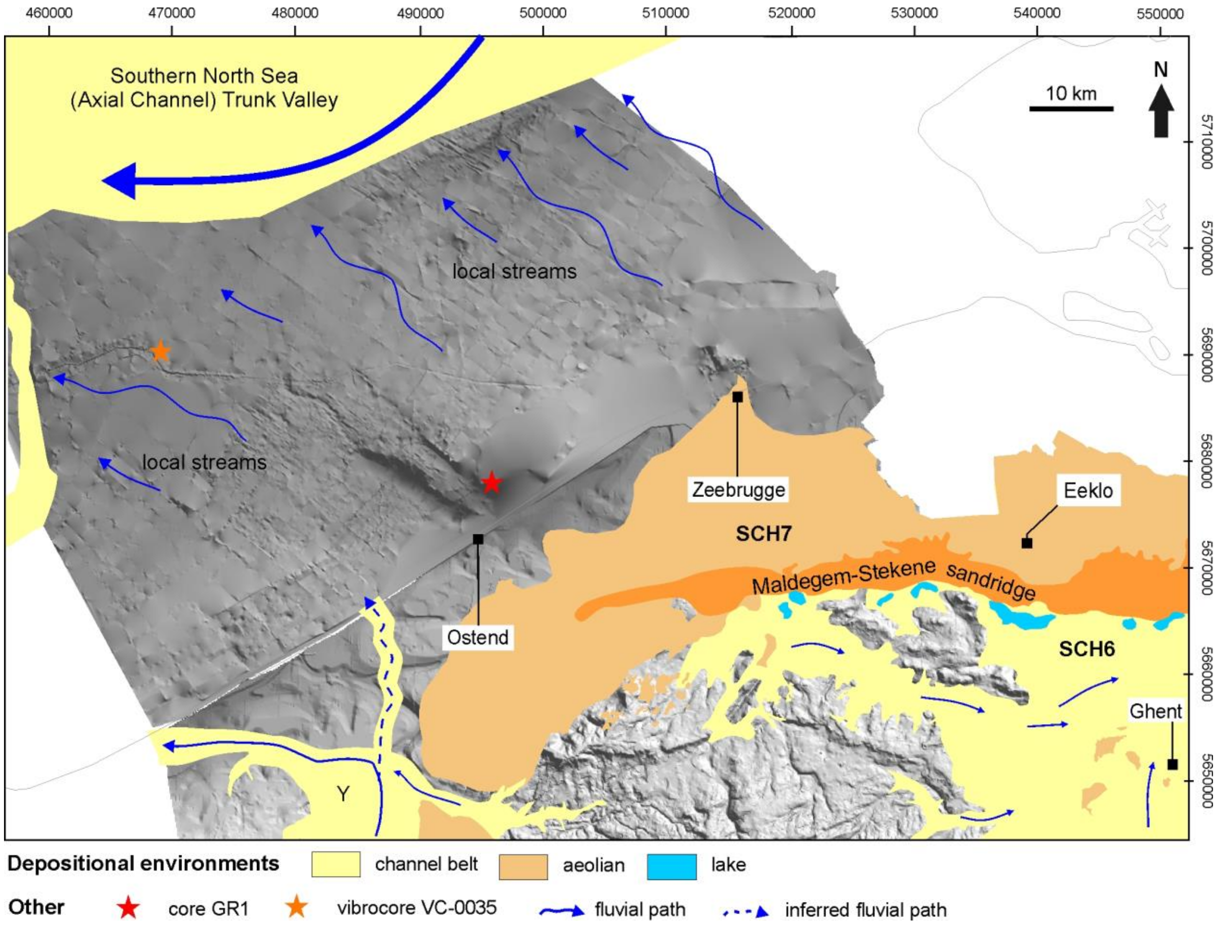


Figure 5.19 Local scenario for the Weichselian late Middle and Late Pleniglacial (MIS 3–2). Yser River: Bogemans et al. (2016). Aeolian deposits and onshore fluvial pathways: Jacobs et al. (2004). Maldegem-Stekene sandridge: Derese et al., 2010 and Crombé et al. (2013). Palaeolakes: Crombé and Robinson (2017). The reader is referred to Figure 5.1 for abbreviations used.

The dry climate and increased aeolian activity that prevailed during the Late Pleniglacial resulted in the formation of a large coversand belt in the upstream area of the palaeo-Scheldt Valley (unit SCH7: Figure 5.4 A; Heyse, 1979). Fixation of the northeast-southwest moving aeolian coversand was partly influenced by humid depressions, locally by vegetation and mainly by damming of the drainage itself, which also caused a rise of the groundwater table in the nearby depressions (Heyse and De Moor, 1979). An east-west oriented coversand ridge, the Maldegem-Stekene sand ridge, formed (Heyse, 1979: Figure 5.1 and Figure 5.19) across the palaeo-Scheldt Valley gradually obstructing its downstream course towards Zeebrugge forcing it to deflect east near Ghent in the direction of Antwerp. This blockade resulted in the formation of small palaeolakes, with a relatively short life span, south of the sand ridge. These intermittent palaeolakes guide the relatively weak palaeo-Scheldt River on the aggradation plain in the present-day direction of Antwerp (Crombé et al., 2013b). OSL-dating across the whole section of the sand ridge show that its formation started around 17.5–15.55 ka ago (Derese et al., 2010) and is the estimated age for the completion of the palaeo-Scheldt redirection towards Antwerp.

5.7 The Eemian sea-level curve in Belgium compared to the Netherlands

In this study seismic mapping, sedimentary geological sampling, palaeoenvironmental investigation and OSL dating were applied on the Eemian estuarine record of the palaeo-Scheldt Valley (Figure 5.9). The palaeoenvironmental analysis (malacology of this study and diatoms of Mathys, 2009) can be applied to develop a relative sea-level curve of the Eemian transgression of the palaeo-Scheldt Valley. The sea-level curve is constructed from four malacological samples from core GR1 (M1, M3–M5; this study) and three additional diatom samples from Mathys (2009) (D1-D3 Supplementary Table 5.2; Figure 5.19). The malacological samples provide information on the water depth during deposition. Water depths ranged around 10 m or even a bit more (e.g. Bogemans et al., 2016; Heyse, 1979) during transgression (M1, M3–M5 Supplementary Table 5.1; Figure 5.9). Only sample M1 was deposited below 10 m water depth.

The observed sea-level rise in Belgium seems to continue longer than in the Netherlands, while the high-stand in Belgium lags that of the Netherlands. Similar to the high-stand sea-level fall would also have lagged in Belgium compared to the Netherlands. These subtle differences may result from the timing and rates of forebulge relaxation (see Cohen et al., 2014). If so, forebulge relaxation appears to have lasted longer to die out in Belgium compared to the Netherlands or if similar rates of relaxation occurred the sea-level curve in Figure 5.11 testifies to a more nearby position of the forebulge crest near Belgium, while the Netherlands are situated at the ice-sheet facing side of the peripheral forebulge. However, tectonic structural differences over short distances, in this case the North Sea Basin and the London Brabant Massif, may also explain the observed differences between both areas. Kiden et al. (2002) assessed that tectonic subsidence between the western Netherlands with respect to Belgium ranges between 0.06 and 0.16 m ka^{-1} since the Eemian period. These differences are likely to have assisted in the observed differences in sea-level rise between Belgium and the Netherlands.

Figure 5.11 also shows the high-stand deposits from Heyse (1979) and Bogemans et al. (2016) (see Figure 5.14 to Figure 5.16 for location). The results from Heyse (1979) and Bogemans et al. (2016) imply that sea-level rise in Belgium reached higher levels than the Netherlands. The same is also inferred from our maximum position of unit SCH4a. This suggests that the study area was located farther away from the forebulge crest than the Rhine delta-estuary. The drowning history of the palaeo-Scheldt however shows that the inundation this far upstream occurred not before PAZ E5. This could mean that the tidal range of 4 m from the model of Baeteman (2008) is an underestimation of the true tidal range caused by tidal conversion upstream or the dated E4a peat sediments are reworked and incorporated in younger E5-E6 deposits (i.e. now they are seen as younger high-stand deposits). This is not exceptional considering the strong tidal forces that entered this area of the palaeo-Scheldt Valley through the Zeebrugge Valley (Figure 5.15 and Figure 5.16 A). This caused major reworking of older deposits that were reworked and redeposited elsewhere within the valley system and were later incorporated in younger sediments.

The inter- to supratidal sediments from Bogemans et al. (2016) similarly show a strong sea-level rise and appears to be in line with that of Heyse (1979). However, only the lower part of this inter- to supratidal sequence was correlated to PAZ E4. It is likely that the upper section dates to PAZ E5 and possibly also E6. These results would then agree with observations from our study area (i.e. a shift to the right on the time axis to a younger high-stand). The overlying Weichselian fluvial channel deposits may have even eroded the upper section of the Eemian drowning sequence in this area. This observation supports the assumption that the E4a peat from Heyse (1979) may have been reworked and incorporated in younger sediments, possibly by the Meetkerke Formation (e.g. unit SCH4c) during E5-E6. Finally, no evidence of a sea-level jump, as postulated by Mathys (2009), has been observed in the present study. This supports the review of Long et al. (2015) for northwest Europe where similar to Belgium no sea-level oscillations could be positively identified.

Future chronological and palaeobiological analysis from the onshore-offshore sections of the palaeo-Scheldt Valley system and its bifurcations will allow us to better understand the formation and demise of this valley system in relation to glacio-isostatic adjustment and sea-level change. This will undoubtedly provide extra information on the coupled relationship between a Late Saalian forebulge between East Anglia and Belgium and its impact on the southern North Sea landscape evolution where apparently small-scale differences in glacio-isostatic adjustment and tectonic movement impacted the landscape in profound ways never considered before.

5.8 Landscape evolution in relation to glacio-isostatic adjustment and spillway erosion of a land bridge

5.8.1 Northwest Europe

Glacial reconstructions of Scandinavia and Britain show that during the Saalian glaciation the FIS and BIIS coalesced over the North Sea Basin (Figure 5.12 A) (Clark et al., 2012; Ehlers et al., 2011; Moreau et al., 2012; Patton et al., 2016; Sejrup et al., 2009; Svendsen et al., 2004). Short-lived ice lobes of this coalesced ice-sheet, up to several hundred meters thick, reached as far as the central part of the Netherlands. Geophysical modelling results (Fjeldskaar and Kjemperud, 1992; Lambeck et al., 2006) show that the region south of the ice mass is located in the peripheral upwarping zone. The exact location and extent of this zone is not known but it is considered here to play a pivotal role in the existence of a Late Saalian proglacial lake in the southern North Sea Basin, a consideration shared by Cohen et al. (2017). Indeed, Busschers et al. (2008) demonstrate that the Meuse River in the Netherlands was situated between the peripheral forebulge to the south and the peripheral depressed region to the north. Consequently, the region between East Anglia and Belgium (perhaps also southwest Netherlands: Gibbard and Cohen, 2015; Hijma et al., 2012) are situated in the immediate vicinity of the peripheral forebulge crest, in the Drenthe-to-Warthe and Warthe-to-Eemian time frames (i.e. 155–120 ka). This position is now corroborated by the Eemian sea-level curve in Figure 5.11 demonstrating that the forebulge crest was indeed located somewhere close to the study area. Compared to the erosive-resistant Elsterian chalk ridge at the Dover Strait this forebulge area has a substrate that consists of relatively easy erodible Paleogene formations (De Batist and Henriet, 1995; Hijma et al., 2012; Le Bot et al., 2003).

Two end-member possibilities arise here regarding the importance of glacio-isostasy to the land bridge ideas: The first end-member implies that the land bridge is not the result of unfinished retrogressive proglacial erosion that started during the Elsterian (ca. 450 ka ago; Gupta et al., 2017), as repeatedly inferred in Toucanne et al. (2010), Hijma et al. (2012), Gibbard and Cohen (2015), Cohen et al. (2017) and Gupta et al. (2017), and that the land bridge may be the direct result of glacio-isostatic upwarping along the forebulge; The second end-member implies that a former land bridge was still present in the landscape as a result of retrogressive erosion (from the former Elsterian land bridge) that was further uplifted by glacio-isostatic adjustment. The presence of a former land bridge is supported by evidence of a Lusitanian marine incursion in the Belgian western coastal plain Yser Valley during the Holsteinian interglacial and the absence of proglacial lake sediments near/at the Yser Valley (Bogemans et al., 2016). The same can also be said based on the British Holsteinian to early Saalian interglacial records (Roe et al., 2009). This implies that a former land bridge was still present after the Elsterian land bridge denudation as also inferred on the basis of Late Saalian proglacial river deposition in the Netherlands (Busschers et al., 2008). In other words, both end members do not exclusively explain the formation of a Late Saalian land bridge between East Anglia and the study area, and we therefore suggest that it was likely caused by a combination of both end members.

Evidence on the timing and magnitude of Late Saalian land bridge denudation is provided by Toucanne et al. (2009a) by terrigenous mass accumulation rates reported from the West European Atlantic margin, the shelf-edge of the English Channel (ca. 1.000 km downstream of the study area). These authors invoke proglacial lake spillway valley erosion to levels similar to that of the Elsterian glaciation as a possible explanation for these events. The large accumulation rates at the West European Atlantic margin comply with the predicted result of increased meltwater due to oscillatory ice mass disintegration since the Drenthe deglaciation phase and with the routing of these masses over the BCS, the Strait of Dover and English Channel at large (Cohen et al., 2017; Gupta et al., 2007, 2017; Hijma et al., 2012; Toucanne et al., 2009a, 2010). The major effect in the study area was that the land bridge lowered, being eroded by massive discharge towards the English Channel. Although it remains unclear, how far north of the landbridge the isostatically-depressed region was located, the depth and thickness of the Eemian estuarine infill in combination with the relative sea-level curve of the palaeo-Scheldt Valley suggests that glacio-isostatic adjustment may have been significant within the study area.

5.8.2 Palaeo-Scheldt Valley

The nearby presence of an isostatically upwarped zone between East Anglia and the study area had major impact on local river systems. The exceptional deep channel incisions WSC and ESC in the palaeo-Scheldt Valley, the thick Eemian estuarine sequence, up to 40 m thick, and the limited timeframe in which these are deposited (e.g. palynology and OSL) suggest that glacio-isostatic adjustment is the main reason why the palaeo-Scheldt, Waardamme and Lys Rivers deflected southwest and formed the palaeo-Scheldt Valley system. This assumes that the palaeo-Scheldt Valley formed sometime earlier than previously suggested, i.e. before, and not as a result of, proglacial lake spillway erosion (Mathys, 2009). It remains unclear if the deep erosion channels WSC and ESC in the Ostend Valley were formed by base-level lowering, as a result of proglacial lake spillway erosion, or that their parallel position to the Paleogene strata is a controlling factor and that their westward diversion, starting at their bifurcations, creates an angle to the Paleogene clays influences incisional depth. It is sure however, that the deposition of fluvial unit SCH1 was well out of reach of tidal forces. From this we interpret that fluvial processes, somewhere between the Late Saalian and the Early Eemian, formed these channels and that the setting may have controlled their incisional depths.

During the following rapid Eemian sea-level rise glacio-isostatic adjustment in the southern North Sea Basin was still an on-going process (Long et al., 2015; Peeters et al., 2016). This is demonstrated by palaeomagnetic investigations and OSL dating of Eemian sediments in the Rhine delta/estuary suggesting sea-level lagged several thousand years after global high-stand conditions were met, i.e. during eustatic sea-level fall (Peeters et al., 2016; Sier et al., 2015; Figure 5.19). Peeters et al. (2016) therefore stress that the Eemian sea-level curve for the Netherlands is a *regional* sea level curve for just the southern North Sea. However, as our study area and the Netherlands are located on adjacent parts of the peripheral forebulge, a similar lag in sea level would be expected to be observed (Figure 5.11). Relaxation of the forebulge during the Eemian interglacial should therefore result in a rapid sea-level rise in the palaeo-Scheldt Valley relative to the Rhine delta/estuary but at a slightly later time. Indeed, the sedimentology and PAZ time control of the palaeo-Scheldt Valley (core GR1) show very rapid infilling of the Ostend Valley section with persistent water depths of ca. 10 m for a prolonged period (PAZ E3–E4: Figure 5.13 and Figure 5.14). It wasn’t until PAZ E5–E6 (Figure 5.16 and Figure 5.17) that water depths slightly decreased suggesting that forebulge collapse was almost completed. This is indeed later than the observed E4–E5 high-stand position observed in the Netherlands (Peeters et al., 2016; Zagwijn, 1983). The Eemian sea-level curve constructed in Figure 5.11 now corroborates these assumptions. Furthermore the indistinct OSL ages are of the same age as those of the Netherlands, meaning that the Eemian correlates to the younger half of MIS 5e, and goes a little bit across the boundary to MIS 5d.

By the Middle Pleniglacial, the Rhine River shifted south, shortening its course and joining the Meuse Valley in the central Netherlands instead of in the Dutch offshore (Busschers et al., 2007; Peeters et al., 2016). This way the Rhine River bypasses the less erodible Early Glacial clay delta deposits. This initiated a renewed direct contact of the main Rhine Valley with the Axial Channel trunk valley, through the Meuse Valley, similar to the palaeo Scheldt and Thames Rivers. Northwest European rivers responded to the Early-Middle Pleniglacial climate deterioration and sea-level lowering first by vertical incision and later experienced lateral erosion with valley aggradation (Huissteden et al., 2001). In Belgium and the Netherlands these incisions were easily obtained by reworking older unconsolidated Pleistocene sandy deposits on a low gradient surface (Huissteden et al., 2001). In England rivers have a steeper gradient and harder bedrock resulting in a gravelly infill and less aggradation as on continental Europe (Huissteden et al., 2001).

During the LGM the European ice-sheet configuration was different compared to that of the Late Saalian. Geophysical modelling results project the peripheral forebulge to be located over the northern part of the Netherlands (Lambeck et al., 2006; Peltier, 2004; Steffen, 2006; Vink et al., 2007) with relative upwarping in the study area amounting between 0 and 5 m higher than northern France and 0–5 m lower than southwest Netherlands. Although undocumented it should be considered if glacio-isostatic adjustment exercised additional external control on the palaeo-Scheldt River system resulting in its deflection towards Antwerp ca. 17.5–15.55 ka ago. This implies that the formation of the Maldegem-Stekene sandridge is not the exclusive reason why the palaeo-Scheldt River deflected at the end of the LGM, which is, up till now, the only considered cause (Crombé et al., 2013; Derese et al., 2010; Heyse, 1979; Heyse and Demoulin, 2018).

5.9 Conclusions

In this study we have integrated core data of the eastern Belgian coastal plain with very-high-resolution seismic reflection and core data of the Inner-Middle BCS with specific focus on the palaeo-Scheldt Valley system. In combination with palaeobiological analyses and OSL dating we have demonstrated that even with limited analysis at crucial locations throughout a buried palaeovalley system a good understanding of landscape evolution can be established. In this study this resulted in a detailed reconstruction of the sedimentary build-up and new advanced insights in to landscape evolution of the palaeo-Scheldt Valley system and the wider southern North Sea during the Late Saalian to Late Pleniglacial period. The development of the palaeo-Scheldt Valley as an incised-valley fill consists of five transitional phases each represented in its palaeogeographic, architectural and sedimentary expression. Finally these transitional phases are the result of interactions between climate, sea-level change and glacio-isostatic adjustment.

A first incision phase of the palaeo-Scheldt Valley started at the end of the Late Saalian and was quickly followed by fine-grained deposition in deeply scoured channels (unit SCH1). This incision is attributed to glacio-isostatic peripheral upwarping possibly of a remnant of an Elsterian land bridge within the study area forming an updoming land bridge between East Anglia and Belgium. This peripheral upwarping resulted in the gradual deflection of the palaeo-Scheldt, Lys and Waardamme Rivers to the study area across the land bridge.

During the following Eemian transgression Late Saalian fluvial sediments were largely removed from the palaeo-Scheldt Valley when it transforms from an incised-valley to an estuary and finally into an marine seaway (units SCH2 to SCH4). The timing of this transgression lagged behind the global eustatic sea-level curve, suggesting that at the onset of the Eemian interglacial the area was experiencing glacio-isostatic forebulge collapse as a result of ice unloading in northwest Europe. Comparing these results with the Eemian transgression type area in the Central Netherlands that is affected by the Rhine show that subtle differences in the time and rates of forebulge collapse across short distances result in significant palaeogeographical changes during transgression after ice-sheet retreat. Within the southern North Sea Basin structural tectonic differences, between the London-Brabant Massif and the North Sea Basin, likely play an important role in different rates of sea-level fluctuations.

The Weichselian Early Glacial is characterised by relatively low-energy conditions throughout the palaeo-Scheldt Valley system, although erosion of the Eemian interglacial high-stand deposits did occur. During this stage the palaeo-Scheldt Valley extended into French waters as a series of shallow channels filled with clayey and organic fine-grained sediments (unit SCH5).

During the Early-Middle Pleniglacial the palaeo-Scheldt Valley transforms into a braided river system (unit SCH6). Sediment deposition stops halfway the Coastal Valley and is assumed to focus through the Zeebrugge Valley. This is attributed to climate change resulting in decreased fluvial activity and channel focussing. No Weichselian Early Pleniglacial palaeo-Scheldt sediments are encountered in the incised-valley fill record.

The final transition of the palaeo-Scheldt system was the deflection eastward near Ghent in the direction of Antwerp that resulted in the abandonment of the palaeo-Scheldt Valley between 17.5 and 15.55 ka. This deflection is mainly attributed to climate change driven aeolian sedimentation (unit SCH7) and decreased fluvial activity possibly facilitated by minor glacio-isostatic adjustments of the study area. The Late Pleistocene sedimentary succession did not experience significant erosion afterwards, resulting in its incorporation in the sedimentary record.

5.10 Acknowledgements

This study was carried out within the framework of the IWT SBO Project 'SeArch' (Archaeological heritage in the North Sea, project nr. 120003) and is funded by IWT grant project nr. 131308. The Netherlands Organisation for Scientific research (NOW-ALW; Grant 834.03.003) supports the Netherlands Centre for Luminescence dating. We would also like to thank the captains and crews of the RV Belgica and RV Simon Stevin for their excellent work during the geophysical campaigns. We would also like to thank Koen De Rycker for the great job he did during the data acquisition. This research greatly benefited from discussions with Freek Busschers (TNO-GSN) and David Garcia Moreno (Ghent University) at various stages of the work. The reviewers are greatly acknowledged for improving this paper with their valuable and constructive comments. The reviewers Kim Cohen and Philip Gibbard are greatly thanked for their constructive comments on the manuscript at the review stage and for advice. We dedicate this research to Jean-Pierre Henriet who passed away at the age of 72 in April 2017.

5.11 Post-scriptum

We would like to inform the reader that a mistake was made in the sea-level curve of De Clercq et al. (2018). The presented sea-level positions of Bogemans et al. (2016) and the top of unit SCH4a of De Clercq et al. (2018) were calculated wrongly. This was corrected for within the presented chapter. The mistakes did not change any of the conclusions made.

5.12 Supplementary information

5.12.1 OSL dating

5.12.1.1 Sampling

Samples were obtained from sediment core GR1 that was stored at the Royal Belgian Institute of Natural Sciences, Brussels, Belgium. For each sample, 400–800 grams (ca. 10–15 cm of inner part of core) of sediment was collected from the split core and stored in opaque black plastic bags. This sample size provided sufficient material for both the dose-rate and the equivalent-dose measurements. To prevent any possible disturbances, samples were preferentially taken from homogenous intervals with clear sedimentological structures. Sampling close to bounding surfaces was avoided to prevent mixture of sediments from different sedimentary units; material adjacent to the core tube was not included in the sample to avoid contamination due to coring. Sample bags were transported to the laboratory of the Netherlands Centre for Luminescence dating (NCL) for further preparation and analysis. At the NCL laboratory, material from each of the bags is split in two parts under orange/amber safelights. One part is prepared for dose rate analysis and the other part for equivalent dose estimation.

5.12.1.2 Dose rate

For dose rate estimation a puck is prepared by mixing sediment with wax. The wax serves to provide a pre-determined geometry and to retain radon in the sample. Activity concentrations of ^{40}K and several nuclides from the U and Th decay chains are measured using high-resolution gamma ray spectrometer. Results are combined with information on burial history, water and organic content history, and the grain size fraction used for luminescence measurements to calculate the effective dose rate. We assumed gradual burial of the samples to the present depth below the surface, and estimated water contents to be 20 ± 3% by weight based on the assumption of full saturation throughout the burial history in combination with known saturation levels of fluvial sandy deposits. There were no signs of disequilibrium in the U decay chain. Resulting dose rate values range from 0.68 ± 0.03 till 1.01 ± 0.04 Gy/ka, which is relatively low but in line with previous results on similar deposits.

5.12.1.3 Palaeodose

The quartz/feldspar grain size fraction of 212–250 μm was purified by sieving and chemical treatment (HCl, H_2O_2 en HF and rinsing with HCl). Density separation at ρ=2.70 and ρ=2.58 was performed to obtain quartz.

Based on a number of tests, suitable measurement parameters were selected for use in the Single Aliquot Regenerative (SAR) dose procedure (Murray and Wintle, 2003). Within the SAR procedure, the natural luminescence signal is first measured, and then compared to laboratory induced luminescence signals to find what radiation dose is needed to induce an equally bright signal. This value provides the palaeodose. The procedure monitors and corrects for luminescence sensitivity changes during the measurement.

The most light-sensitive OSL signal of quartz grains is selected using the ‘Early Background’ approach (Cunningham and Wallinga, 2010). To obtain a good estimate of the burial dose, measurements are repeated on at least 21 aliquots per sample. Each of these aliquots consists of a sample disc containing on average 49 grains, 2-mm diameter sample (Heer et al., 2012). To test the SAR procedure and the selected measurement parameters, a laboratory given dose is retrieved with the adopted procedure; the measured dose agreed reasonably with the given dose (dose recovery ratio 0.93 ± 0.01, n=32). Although this value is comfortably within the 10% limit widely used, we note that the dose recovery ratio is significantly below unity. An investigation of dose recovery as a function of preheat temperature did not provide any better alternative preheat parameters. Potentially, the low dose recovery ratio may be related to the high dose that is recovered in this experiment (~82 Gy, similar to expected natural dose).

Equivalent doses were close to saturation limits of the quartz OSL signal. For each aliquot, OSL data was fitted with a single exponential equation and we recorded whether the signal plotted above the saturation limit of $2D_0$, suggested by Wintle and Murray (2006). The percentage of aliquots above this threshold ranged from 10 to 39%. The sample burial dose is obtained from the single aliquot equivalent dose distribution. We used an unweighted mean, to avoid bias towards lower ages (high estimates tend to be less precise due to curvature of the dose response curve). Single aliquot estimates above $2D_0$ are taken into account, again to avoid bias. To eliminate the influence of outliers, an iterative procedure was used to remove single aliquot equivalent dose estimates that deviate more than 2 standard deviations from the sample mean. This resulted in rejection of 4 to 25% of the aliquots for each sample (see Supplementary Information S4 for radial plots for each sample). Thus obtained sample burial dose estimates were below the saturation threshold of $2D_0$ indicating they are within the datable range. For one sample with suspiciously high percentage of aliquots above the saturation limit, we judge the result to be questionable.

A summary of all luminescence ages is presented in Table 5.3 and information pertaining to the accuracy of published ages can be found in Supplementary Figure 5.1.

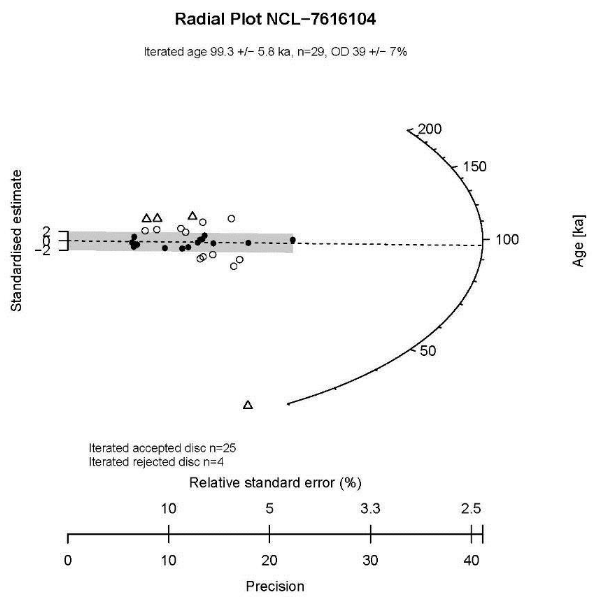
Supplementary Table 5.1 Interpretation of the depositional environment based on mollusc analysis. Analysis performed by Frank Wesselingh from Naturalis Biodiversity Center in the Netherlands.

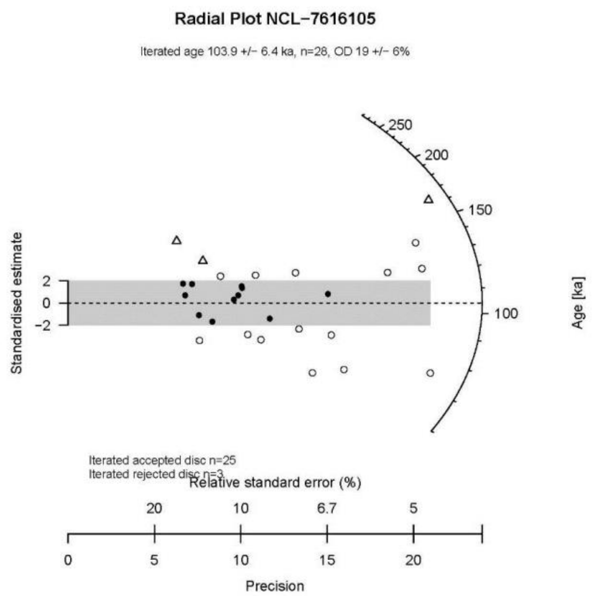
Sample name	Sample depth (m)	Unit	Dominant mollusc assemblage and indicator species	Depositional environment
M1	40.41–40.56	SCH3	*Ostrea edule* *Mimachlamys varia* *Abra ovata* *Bela cf. nebula*	Marine conditions between fair-weather wave base and storm wave base (<10 m water depth)
M2	38.00–38.92	SCH3	No identifiable species	Fluvial transport to very high energy marine environment
M3	36.60–36.76	SCH3	*Ostrea edulis* *Abra ovata* *Mimachlamys varia* *Cerastoderma glaucum*	Between fair-weather wave base and storm wave base with a slightly depressed salinity (>10 m water depth)
M4	30.45–30.55	SCH4	*Mimachlamys varia* *Spisula subtruncata* *Cerastoderma edule* *Donax vittatus* *Abra ovata*	Fluvial transport of above fair-weather wave base
M5	14.47–14.60	SCH4	*Bittium cf. scabrum* *Parvicardium exiguum* *Anomia ephippium* *Spisula subtruncata* *Cerastoderma edule* *Ostrea edulis* *Mimachlamys varia* *Tritia reticulata* *Donax vittatus* *Diplodonta rotundata* *Glycymeris glycymeris* *Trivia monacha (?)* *Macoma obliqua*	Above storm wave base and possibly even above fair weather wave base (ca. 10 m water depth). Fauna is full marine
M6	12.55–12.70	SCH5	*Cerastoderma edule* *Donax vittatus* *Flexopecten flexuosus* *Jujubinus sp.*	Common reworking at shallow depositional depths (possibly fair-weather wave base)

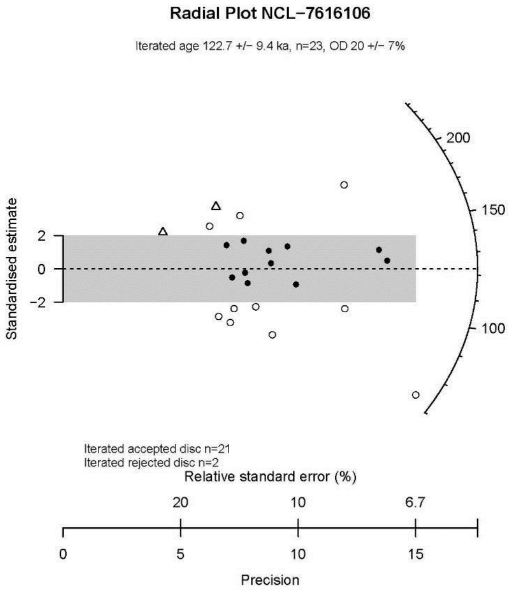
Supplementary Table 5.2 Interpretation of the depositional environment based on diatom analysis (Mathys, 2009). D1-D3: samples used for construction sea-level curve in Figure 5.11.

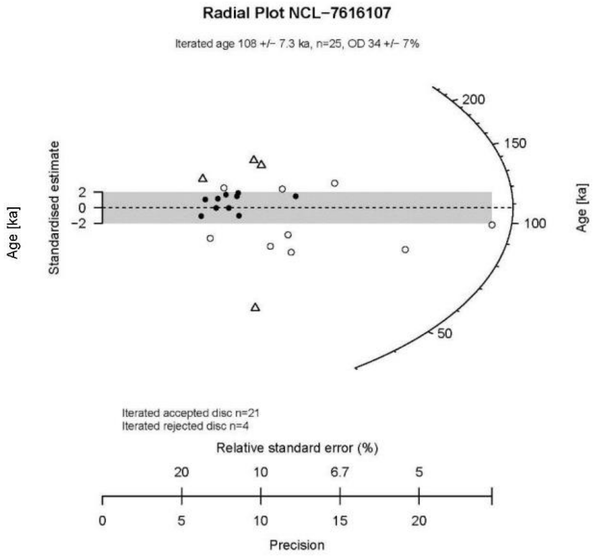
Core	Sample depth (m LAT)	Unit	Diatom species	Depositional environment
GR1	27.60	SCH4a	*Paralia sulcata* *Rhaphoneis amphiceros*	Marine-brackish (D3)
	36.30	SCH3	*Paralia sulcata*	-
	43.10	SCH2	Sparse	Tidal flat (D2)
SB2	18.84	SCH4a	None	-
OSB	11.68	SCH4a	*Rhaphoneis surirella* *Rhaphoneis amphiceros* *Paralia sulcata*	Marine (D1)
	19.30		None	-
UIT	9.50	Ho	None	-
	12.80	SCH4a	None	-

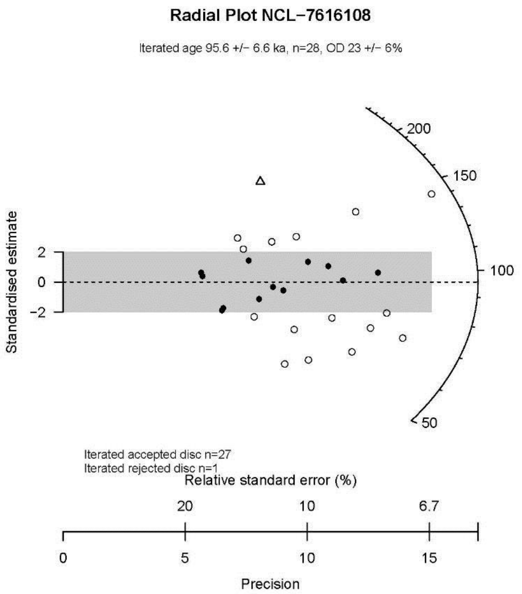
Supplementary Figure 5.1 Age distributions for quartz dating of samples from core GR1**.** The radial plots shown for each sample below single-aliquot luminescence ages (open and filled dots) and the sample age obtained through an iteration model (grey box). The curved y-axis indicates the age estimate, whereas the x-axis reflects the precision of the individual estimates (most well-known points plot on the right-hand side). To construct these graphs, single-aliquot palaeodose estimates were divided by the sample dose rate. Uncertainties in dose rate and systematic uncertainties in palaeodose estimation are not included in the graph. Solid data points fall within the shaded area and agree with the final age estimate. The triangles represent aliquots that are rejected in the iteration (2-σ) process used to clean-up equivalent dose distributions; circles are the accepted aliquots. The robustness of the age obtained is reflected by the percentage of single-aliquot ages within the shaded band, and by the over dispersion percentage (OD; indicated at the top of each graph).

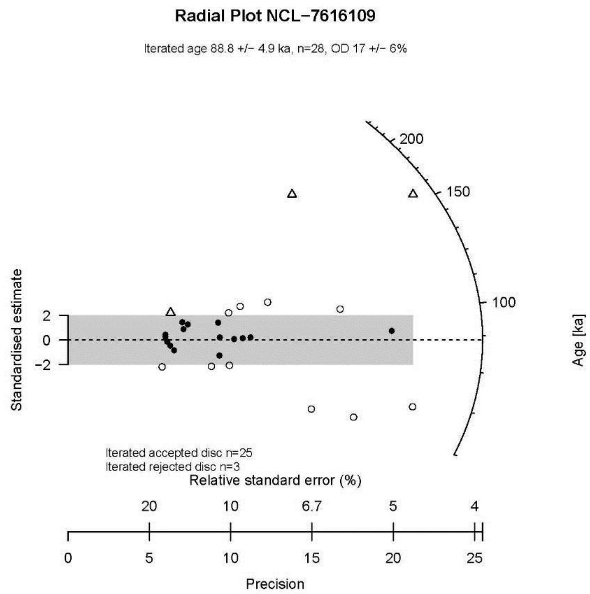





Radial Plot NCL−7616106
Iterated age 122.7 +/− 9.4 ka, n=23, OD 20 +/− 7%
Standardised estimate
2
0
−2
200
150
100
Iterated accepted disc n=21
Iterated rejected disc n=2
Relative standard error (%)
20
10
6.7
0
5
10
15
Precision


Radial Plot NCL−7616107
Iterated age 108 +/− 7.3 ka, n=25, OD 34 +/− 7%
Age [ka]
Standardised estimate
2
0
−2
200
150
100
50
Age [ka]
Iterated accepted disc n=21
Iterated rejected disc n=4
Relative standard error (%)
20
10
6.7
5
0
5
10
15
20
Precision


Radial Plot NCL−7616108
Iterated age 95.6 +/− 6.6 ka, n=28, OD 23 +/− 6%
Standardised estimate
2
0
−2
200
150
100
50
Iterated accepted disc n=27
Iterated rejected disc n=1
Relative standard error (%)
20
10
6.7
0
5
10
15
Precision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Iterated age 88.8 +/− 4.9 ka, n=28, OD 17 +/− 6%
Age [ka]
Standardised estimate
2
0
−2
200
150
100
Age [ka]
Iterated accepted disc n=25
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Relative standard error (%)
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5
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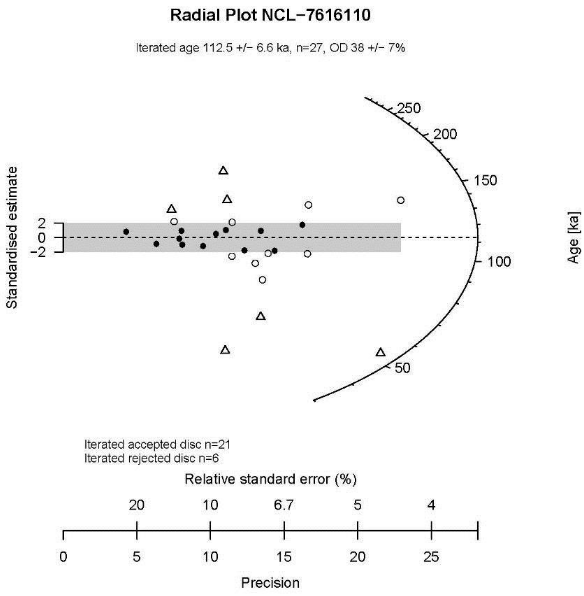
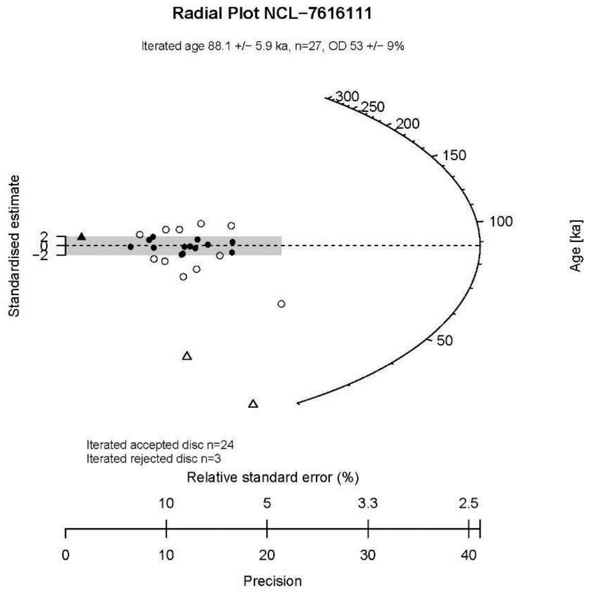
0
5
10
15
20
25
Precision


Radial Plot NCL−7616110
Iterated age 112.5 +/− 6.6 ka, n=27, OD 38 +/− 7%
Standardised estimate
Age [ka]
250
200
150
100
50
2
0
−2
Iterated accepted disc n=21
Iterated rejected disc n=6
Relative standard error (%)
20
10
6.7
5
4
0
5
10
15
20
25
Precision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Standardised estimate
Age [ka]
300
250
200
150
100
50
2
0
−2
Iterated accepted disc n=24
Iterated rejected disc n=3
Relative standard error (%)
10
5
3.3
2.5
0
10
20
30
40
Precision

6 EVIDENCE OF LATE PLEISTOCENE COASTAL ENVIRONMENTS AND GLOFS ON THE MIDDLE-OUTER BELGIAN CONTINENTAL SHELF, SOUTHERN NORTH SEA BASIN: IMPLICATIONS FOR NORTHWEST EUROPEAN LANDSCAPE EVOLUTION

This chapter is based on De Clercq M, García-Moreno D, Missiaen T, Hurtado Zurita O, De Batist M. Submitted. Evidence of Late Pleistocene coastal environments and GLOFs on the Middle-Outer Belgian Continental Shelf, southern North Sea Basin: implications for northwest European landscape evolution. Earth Surface Processes and Landforms.

Abstract

The Outer Belgian Continental Shelf in the southern North Sea is characterised by Last Interglacial sediments that have been truncated by Last Glacial sediments that infill a network of north to south-southwest-oriented, elongated and interconnected palaeodepressions. At the boundary of the Middle-Outer Belgian Continental Shelf lies a large gravel field with a provenance area determined as the Scottish Grampian Highlands and the British East Coast. Despite the fact that these erosional/sedimentary features and gravels have been studied in several previous works, their origin and interrelationship remained largely unknown. Here, we re-evaluate their formation, relationship and transportation mechanism by combining a large set of archived sediment cores with a high-resolution seismic grid, and comparing their combined interpretation with recent palaeoclimatic/palaeogeographic reconstructions of northwest Europe. Our dataset shows evidence for the formation of an Early Last Interglacial shallow marine embayment that was to the south controlled by a Late Penultimate Glaciation Meuse River terrace. These sediments are truncated to their northwest by the palaeodepression network, which we interpret as an erosional feature carved within a floodplain of the Axial Channel trunk valley that experienced one or several episodes of major flooding during the 70 ka and 30–18.7 ka glaciations (MIS 4 and MIS 2). During each glaciation glacio-isostatic adjustment combined with a merged ice-sheet over the North Sea resulted in proglacial lake formation. Subsequent ice-sheet destabilisation released meltwater and icebergs into the lake. This led to drainage of a proglacial lake in the southern North Sea southwards via the Axial Channel into the English Channel. Along this path, about 200 km downstream, an up to 10 m sediment package (locally up to 24 m within palaeodepressions) has been deposited. This sediment package was deposited in a high-energy environment, immediately following extensive erosion of the underlying pre-Quaternary substratum and Last Interglacial deposits.

6.1 Introduction

The Late Pleistocene sedimentary and stratigraphic record of the Middle-Outer BCS is poorly understood, yet it holds some of the key information to fully understand the Late Pleistocene landscape evolution of northwest Europe.

Detailed palaeogeographic reconstructions of the Rhine-Meuse River system in the Netherlands (Busschers et al., 2005, 2007, 2008, Peeters et al., 2015, 2016) and wider southern North Sea Basin (Hijma et al., 2012) hardly ever cross national boundaries. Transnational correlations of stratigraphic units and geomorphological features outside the BCS are not straightforward and have never really been attempted. However, the Outer BCS is most likely located close to or within the area of confluence of two of the largest palaeodrainage systems in northwest Europe, i.e. the Axial Channel trunk valley and the Rhine–Meuse River, during Middle and Late Pleistocene glacial periods (Figure 6.1 A; Toucanne et al., 2010, 2015; García-Moreno, 2017).

Palaeogeographic reconstructions of the Middle-Outer BCS are traditionally based on the assumption that the various Late Pleistocene sedimentary units were deposited *in situ* under marine conditions, during sea-level high-stands (e.g. Mathys, 2009; Van Den Broeke, 1984). This is supported by the observation of infilled ‘scour hollows’ attributed to marine erosion (Le Bot et al., 2003; Liu, 1990; Mathys, 2009). In addition, it is traditionally also assumed that every interglacial transgression of the relatively low-accommodation BCS resulted in the reworking of all continental deposits accumulated outside deeply incised-valley systems during the preceding low-stand period (Cameron et al., 1993; Mathys, 2009; Van Lancker et al., 2007). Most terrestrial deposits found on the BCS have thus been interpreted as reworked by subsequent marine transgressions and submarine erosion (Le Bot et al., 2003; Mathys, 2009).

A recent very-high-resolution DCSM of the BCS pre-Quaternary surface shows that this surface is characterised by strong elongated north to south-southwest-oriented topographical features on the Outer BCS (De Clercq et al., 2016, 2018; Figure 6.1 B). This, together with the recent discovery of cobble- to boulder-sized erratic clasts found throughout the BCS, originating from the Scottish Grampian Highlands and the British east coast (Dusar, 2014; Dusar et al., 2016), suggest that fluvial and/or subaerial flood erosion, possibly GLOFs, and deposition may have significantly contributed to define the geomorphology and sedimentology of the BCS during Pleistocene glaciations. The marine Eemian and/or Holocene reworking hypotheses may thus not be entirely correct.

It is of importance to known when GLOFs occurred during the Late Pleistocene, and what their associated processes are, as they provide information on ice-sheet disintegration and downstream landscape evolution and to understand how strong their impact is on climate and ocean circulation (Carrivick and Tweed, 2013).

In this study, we investigate the palaeogeographic evolution of the Middle-Outer BCS and its link with a proglacial lake that may have existed during the 70 ka and 30–18.7 ka glaciations during the LGM as predicted by several authors (e.g. Carr et al., 2006; Hijma et al., 2012, Hjelstuen et al., 2017; Sejrup et al., 2016). In this effort, we integrate a recently updated DCSM of the top pre-Quaternary surface, derived from the seismic data, with sedimentary descriptions available from archived sediment-cores (see De Clercq et al., 2016; 2018), and (1) evaluate geomorphological features and their possible relationship with GLOFs, (2) identify seismic units using seismic facies analyses, (3) study sediment character of identified seismic facies and seismic units, and (4) discuss a relative chronological framework for the identified processes. Finally, we discuss the regional palaeolandscape evolution for the Eemian-LGM period and put the GOFs in context of the palaeoenvironmental history of the southern North Sea Basin.

6.2 Geological and palaeogeographical setting

6.2.1 Southern North Sea Basin

The rapid Eemian interglacial sea-level rise resulted in the formation of a continental shelf sea similar to the present-day North Sea. This resulted in the deposition of subtidal and 'intertidal coastal-lagoon' deposits from the landward retreating Waardamme and Lys Rivers in the western part of the DCS, near the Belgian-Dutch border, as part of the palaeo-Scheldt Valley system (De Clercq et al., 2018; Hijma et al., 2012) and estuarine deposits in the Thames, Rhine and Meuse River mouths (Bridgland and D'Olier, 1995; Busschers et al., 2005, 2007; Hijma et al., 2012; Peeters et al., 2015, 2016).

The climatic cooling during the following Weichselian Early Glacial resulted in a gradual sea-level lowering (Medina-Elizalde, 2013). This period is characterised by an alternation of cold stadial periods and phases of more temperate interstadial climate, in which global sea level ranged between 0 and -60 m relative to present. This led to the formation of a large prograding delta in the central southern North Sea Basin at the mouth of the Rhine-Meuse River system (Busschers et al., 2007; Hijma et al., 2012; Peeters et al., 2016). Other rivers such as the Thames and the palaeo-Scheldt flowed into an elongated narrow sea positioned in the Axial Channel trunk valley (Hijma et al., 2012; De Clercq et al., 2018).

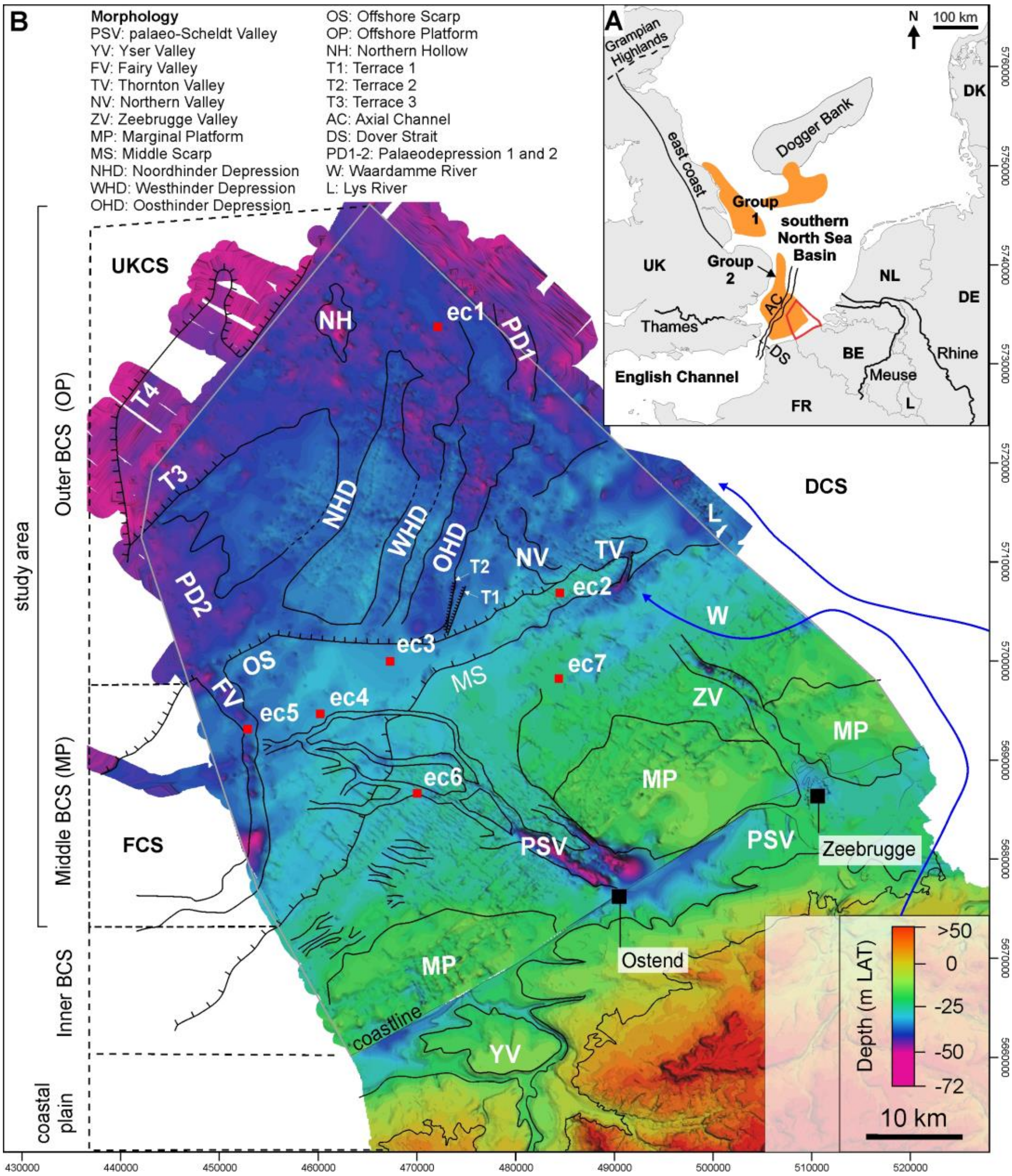


Figure 6.1 A) Northwest Europe with indication of the BCS (red) and the distribution of the two distinct groups of erratic clasts (in orange) encountered in the southern North Sea according to Veenstra (1969). Group 2 is modified following the observations of erratic clasts encountered on the BCS (Dusar, 2014; Dusar et al., 2016). B) Merged map of the top of the pre-Quaternary surface for northwest Belgium and the BCS extending into Dutch, French and English waters (modified from De Clercq et al., 2016). Red squares indicate locations of erratic clasts (ec1 to ec7) derived from the Grampian Highlands and the British east coast (Dusar, 2014; Dusar et al., 2016).

At the onset of the Pleniglacial, climate deterioration continued, resulting in a period of severe cold conditions around 70 ka (Guiter et al., 2003; Helmens, 2014). This induced a significant drop of global sea level to values 100 m below present (Medina-Elizalde, 2013), causing widespread fluvial reworking in the southern North Sea Basin (Huijzer and Vandenberghe, 1998). This fluvial reworking appears to have been intensified by permafrost, which possibly induced a decrease in soil permeability and may have significantly increased the water volume flowing across this area during fluvial peak-discharges (Vandenberghe and Pissart, 1993; Huijzer and Vandenberghe, 1998; Busschers et al., 2007; Westaway and Bridgland, 2010).

Palaeogeographic reconstructions for the North Sea Basin suggest a major expansion of the Fennoscandian and British-Irish ice-sheets (respectively FIS and BIIS) at 70 ka with a possible merging across the North Sea (Carr et al., 2006; Mangerud et al., 2011; Svendsen et al., 2004). In that context, northwest European fluvial systems, previously running north, would have been diverted owing to the ice-sheets blocking their northern routes, forming a major river system that extended from the North European Plain to the Bay of Biscay of the English Channel (Toucanne et al., 2009a). This river system purportedly traversed the southern North Sea, converging with the Thames and Rhine-Meuse River systems in the Axial Channel trunk valley and continuing southward across the Dover Strait (Hijma et al., 2012; García-Moreno, 2017; Patton et al., 2017). This may have resulted in intensive reworking of older sediments deposited during previous high- and low-stands in the DCS and BCS (Hijma et al., 2012). Carr et al. (2006) hypothesize that during the 70 ka glaciation, if the ice-sheets merged, a proglacial lake may have formed to its south and that evidence for this lake is present as the Brown Bank Formation. Other authors suggest that the brackish-water indicators of the Brown Bank Formation is an indication of dropping sea levels, ca. 20 m lower, from the previous Eemian interglacial and that the sediments are therefore of Early Glacial age (Busschers et al., 2007; Peeters et al., 2016).

During the LGM, the FIS and BIIS appear to have merged once again between ca. 30 and 18 ka (Clark et al., 2012; Sejrup et al., 2016). Similarly to the possible 70 ka glaciation, this resulted again in the deviation of the North European Plain and north German Rivers toward the Dover Strait (e.g. Hijma et al., 2012; Patton et al., 2017; Sejrup et al., 2016; Toucanne et al., 2009). Sejrup et al. (2016) and Patton et al. (2017) hypothesised this river system may have formed a proglacial lake in the German-Danish continental shelves, although no direct evidence for this has been presented yet. However, recent evidence for GLOFs in the northern North Sea has been presented and suggest that a proglacial lake was present at that location before the ice-sheets decoupled (Hjelstuen et al., 2017). In any case, both (the rivers formed by the deviated drainage system and the outflow of a proglacial lake) would have flowed southward joining the Rhine–Meuse and Thames Rivers on their way toward the Dover Strait (e.g. Gibbard, 1995).

Sedimentary evidence suggest that this river system may have been subjected to major meltwater injections prior to and during the decoupling of a merged FIS and BIIS due to episodic rapid melting phases (Toucanne et al., 2015). During the time intervals when the FIS and BIIS remained merged, GLOFs were channelled through the Axial Channel trunk valley that connected this lake with the Dover Strait. Evidence for these GLOFs at the Dover Strait have been described by García-Moreno (2017) as a series of scoured geomorphological features.

At various locations in the present southern North Sea far-travelled erratics clasts are encountered. These are spatially distributed in two groups (Figure 6.1 A). The first group belongs to the Middle to Late Pleniglacial Bolders Bank Formation (Stoker et al., 2011; Dove et al., 2017) and is located to the south of the Dogger Bank and along the British East Coast (Veenstra, 1969; Group 1 Figure 6.1 A). The second group of erratic clasts extends as far south as the Dover Strait (Veenstra, 1969) and the BCS (see Dusar, 2014; Dusar et al. 2016) (Figure 6.1 B red boxes). Clast petrology, together with heavy minerals and derived palynomorphs from the first clast group, suggest a provenance from the Grampian Highlands and the Midland Valley of Scotland (northern United Kingdom) and the western margins of the North Sea Basin (Busfield et al., 2015; Clark et al., 2012; Davies et al., 2011; Dove et al., 2017; Larkin et al., 2011; Roberts et al., 2013). The presence of these erratics in those locations have been associated with southward expansions of the North Sea Lobe from the BIIS during successive Pleistocene glaciations (Davies et al., 2011; Dove et al., 2017; Larkin et al., 2011).

The origin of the second group is debated. Some authors proposed that they were deposited by either the Thames River, due to its large concentrations in flint, or the Rhine-Meuse River (Deleu and Van Lancker, 2007; Veenstra, 1969). Other authors have suggested a more local origin, proposing that they were brought there due to marine processes (Mathys, 2009; Veenstra, 1969). Recently, Dusar (2014) and Dusar et al. (2016) analysed a new bulk sample of erratic clasts extracted from the BCS (Table 6.1, Figure 6.1 B: erratic clast locations ec1–7; these samples are stored in the RBINS collection and can be consulted). These clasts range in size from cobbles to boulders (maximum 30 cm in diameter) and have an angular to sub-rounded form. Most material consists of flint fragments; however, some contain abnormally high amounts of sandstone, granite and carbonate, foreign to the BCS (Table 6.1 ec1 and ec3 to ec4). Between clast locations ec3 and ec4 sightings have been reported of even larger clasts ranging in size up to 2–3 m (Van Lancker et al., 2007). In locations, closer to the DCS or the present-day coast (Table 6.1 ec2 and ec7), these foreign materials are absent and replaced by sandstones from the local Paleogene substrate (Dusar et al., 2016). Clast petrology, heavy minerals and derived palynomorphs suggest that the locations containing foreign materials have the same source origin as Group 1. Dusar (2014) and Dusar et al. (2016) hence interpret the BCS erratics as a mixture of far-travelled and local material with an unknown dispersal mechanism and associated them to the Middle Pleniglacial Well Ground Formation.

Table 6.1 Composition summary of the material in the erratic clasts encountered on the BCS (modified from Dusar, 2014 and Dusar et al., 2016).

	ec1	ec2	ec3	ec4	ec5	ec6	ec7
Silex	18	4	393	1275	58	2	283
Sandstone	13		115	244	8	1	
Granite			19	69	2		
Carbonite	7			59	5		
Paleogene		5					386
Total	38	9	527	1647	73	3	669

6.2.2 Belgian Continental Shelf

The pre-Quaternary substratum of the BCS consists of Eocene deposits, mostly of marine origin, slightly tilted (0.5–1°) towards the north-northeast. In the west, the BCS is underlain by homogeneous heavily compacted clays of Ypresian age, while in the east an alternation of sand-silt-clay is present, up to the Belgian-Dutch border (see Le Bot et al., 2003 for a synthesis). From the late Neogene onwards, this substratum was exposed to marine, fluvial, estuarine, and periglacial environmental conditions (De Clercq et al., 2016; Mathys, 2009), resulting in a polygenetic, mostly erosional geomorphological surface (Figure 6.1 B), comprising palaeoriver terraces (e.g. the Middle and Offshore Scarps, T1–T3; Liu et al., 1992), platforms (e.g. the Marginal, Middle and Offshore Platforms; Liu et al., 1992) and numerous palaeovalleys (e.g. the Ostend, Zeebrugge, Waardamme, Lys, Thornton and Northern Valleys and the Ostend Valley Outflow, all attributed to the drainage system of the palaeo-Scheldt River: Liu et al., 1992; De Clercq et al, 2018).

The geomorphology of the pre-Quaternary surface of the BCS was shaped by multiple erosional episodes from the Late Saalian onwards (De Clercq et al., 2018), when the southern North Sea Basin experienced strong geomorphological alterations imposed by the merging FIS and BIIS (Cohen et al., 2014; Ehlers et al., 2011; Gibbard and Lewin, 2016). One of these alterations was the complete deflection of the palaeo-Scheldt River system from the Netherlands to the BCS as a result of glacio-isostatic upwarping (De Clercq et al., 2018). The final breaching of a land bridge, located between East Anglia and northwest Belgium, and the subsequent drainage of a proglacial lake in the southern North Sea, resulted in net vertical incision of fluvial networks in the southern North Sea Basin (Busschers et al., 2008; De Clercq et al., 2018; Toucanne et al., 2009a; 2015). Indeed, the Axial Channel trunk valley (Figure 6.1 A) became the primary drainage system during marine low-stands for the whole southern North Sea Basin and the adjacent European mainland.

The Rhine-Meuse River had a significant impact on the geomorphology and sedimentology of our study area, the Middle-Outer BCS (Busschers et al., 2007, 2008; Hijma et al., 2012; Peeters et al., 2015, 2016). On the DCS the Rhine-Meuse River eroded large parts of older Pleistocene deposits. Whereas in the BCS significant parts of the pre-Quaternary surface were removed (Hijma et al., 2012; Mathys, 2009). Mathys (2009) and De Clercq et al. (2018) suggested that during the Late Saalian Glacial Maximum this river system carved an escarpment or river terrace in the pre-Quaternary surface of the BCS, extending into the chalk bedrock of the FCS down to the Dover Strait, i.e. the so-called Offshore Scarp (García-Moreno, 2017). This Offshore Scarp has a mean height of 6–8 m (locally reaching up to 14 m near the Belgian-Dutch border: De Clercq et al., 2016).

During the following Eemian interglacial and Weichselian glacial periods, the BCS passed through an infilling episode, which resulted in the deposition of sediments into the Late Saalian Meuse palaeovalley different from those of similar age found in the FCS. There, the area north of the Offshore Scarp is infilled with marine Eemian and lacustrine to fluvial Weichselian sediments (Kirby and Oele, 1975). On the Outer BCS, however, the sediments deposited to the north of that scarp are interpreted as Eemian marine deposits overlain by Holocene marine sediments (Balson, 1992; Le Bot et al., 2003; Mathys, 2009; Van Den Broeke, 1984). The authors supporting the latter interpretation assumed that the Eemian and Holocene transgressions along the flat accommodation-deficient BCS would have resulted in the complete reworking of previous low-stand deposits. Based on that assumption, these authors hypothesised that the various scour hollows observed in the pre-Quaternary surface of the BCS were the result of tidal forces during the Eemian transgression, being later infilled with Eemian marine sediments (Liu, 1990; Liu et al., 1993; Mostaert et al., 1989). This view was also supported by the wide-spread presence of Eemian marine molluscs encountered on the BCS (Van Den Broeke, 1984).

6.3 Data and Methods

6.3.1 Seismic reflection dataset

Over the past 40 years the Renard Centre of Marine Geology (Ghent University) collected a large amount of very-high-resolution sub-bottom 2D single-channel seismic reflection data covering the entire BCS (Figure 6.2). These seismic data were collected using various acoustic sources (in the study area mainly multi-tip sparker), operated at different frequencies, and they penetrate and resolve unconsolidated and consolidated sediments up to ca. 100 m below the seafloor (for a summary see Mathys, 2009 and Zurita Hurtado and Missiaen, 2016). Vertical resolution of the seismic data ranges on average between 0.3 and 0.5 m. A large part of this database consisted of analogue paper data that was converted by Mathys (2009) into a reusable digital format. The most recent (2014–2016) seismic reflection data allowed to improve the depth-converted structure map of the pre-Quaternary surface produced by De Clercq et al. (2016) and extends it beyond the BCS. The current version of this structure map of the pre-Quaternary surface within the BCS was gridded and imaged using a 100 m pixel size. The pre-Quaternary surface outside the BCS was gridded and imaged at 25 m pixel size in order to show areas with no data. No cores were available for the time-to-depth conversion of the seismic reflection data acquired outside the BCS. We therefore applied a constant sound velocity for the seismic unit(s) between the seafloor and the top pre-Quaternary surface of 1.560 ms^{-1}. This lack of cores may have added some distortion to the depth-converted structure map in those areas, but we estimate that this distortion is limited to 0–1 m in areas where the thickness of this sediment package does not exceed 10–15 m, which is the case for most of the area in which we apply a constant velocity.

6.3.2 Bathymetry dataset

The bathymetric data comprise three datasets: one dataset extending over the entire BCS, a second dataset over the DCS, and finally the Emodnet bathymetric dataset covering the North Sea Basin and English Channel. The BCS dataset is a compilation of single-beam and multi-beam swath bathymetry data collected between 1990 and 2015 by the Belgian Flemish Hydrography and FPS Economy Continental Shelf Services (www.vlaamsehydrografie.be). This dataset was provided fully processed and gridded at 20 m bin size. Deltares (www.deltares.nl) provided the DCS dataset. It comprises single-beam and multi-beam bathymetric data available from the DCS until 2013. Deltares furnished this dataset fully processed and gridded at 25 m bin size. The Emodnet dataset (bin size: 230 m) is freely accessible at the European Marine Observation and Data Network (www.emodnet-hydrography.eu).

6.3.3 Sediment core dataset

The interpretation of the marine seismic reflection data is supported by sedimentological core descriptions (Figure 6.2). The comprehensive offshore sediment core database (Kint and Van Lancker, 2016) is the result of a wide range of academic and commercial-industrial studies and consists of 74 cores and 172 vibrocores. Mean depth of the sediment cores is 24.5 m and of the vibrocores is 3.97 m. In total 52 of these sediment cores/vibrocores penetrate the Quaternary cover. Data obtained from this database forms the basis for lithostratigraphic analysis and characterisation of the sedimentary facies.

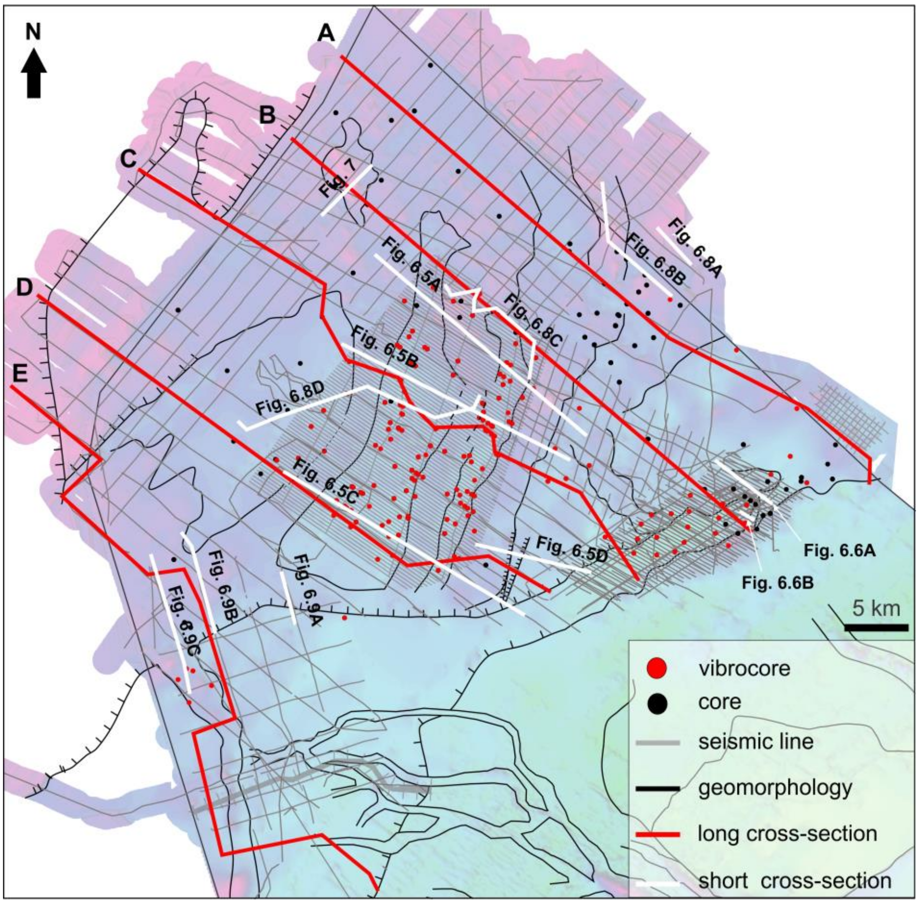


Figure 6.2 Overview of all available seismic reflection and core-vibrocore data used in this study. The long cross-sections are visualised in Figure 6.4.

6.3.4 Methodology

For this study, both seismic and sedimentary unit boundaries were identified and correlated with each other across the Middle-Outer BCS. Characterisation and interpretation of seismic facies and stratigraphy was carried out according to the criteria of Mitchum and Vail (1977) and Mitchum et al. (1977a, 1977b). The goal was to recognise, correlate and characterise the various depositional sequences and their different depositional environments/mechanisms.

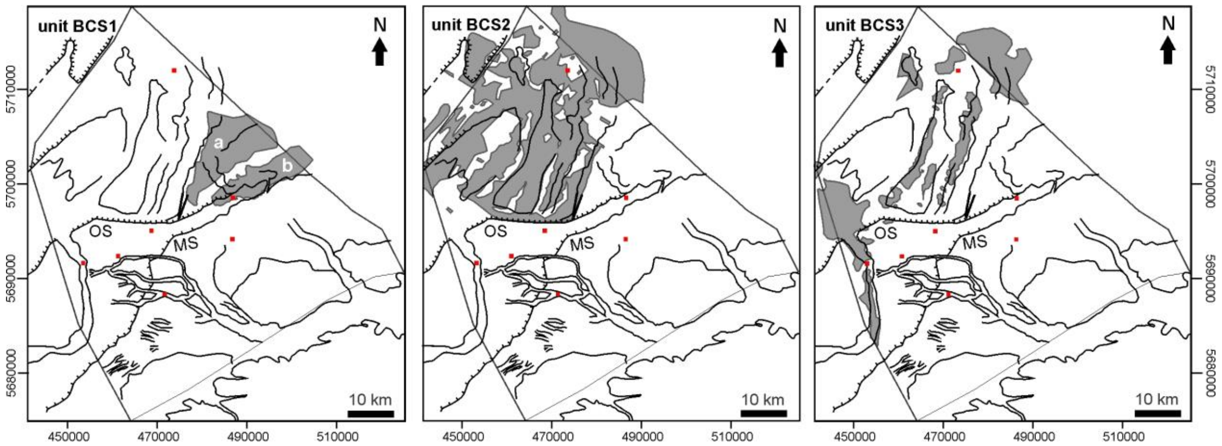


Figure 6.3 Geographical distribution of the identified units within this study. For convenience the erratic clast locations ec1 to ec7 are visualised (red squares).

Sedimentary and faunal descriptions were extracted from the SediLITHO@SEA database (Kint and Van Lancker, 2016) and compared and combined with the seismic interpretation in order to better define the different units and their depositional environments. No dating results are available for this study. When ages are used for the interpretation or discussion, they are based on results of De Clercq et al. (2018) and correlations with events across the southern North Sea Basin and the Bay of Biscay (Toucanne et al., 2015). The geographical distribution of the units is presented in Figure 6.3; while the final interpretations are represented in six northwest–southeast oriented geological cross-sections (Figure 6.4) and a series of palaeogeographical maps (Figure 6.11 to Figure 6.15). A summary of all sedimentary and seismic facies characteristics is presented in Table 6.2. An overview of shell material encountered in each unit is presented in Table 6.3.

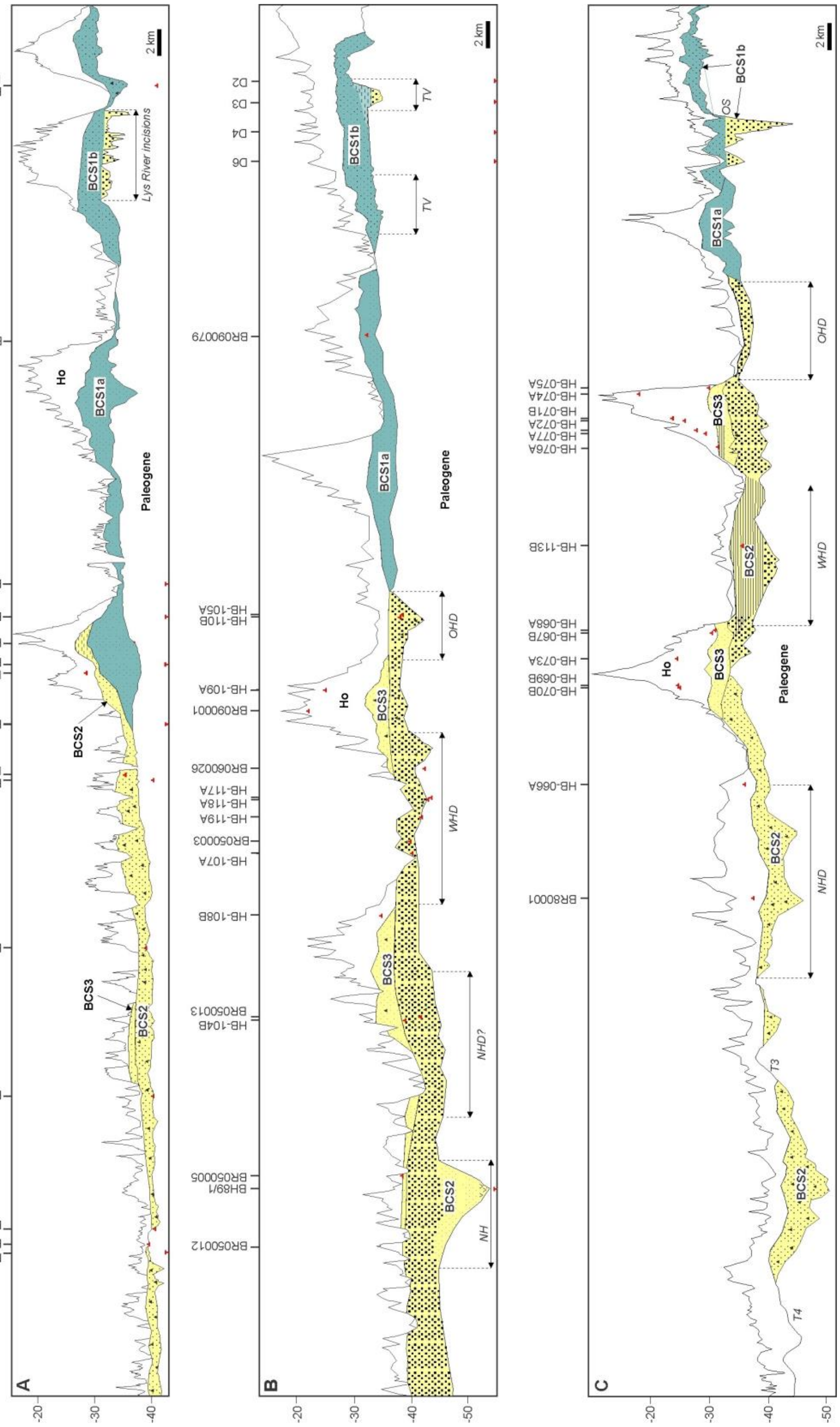


Figure 6.4 Geological cross-sections showing the sedimentary architecture and stratigraphical position of the units identified on the Middle-Outer BCS. Cross-section locations are plotted in Figure 6.2. Horizontal and vertical scales differ between the cross-sections. The colours of the units are matched to the landscape environments in Figure 6.11–13 and Figure 6.15. The reader is referred to Figure 6.1 for used abbreviations. Depths are in meter LAT.

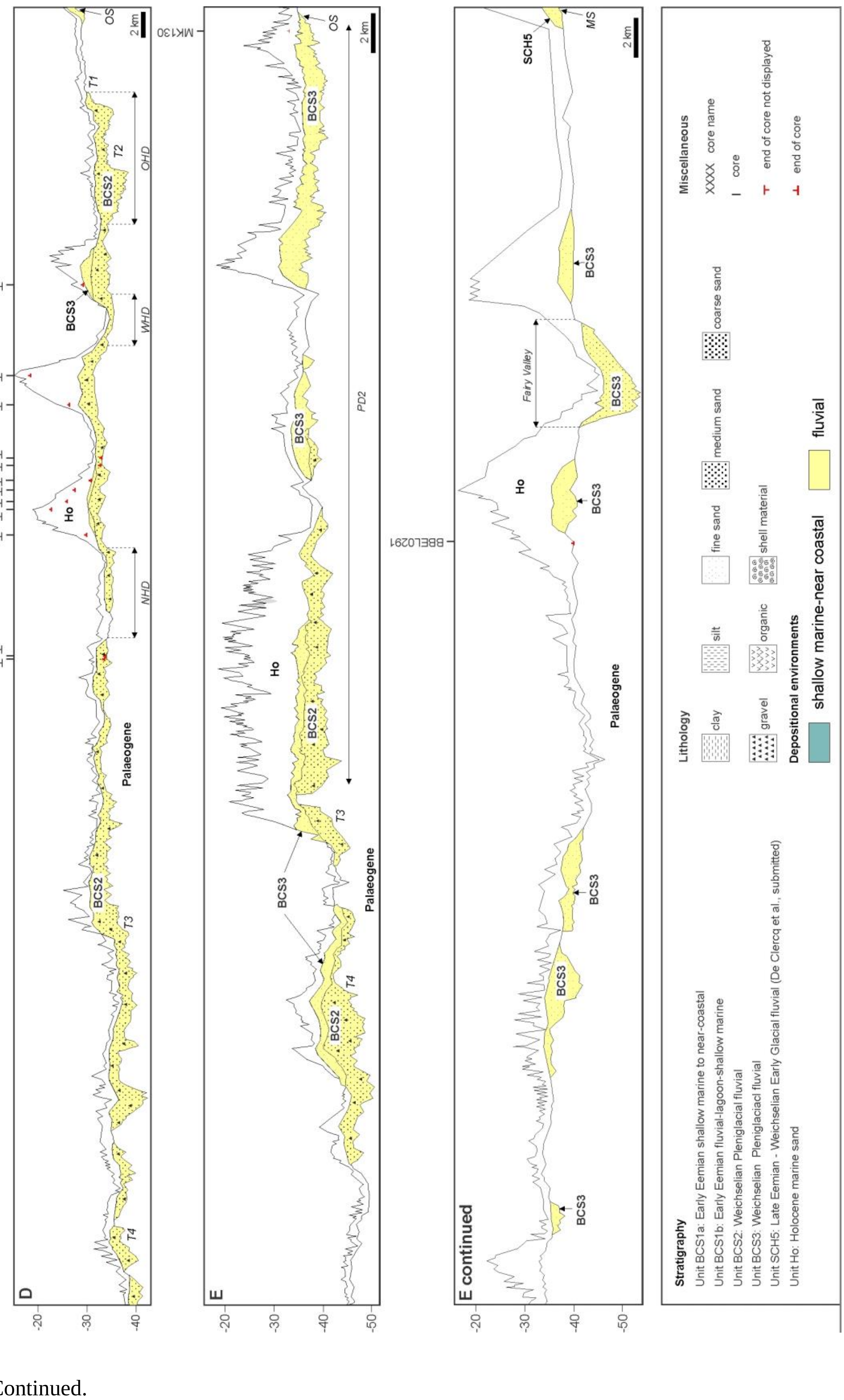


Figure 6.4 Continued.

Table 6.2 Summary of sedimentary characteristics, seismic and environmental interpretation of the seismo-sedimentary units for geological cross-sections A to E in Figure 6.4 and detailed cross-sections in Figure 6.6 to Figure 6.9.

(sub-) Unit	Facies	Sedimentary characteristics	Seismic characteristics (cf. Mitchum et al., 1977)	Depositional environment
BCS1a	-	Very fine (very silty) to medium fine sand, sometimes coarse grained, shell material, fine gravel	Continuous sub-parallel with onlap; discontinuous hummocky-chaotic and oblique	Shallow marine to near-coastal
BCS1b	F1	Coarsening upward fine to medium sand, shell material increases upwards	Sigmoid-oblique (PBDs)	Fluvial
	F2	Slightly sandy to sandy clay, some shell material, $CaCo_3$-rich	Continuous sub-parallel	Central basin estuary
	F3	Fine to medium sand, shell material, some gravel and clay fragments	Continuous hummocky-oblique	Transgressive shallow marine
BCS2	F1	Medium sand with gravel and shell material; at the edges finer-grained	Continuous sigmoid-oblique with onlap; U-shaped upward prograding oblique-parallel	Braided channel belt
	F2	Alternation of very fine- to fine-grained with clay fragments and medium- to very coarse-grained layers rich in mollusc material (marine and fresh) and gravel	Discontinuous sub-parallel with internal channeling; hummocky- to sigmoid-oblique with internal channeling	Braided channel belt
BCS3	-	Alternation of very fine- to fine-grained with clay fragments and medium- to very coarse-grained layers rich in mollusc material (marine and fresh) and gravel; Local stiff clayey sand	Hummocky-chaotic to –oblique; local discontinuous sub-parallel; Internal channel configurations are ubiquitous; sigmoid-oblique, onlap	Braided channel belt

6.4 Results

6.4.1 Geomorphology

De Clercq et al. (2016) reported the presence of three north to south-southwest oriented elongated palaeodepressions, divided by ridges, in the top of the pre-Quaternary surface of the central part of the Outer BCS. Generally these depressions are sub-parallel, 2–7 km wide, 20–30 km long, 5–10 m deep and filled with several meters of Late Pleistocene sediments. The ridges in between these depressions are 2–2.5 km wide and several meters high and are mostly situated beneath the present-day sandbanks (Figure 6.1 and Figure 6.5).

The Oosthinder Depression (OHD) has a relatively stable width between 2.3 and 4.9 km, is ca. 20 km long and reaches depths over -40 m Lowest Astronomical Tide (LAT). The slopes of the depression range from 0.3–0.9° and 0.1–1.2° along its northwest and southeast slopes, respectively. This depression is characterised by two terraces, T1 and T2, at its southeast edge north of the Offshore Scarp (Figure 6.5 D). T1 and T2 are each ca. 3.8 km long and reach a maximum elevation difference of ca. 8.4 m and 10 m, respectively. Towards the south, these escarpments appear to merge, first with each other and then with the Offshore Scarp (Figure 6.1 B).

The Westhinder Depression (WHD) is a channel that narrows from 7 km in the north to ca. 2 km just north of the Offshore Scarp. It has a length of ca. 28 km long and reaches depths up to -48 m LAT. The slopes of the depression are variable and range between 0.2–2.19° and 0.5° respectively along its southeast and northwest slopes.

The Offshore Scarp bounds the OHD and WHD in the south. It appears both channels are diverted here to the west, following the Offshore Scarp where they converge with the Noordhinder Depression (NHD) in what appears to be a deeper incised region of the Outer BCS (i.e. Palaeodepression 2: PD2). The NHD is 2.2–6 km wide and is up to 30 km long with depths up to -43 m LAT. The slopes of the depression range between 0.2 and 0.6° along its southeast slope, and average below 0.35° along its northwest slope.

Alongside these three sub-parallel elongated depressions several other palaeodepressions have been reported or are now newly recognised. The Northern Hollow is a ca. 5 km long to 2.6 km wide north-south incised palaeodepression north of the NHD. Its depths vary between 42.5 and 59 m below LAT and its slopes range between 0.6 and 1.2° (see also Liu et al., 1993; Maenhaut Van Lemberghe, 1992).

Palaeodepression 1 (PD1) is located across the Belgium-Dutch border and has a north-south orientation. The depression is ca. 10 km long although its northern continuation is unclear. Its depths range between 40 and 51 m below LAT and its slopes range between 0.35 and 1.2°.

Palaeodepression 2 (PD2) is located near the Belgian-French border and is bounded by the Offshore Scarp in the south and T3 in the north. The depths of this palaeodepression is approximately 42.5 m below LAT on average and has a gentle slope between 0.1 and 0.2° along its northeast flank. To the southeast it is connected to the elongated palaeodepressions OHD-WHD-NHD while to the south it connects to the Fairy Valley. This valley is ca. 22 km long and extends into the FCS where it was interpreted as an isolated palaeodepression (Liu et al., 1993). The depth of the valley ranges between 37 and 51 m below LAT. In the FCS Palaeodepression 2 likely represents the so-called Quaternary Basin of Kirby and Oele (1975), i.e. the depression north of the Offshore Scarp on the FCS.

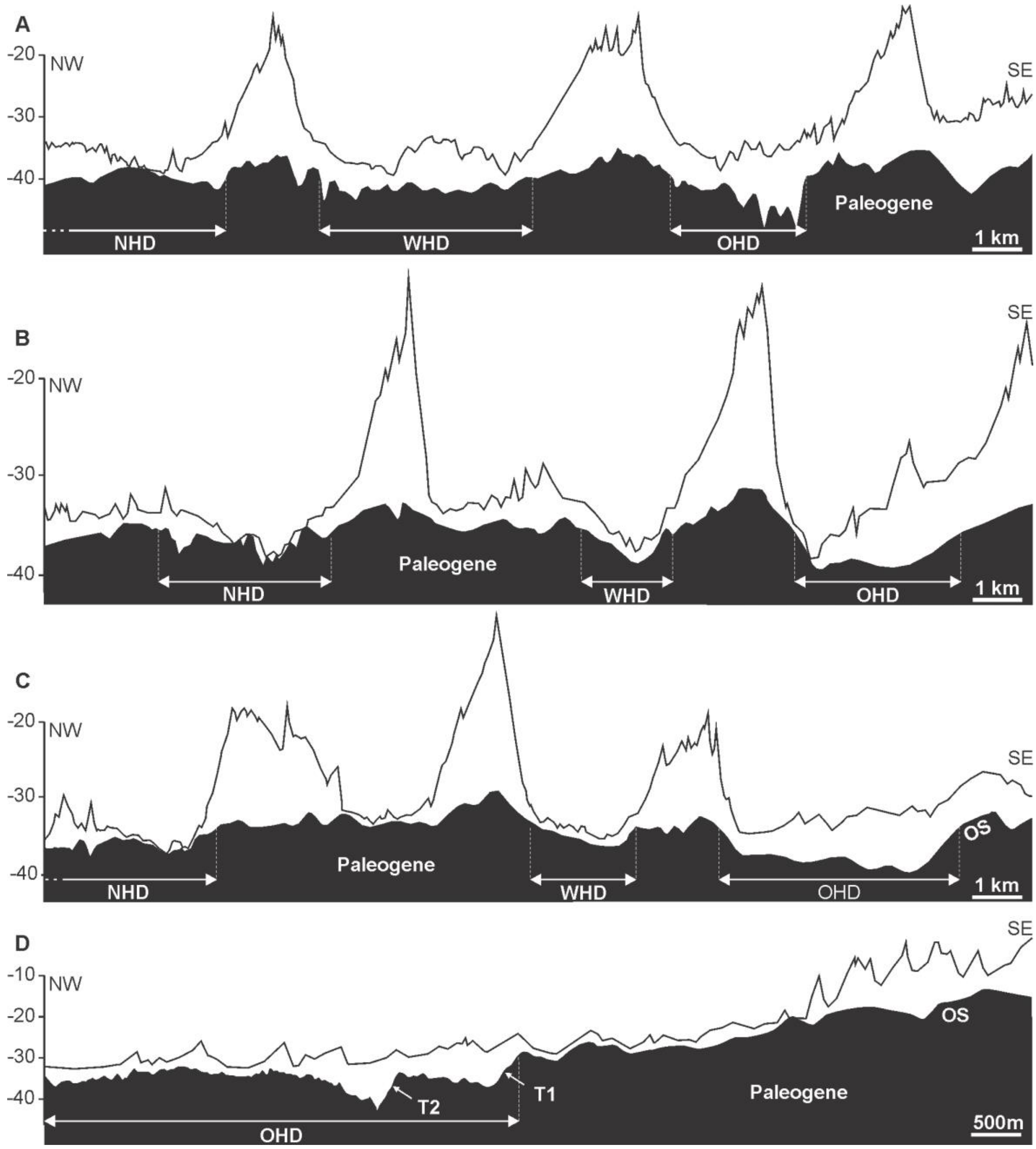


Figure 6.5 Schematic cross-sections of palaeodepressions WHD, OHD and NHD scoured in the pre-Quaternary surface of the Outer BCS. Depths are in meter LAT. Abbreviations mentioned in the figure can be found in Figure 6.1. The locations of the cross-sections are imaged in Figure 6.2.

Escarpments T3 and T4 have recently been described by García-Moreno (2017). Both escarpments run sub-parallel to each other, although T4 has a more complex pattern towards the northeast. T4 appears to be 7–8 m high where it closes in on T3. Indeed, T4 appears to truncate T3. In addition, both scarps appear to meet at the northern section of Palaeodepression 2. Elsewhere, T3 and T4 are ca. 4–10 m high. Liu et al. (1992) already identified part of T3 and T4 and interpreted T4 as the southwest slope of the Axial Channel trunk valley. However, these authors did not associate this scarp with any specific age. More recently, Hijma et al. (2012) modelled the 80–20 ka Rhine-Meuse-Axial Channel palaeovalley assuming escarpment T3 as its southeast slope somewhere during this period.

6.4.2 Stratigraphy

In the study area, three sedimentary units were identified; these are units BCS1, BCS2 and BCS3 (Figure 6.3). Unit BCS1 is divided into two subunits: BCS1a and BCS1b. Subunit BCS1a is located northwest of BCS1b and consists of very fine- to coarse-grained sands of a shallow marine to near coastal origin; subunit BCS1b is located in the southeast corner of the Outer BCS, in the incised Thornton Valley, and consists of very-fine to medium-grained sediments of fluvial, estuarine and shallow marine origin. To its northwest edge unit BCS2 erodes subunit BCS1a. BCS2 fills in buried elongated palaeodepressions scoured in the pre-Quaternary surface, while unit BCS3 is mostly situated on top of BCS2. Both units are attributed a fluvial origin.

6.4.2.1 Subunit BCS1a

Description – Subunit BCS1a is 3–7 m thick and extends over 190 km^2. It is composed of calcium carbonate-rich very fine- to, sometimes very silty, fine-grained sand that can become coarse-grained near the top (core BR090079, Figure 6.4 B). From northwest to southeast (Figure 6.4 A–C), the shell concentration (grit, fragments and complete) increases, appearing mixed with fine gravel. Seismic profiles show a range of seismic facies; from bottom to top, first low-amplitude continuous sub-parallel reflections are present, onlapping the pre-Quaternary surface in a northwest to southeast orientation, overlain by hummocky-chaotic reflections and, finally, by low-amplitude oblique reflections. This unit is truncated by BCS2 along its northwest edge at a depth of ca. -29 m LAT near the Belgian-Dutch border to -35 m LAT further to the southwest (Figure 6.4).

Palaeontology – Core BR090079 shows a dominant Boreal-Lusitanian fauna (Table 6.3; Funder et al., 2002). *Mytilus edulis* and *Macoma balthica* indicate no specific age (Meijer and Cleveringa, 2009); all species are however known to have lived during the Eemian and the Holocene interglacials (Funder et al., 2002).

Table 6.3 Shell assemblage overview of units BCS1 and BCS2. Biogeographic zonations indicate the northern penetration along the European coastlines (see Figure 6.2 in Funder et al., 2002). SA: subarctic, B: Boreal, L: Lusitanian, ST: subtropic.

(sub-) Unit	Name	Biogeographic Zonation	Habitat
BCS1a	*Macoma balthica*	SA, B, L	Shallow marine
	Cerastoderma edule	B, L	Intertidal-shallow marine
	Mytilus edulis	SA, B, L	Intertidal-shallow marine
	Donax vittatus	B	Shoreface
	Barnea candida	B	Shoreface
	Spisula elliptica	B	Shoreface
BCS1b	*Striarca lactea*	ST	Littoral-deep marine
	Diplodonta rotunda	L	Marine
	Turritella communis	B	Shoreface
	Ostrea edulis	B	Estuarine-marine
	Timoclea ovata	B	Marine
	Nucula consentanea	L	Marine
	Littorina littorea	B	Foreshore-tidal zone
	Arctica islandica	SA, B, (L)	Intertidal-shallow marine
	Buccinum undatum	SA	Shoreface
	Cerastoderma edule	B, L	Intertidal
	Astarte montaqui	SA	Shoreface
	Epitonium sp.	SA, B, L	Marine
	Acila cobboldiae	SA	Marine
BCS2	*Mytilus edulis*	SA, B, L	Intertidal-shallow marine
	Chamelea ovata	B, L	Shallow marine
	Macoma balthica	SA, B, L	Shallow marine
	Acila cobboldiae	SA	Shallow-deep marine
	Arctica islandica	SA, B, (L)	Intertidal-shallow marine
	Laevicardium crissum	B, L	Intertidal-deep marine
	Striarca lactea	ST	Littoral-deep marine
	Corbula sp.	B, L	Intertidal-deep marine
	Sphenia binghami	L	Intertidal-deep marine
	Cerastoderma edule	B, L	Intertidal-shallow marine
	Pisidium sp.	-	Freshwater
	Valvata piscinalis	-	Freshwater

Interpretation – Unit BCS1a is interpreted as a shallow marine to near coastal deposit. The dominant marine shell assemblage indicates that the depositional range varies from the tidal zone to the marine shoreface. Furthermore, the species indicate a dominantly Boreal-Lusitanian fauna and only occur during interglacial periods such as the Eemian and the Holocene. According to the landscape models of De Clercq et al. (2018) this area was already inundated during the Early Eemian interglacial. In that situation, the edge of the Offshore Scarp acted as the palaeoshoreline for some time during the Eemian transgression. This suggests that the increasing concentrations of broken shell material and gravel in the direction of this Offshore Scarp palaeoshoreline were due to the wave action (Catuneanu et al., 2011). This relatively homogenous subunit shows a northwest progradation into deeper water indicating possible southwest directed waves combined with energy conditions increasing upwards. This shallow marine setting and upward increasing energy conditions are supported by the seismic facies of BCS1a. Furthermore the sedimentary characteristics of BCS1a are very similar to the Eemian-Weichselian Early Glacial unit MIS5 of Hijma et al. (2012), which it appears to overlap with. Their unit MIS 5 consists of very fine- to medium-grained grey sands with occasionally large amounts of gravel and shell material of a primarily shallow marine origin, including typical Eemian species.

6.4.2.2 Subunit BCS1b

Description – Subunit BCS1b extends into the DCS (Figure 6.3) and has a total areal coverage of at least 120 km² and infills the Thornton Valley. This valley is incised in the pre-Quaternary surface and consists of a channel-like feature (De Clercq et al., 2016, 2018; Mathys, 2009) that enters a shallow basin that is parallel with the palaeoshoreline as depicted by De Clercq et al. (2018; Figure 5.12). This morphological configuration (see Figure 6.1 A) shows a strong resemblance to the lagoon-shaped estuary in a wave-dominated setting following the classification of Reinson (1992). Such lagoon-shaped estuaries have a protective barrier-complex that protects it from the sea, which is not recognised here in the seismic reflection data.

Due to the morphology of the incised-valley the thickness of subunit BSC1b varies greatly. A maximum thickness of ca. 12 m is reached where the channel enters the basin, while more to the north the unit rapidly pinches out. Subunit BCS1b can be divided into four facies, F1 to F4, each reflecting a change in the depositional environment (Figure 6.6 A). Facies F1 constitutes the bottom infill of the palaeovalley and comprises upward coarsening, fine- to medium-grained sand (core D3). The top of this facies consists of medium-grained sand with some gravel and many large shell fragments, possibly indicating higher energy conditions. Facies F2 consists of an up to 8.5 m thick calcium carbonate-rich sandy clay with low concentrations of shell material filling in the central part of the basin. Facies F3 truncates facies F2 in a southeast direction (Figure 6.6 A). It consists of predominantly medium-grained sand with pebbles and shell material that decrease upward. In the north (core D4) the base of facies F3 is finer-grained and clay fragments appear while shell material disappears (core D6). The upper section of facies F3 consists of medium-grained sand with lots of shell material, some gravels and occasionally pebbles. Facies F4 is located just southeast of the basin (core D2), and it is only found in this location. This facies consists of sand–clay alternations with low concentrations of shell material.

In the seismic data, facies F1 is characterised by almost transparent seismic facies with some low-amplitude reflections with a sigmoidal-oblique pattern. Facies F2 consists of low-amplitude to high-amplitude reflections. In this case, the reflections are more or less continuous and slightly undulating. The boundary between facies F2 and F3 consists of a low- to medium-amplitude reflection. Facies F3 is characterised by low-amplitude to medium-amplitude reflections within a more or less continuous hummocky-oblique undulating pattern. Facies F4 has no distinguishable reflectors. The facies F1 channel-fill south of the lagoon-shaped basin consists of medium-amplitude sigmoidal-oblique reflections, resembling lateral accretion at point bar structures, with possible onlap terminations on the pre-Quaternary surface (Figure 6.6 B).

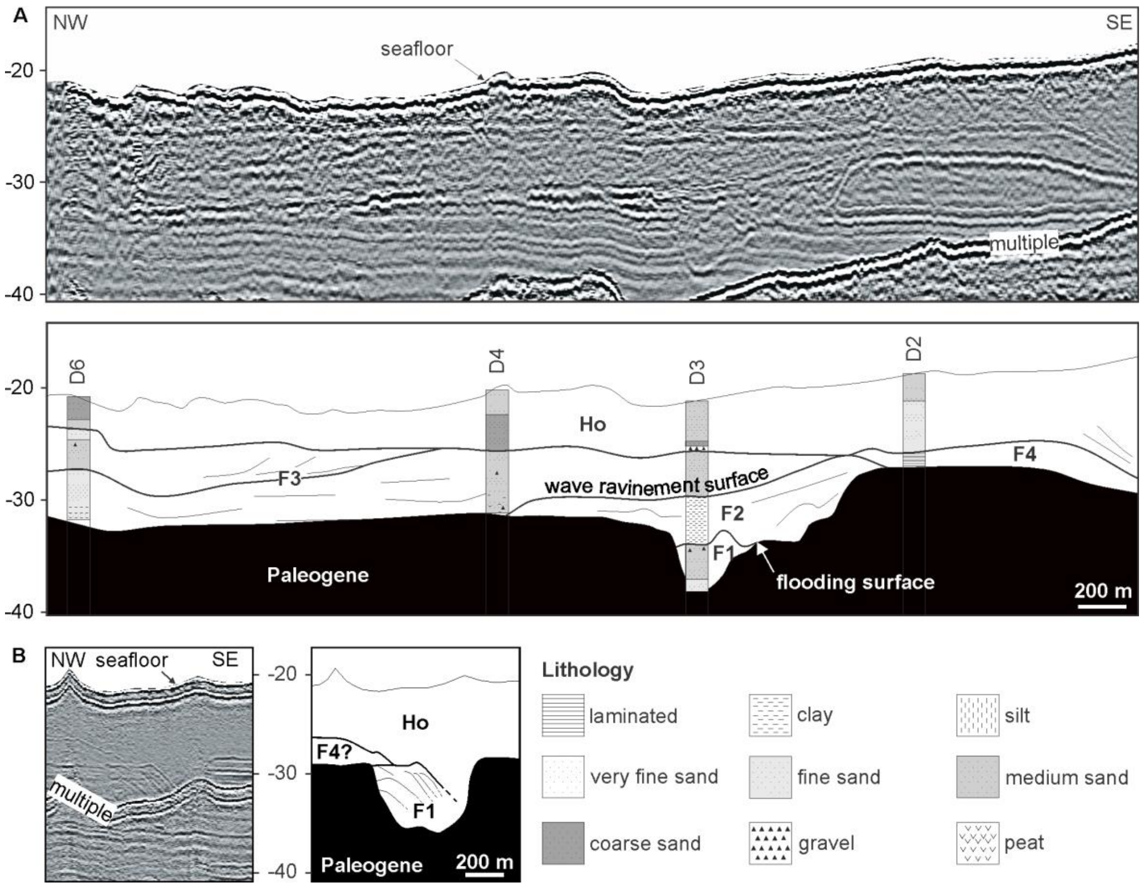


Figure 6.6 Detailed view of subunit BCS1b and its different facies (F1 to F4). Transects A and B (see Figure 6.2 for locations) show the internal stratigraphy of subunit BCS1b at two key locations. A) Cross-sectional view of the central part of the Thornton Valley infill. B) Cross-sectional view of a sediment-filled channel entering the Thornton Valley from the south. Abbreviations used here are referred to in Figure 6.1. All depths are in meters LAT.

Palaeontology – Core BS070187/MK137 (Figure 6.4 A) contains a Boreal-Lusitanian fauna comprising subarctic species (Table 6.3). The fauna represents a relatively shallow marine environment inhabited by species that are found from Late Pleistocene to recent times. Facies F3 is dominated by Boreal-Lusitanian species (e.g. *Acila cobboldiae* and *Crassostrea* sp.) indicating a deep- to shallow-water environment (Gibbard et al., 1991). *Acila cobboldiae,* which had its largest expansion during the cold Early Pleistocene periods, is a species typically found in the near coastal-shallow marine Maassluis Formation and dates back to the Reuverian to Tiglian period (3.4–1.77 Ma; Gibbard et al., 1991).

Interpretation – Subunit BCS1b is interpreted as an Early Eemian wave-dominated estuary, with a lagoon-shaped morphology, filled with fluvial (F1), estuarine (F2), shallow marine (F3) and intertidal (F4) sediments. This morphology and facies succession is typical for an estuarine environment controlled by a microtidal range (Reinsen, 1992). In a complete setting this type of estuary is bounded by a barrier complex parallel to the shoreline. However, no such evidence for a protective barrier was found. This suggests that during transgression the protective barrier may have been eroded during shoreface retreat creating a wave ravinement surface (Boyd et al., 2006; Dalrymple et al., 1992; Figure 6.6 A). If this interpretation is correct facies F2 represents the central basin muddy facies. From this the boundary between facies F1 to F2 represents the flooding surface during transgression (Boyd et al., 2006). The wave ravinement surface is overlain by an upward-coarsening succession passing into transgressive sands (F3). Such a coarsening upward succession is diagnostic for a transgressive wave ravinement surface (Catuneanu, 2006). The margins of wave-dominated estuaries typically contain intertidal and supratidal environments, which, in this case, are represented by facies F4. The Early Eemian age is derived from its transgressive relationship with subunit BCS1a and the palaeogeographic reconstructions of De Clercq et al. (2018). The presence of Early Pleistocene shell material indicates possible reworking of older deposits upstream, as no sediments of this age are encountered on the BCS (see De Clercq et al., 2018). This suggests that these shells are derived from the DCS with the Waardamme River.

6.4.2.3 Unit BCS2

Description – Unit BCS2 is the largest sedimentary unit of the study area and covers ca. 1.500 km². It fills most of the palaeodepressions incised into the pre-Quaternary surface of the Outer BCS. As a result the thickness of BCS2 is highly variable, reaching up to 24 m. This unit is mostly sub-cropping underneath the Holocene sands (Figure 6.4), although it outcrops at the seafloor surface locally. This unit has also been scoured away from several locations, thereby leaving the pre-Quaternary surface exposed, hence its irregular configuration (Figure 6.3).

Based on its sedimentary characteristics unit BCS2 can be subdivided into two distinct facies: A basal facies F1 infilling the palaeodepressions of the pre-Quaternary surface; and the upper facies F2, which is more continuous extending over larger areas underneath Holocene sands, unit BCS3 or is locally exposed at the seafloor.

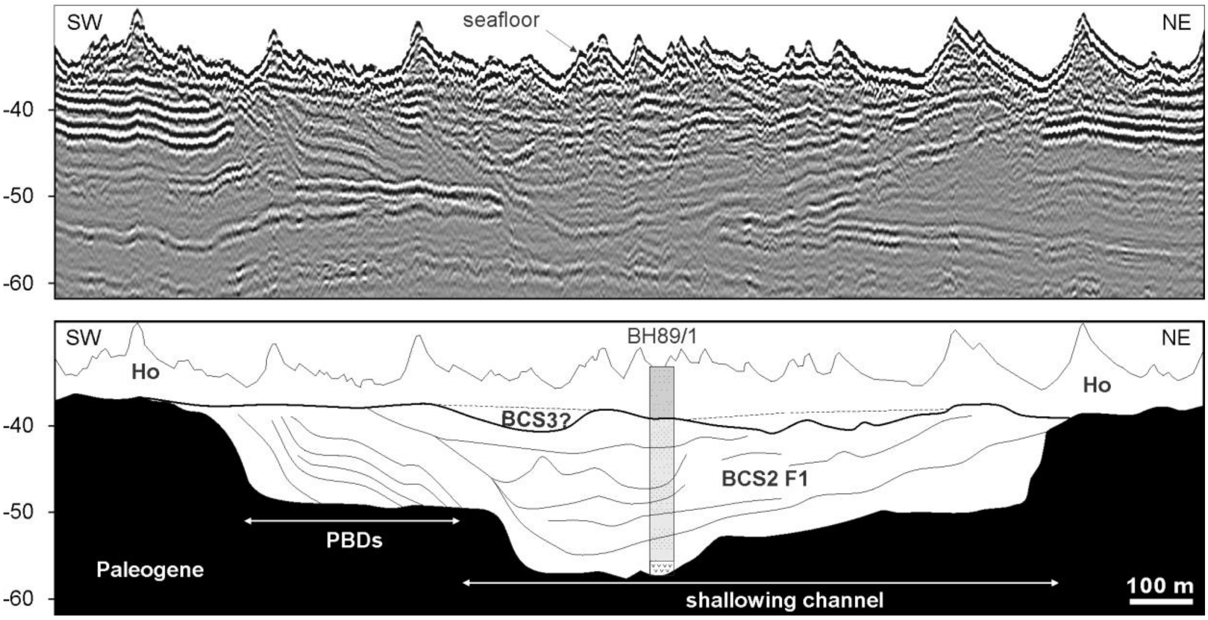


Figure 6.7 Detailed view unit BCS2 facies F1 and its seismic and sedimentary characteristics in the Northern Hollow. Profile location is indicated on Figure 6.2. All depths are in meters LAT.

Facies F1 is identified at only two locations: the Northern Hollow (Figure 6.7) and Palaeodepression 1 (Figure 6.8 B). In the Northern Hollow, facies F1 consists of a calcium carbonate-rich well-sorted fining upward silty fine-grained sand (core BH89/1; for more details see Maenhaut Van Lemberghe, 1992). Gravel concentrations increase upwards while shell material decreases. Peat and wood fragments are abundant and may appear as (sub-) horizontal laminae. At the base of the infill, a one-meter thick gyttja-like peat is preserved. In Palaeodepression 1 facies F1 consists of a brown medium-grained sand mixed with gravel and shell material (core C08; Figure 6.8 B). Near the edges of this palaeodepression the grain size decreases, being replaced by sandy clay and silt (cores BH-102 and F05; Figure 6.8 B). Facies F2 consists of very fine- to fine-grained and medium- to coarse-grained sand layers separated from one another by sharp erosional boundaries. The finer-grained layers consist of fining upward sequences and may have a siltier admixture. The coarser-grained sands contain lots of shell material and gravel. Facies F1 exhibits two distinct seismic-stratigraphic geometries: a continuous sigmoidal to oblique pattern with onlap terminations along its sides, very much like point bar structures; and U-shaped clinoforms prograding upward, showing parallel-oblique geometries (Figure 6.7). Facies F2 also shows two seismic facies with different stratigraphic geometries (Figure 6.8 A). At its base, it shows discontinuous sub-parallel to undulating low- to high-amplitude reflections with low frequencies. Additionally interbedded channels are observed in between these reflections; and in its upper part hummocky- to sigmoidal-oblique bedding consisting of medium- to high-amplitude reflections, occasionally exhibiting shallow channel-like incisions.

Mineralogy – Maenhaut Van Lemberghe (1992) conducted a mineralogical and magnetic analysis on core BH89/1 situated in the centre of the Northern Hollow (Figure 6.7). The analysis showed a dominant Rhine assemblage of garnet, green hornblende and epidote in the lower fine-grained zone, and an increase of the augite concentration in the upper coarser-grained zone.

Palaeontology – Most of the shell material was collected from the northwest area of unit BCS2. It indicates a marine setting, showing a mixture of subarctic, Boreal and Lusitanian species, with the latter being dominant (Table 6.3). The assemblage describes a wide habitat range of environments from shallow intertidal to fully open marine and freshwater. The freshwater species *Pisidium* sp. and *Valvata piscinalis*, encountered in core BR060033 (Figure 6.3 A), appear mixed with marine molluscs that have been identified as Eemian marine (Van Den Broeke, 1984).

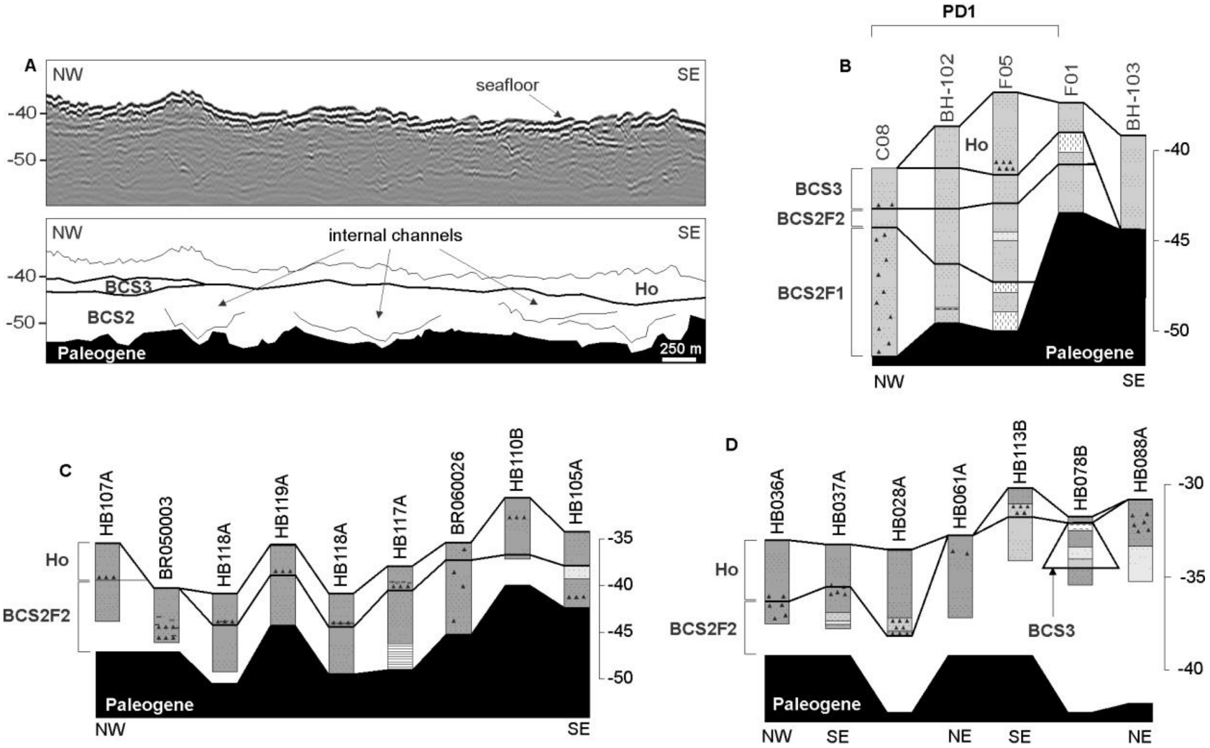


Figure 6.8 Detailed view of a single seismic cross-section (A) and three cross-sections based on cores and vibrocores (B–D). Profile locations are indicated on Figure 6.2. All depths are in meters LAT.

Interpretation – Based on its geometry, cross-sectional grain-size variations and reworked marine and freshwater shells we interpret unit BCS2 as a braided channel belt deposit. A Weichselian age seems plausible given its younger stratigraphic position than the Early Eemian subunit BCS1a, which is eroded by BCS2. We interpret the observed variations in grain size, gravels and shell material within this unit as changes in fluvial discharge. The presence of fine-grained facies alongside the main fluvial streams and vertical variations in fine- and coarse-grained layers with respectively low and high shell material concentrations likely represent episodic flooding events linked to high fluvio-hydraulic conditions. The fluvial origin of unit BCS2 is supported by its seismic facies and the presence of freshwater species. The Rhine mineralogy indicates a northern origin of the discharge where older Rhine-derived sediments have been deposited (Busschers et al., 2008). This is in line with the north to south-southwest orientation of the palaeodepressions. The point bars observed in the palaeodepressions are typical of meandering rivers but are also known to be present in braided river systems (Miall, 1977). The latter is more likely as the morphology of the Northern Hollow and the other palaeodepressions does not show any meandering morphology and the sedimentary characteristics point towards variable energy conditions. The upward migrating channel infill of the Northern Hollow represents multiple phases of erosion and deposition within the channel (Miall, 1996). The shallow internal channels in facies F2 (Figure 6.8 A) are also indicative of a braided river system (Miall, 1996).

6.4.2.4 Unit BCS3

Description – Unit BCS3 is a spatially scattered deposit (Figure 6.3) that is largely stratigraphically positioned above unit BCS2 (Figure 6.4). From the DCS to the FCS, this unit changes from a more or less draped deposit, covering unit BCS2, to a channel infill deposit in the western BCS, where it infills Palaeodepression 2 and the Fairy Valley by truncating BCS2 (Figure 6.4 A–E; Figure 6.9).

Near Palaeodepression 1, unit BCS3 consists of a dominantly silty medium-grained sand to sandy silt and contains fine gravel and shell material (cores C08, BH-102, F05, F01; Figure 6.8 B). Locally, stiff consolidated clayey sand layers occur. In the central Outer BCS, BCS3 is composed of very fine- to fine-grained and medium- to coarse-grained sand layers separated from one another by sharp erosional boundaries. The finer-grained layers are more dominant at the base of this unit, consisting of fining upward sequences with local cross bedding with a possible siltier admixture at their base. Occasionally a clay layer, up to 0.5 m thick, is present interbedded with silty laminae, which coarsen upward into silty and sandy clay. The coarser-grained sands contain much shell material and gravel (up to 4 cm in diameter; core HB078B; Figure 6.8 D). A sharp boundary at the top separates BCS3 from the overlying Holocene sands. Between the central part of the Outer BCS and the Belgian-French border, no core penetrates unit BCS3.

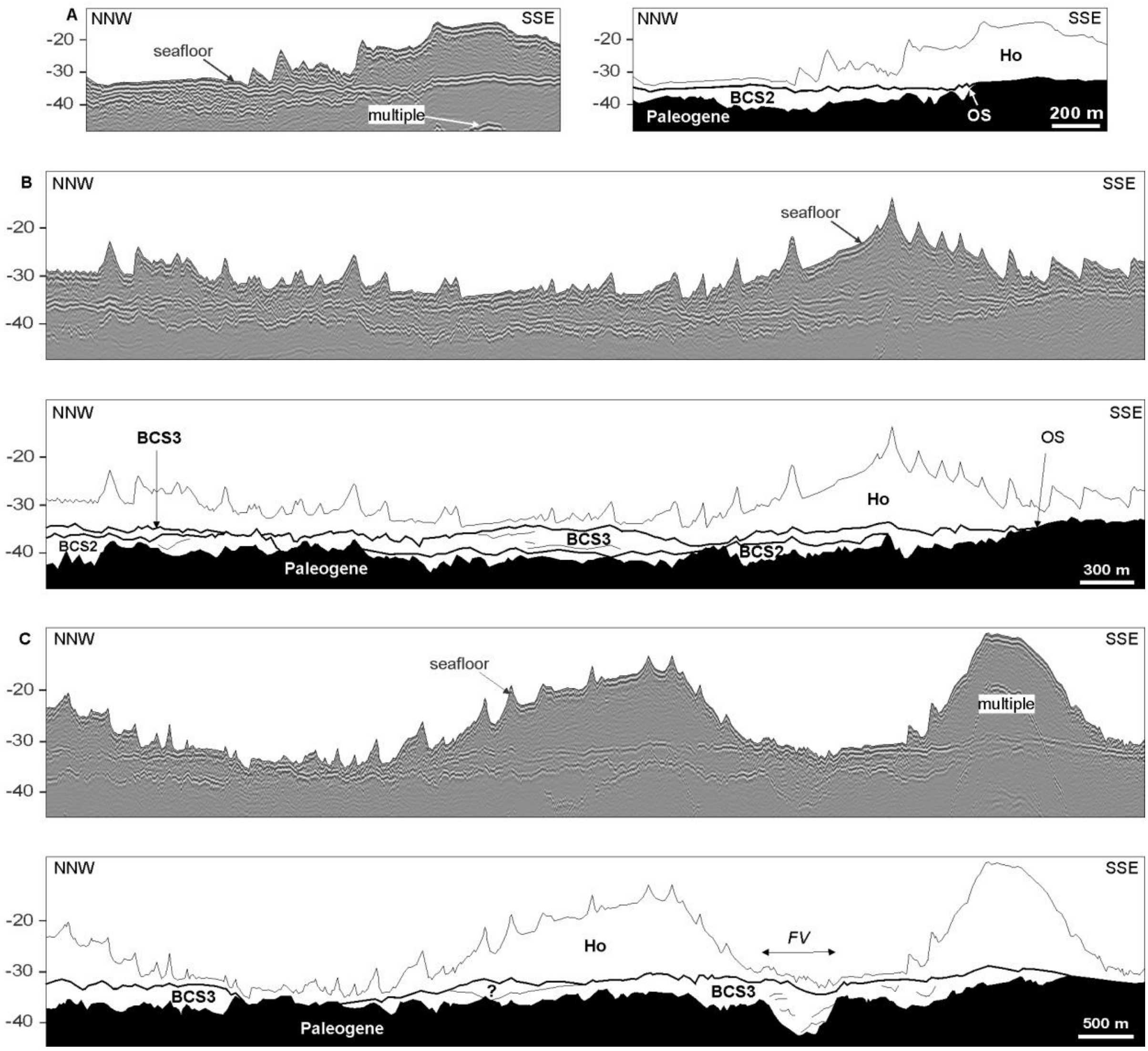


Figure 6.9 Three seismic cross-sections indicating the stratigraphical changes (how unit BCS3 truncates unit BCS2) where the WHD-OHD and the NHD converge into PD2 (A to C)**.** Please note that the horizontal scale between cross-sections differs to provide a clear view for the reader. All depths are in meters present-day LAT.

On the DCS and in Palaeodepression 1, the seismic facies of unit BCS3 is characterised by acoustically almost transparent seismic facies. Some reflections are nevertheless observed indicating zones with hummocky-chaotic to -oblique geometry of predominantly low- to locally high-amplitude reflections, and zones with local discontinuous sub-parallel reflections. In the central Outer BCS unit BCS3 exhibits continuous low frequency, high-amplitude sub-parallel geometries. The boundary between units BCS3 and BCS2 is marked by a series of medium- to high-amplitude reflections. In Palaeodepression 2 the seismic facies is characterised by hummocky-chaotic, laterally continuous sub-parallel reflections. Locally, onlap terminations are observed. In those locations, the unit erodes the lower unit BCS2 (Figure 6.9). In the Fairy Valley, the seismic facies consists of sub-horizontally stratified high-amplitude reflections mostly onlapping onto the pre-Quaternary surface. From north to south, the reflections from the valley infill vary between sub-parallel and sigmoidal-oblique with low to high amplitude. For unit BCS3 no palaeontological information was found.

Interpretation – The presence of cross-sectional grain-size variations and cross-bedding, the erosion of underlying deposits and variable seismic facies suggest unit BCS3 belongs to a braided channel belt. The channel belt is characterised by changing depositional conditions going from the DCS to the FCS. On the DCS very rapid deposition takes place in a setting with highly variable depositional energy resulting in large lithological contrasts. In the central Outer BCS and Palaeodepression 2 deposition takes place in a setting in which the depositional energy changes between relatively stable to variably high. Here unit BCS3 completely erodes BCS2 (Figure 6.9). These intermittent phases of high energy may indicate peak discharges. Unit BCS3 also exhibits multiple interbedded shallow channels similar to those of unit BCS2, suggesting a possible channel belt origin. Based on its relative stratigraphic position, i.e. in between Holocene marine sediments (on top) and unit BCS2, we propose a Weichselian age for unit BCS3.

6.4.3 Erratic clast composition and distribution

The total sampled erratic population of the BCS (ec1–ec7) comprises a diverse suite of rounded to angular far-travelled sedimentary lithologies, alongside subsidiary flat rounded to sub-angular locally derived sedimentary components (Table 6.1; Figure 6.10; Dusar, 2014; Dusar et al., 2016). Flint is the dominant component in locations ec3 to ec5 and ec7, whereas in the remaining locations ec1, ec2 and ec6 flint comprises half of the clast population. The results of Dusar (2014) Dusar et al. (2016) show a remarkable distribution: clast groups ec1 and ec3 to ec5 comprise dominantly of far-travelled clasts, whereas clast groups ec2 and ec7 comprise of locally derived material, originating from the older Paleogene strata. Clast group ec6 has a low clast concentration and is regarded as having a more locally derived affinity.

Figure 6.10 B shows that clast locations ec3, ec4 and ec5 are located within a larger gravel field that is located at the foot and above the Offshore Scarp (Van Lancker et al., 2005, 2007). The gravels range from several centimetres to 3 m in diameter.

Based on these results a tentative boundary can be established between both clast groups that is remarkably similar to the boundary between unit BCS2 and subunit BCS1a. We therefore suggest that the erratic clast group ec1 and ec3 to ec5 are possibly related to the events that resulted in the deposition of unit BCS2 and/or BCS3.

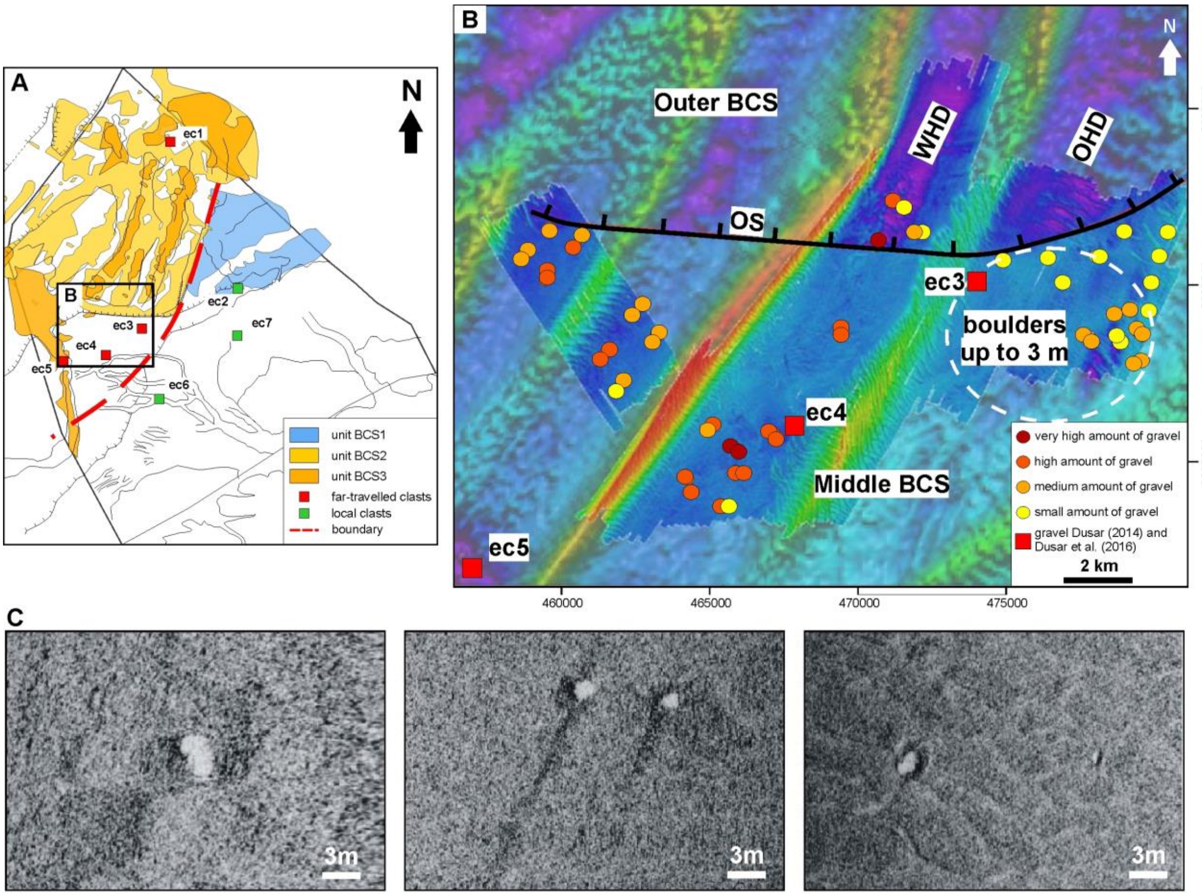


Figure 6.10 A) Erratic clast distribution on the BCS with indication of a division line between local and far-travelled clasts. B) Gravel distribution at the Offshore Scarp. The background shows a detailed multibeam image of the area superposed on less detailed single-beam imagery (modified from Van Lancker et al., 2005, 2007). Erratic gravel locations ec3 to ec5 from Dusar (2014) and Dusar et al. (2016) are also indicated. C) Backscatter images of boulders with a 2–3 m diameter (Van Lancker et al., 2007). See map B for their location.

6.5 Palaeogeographic evolution

6.5.1 Eemian interglacial

At the very end of the Saalian, climate in northwest Europe ameliorated, defining the onset of the Eemian interglacial. This period is characterised by rapidly rising temperatures (e.g. Brewer et al., 2008; Kaspar et al., 2005; Kukla et al., 2002) and sea levels (Dutton and Lambeck, 2012; Kopp et al., 2009; Medina-Elizalde, 2013) and the widespread occurrence of broad-leafed forests (Zagwijn, 1961, 1996). The Eemian North Sea entered the southern North Sea Basin through the Dover Strait and from there extended northward, across lows in the land between Belgium and England. Since the Late Saalian this area is situated on a glacio-isostatically upwarped zone that is now actively subsiding (Figure 6.11 A; see De Clercq et al., 2018).

During the Early Eemian the sea progressively inundated the Outer BCS; when the water reached the topographically higher Offshore Scarp, this possibly became the palaeoshoreline over a significant period of time where prolonged wave action occurred (subunit BCS1a). The waves possibly had a northwest-southeast orientation reaching the coast at an angle. At the same time the palaeo-Scheldt Valley and all its bifurcations, e.g. the Zeebrugge Valley, Waardamme and Lys Rivers, retained their former Late Saalian positions and consolidated their course (Figure 6.11 B). This is supported by the lateral accretion of point bars in subunit BCS1b of the Thornton Valley suggesting a stable river system, probably representing the infilling of a channel on an actively aggrading floodplain. Due to the decreasing slope of the Offshore Scarp from the DCS to the FCS the palaeoshoreline in the east was not able to retreat far inland while to the west, the relatively flat Middle BCS quickly inundated transforming into the palaeo-Scheldt Valley estuary. The steeper coastline in the east, as a result of a ‘hill’ in the pre-Quaternary surface, resulted in the formation of lagoon-shaped wave-dominated estuaries possibly protected by a barrier complex (see also Hijma et al., 2012). Such coastal features can only form in a micro-tidal range, i.e. tidal levels below 2 m (Boyd et al., 2006). This supports the presence of a not yet interconnected southern and northern North Sea at the time, suggesting the Eemian southern North Sea was in fact a shallow bay (Figure 6.11 A).

The rising sea level transgresses the Offshore Scarp palaeoshoreline and forces it to migrate landward. This transgression is observed as coastal onlap in subunit BCS1b as facies F3 that truncates the upper section of the central basin muddy facies F2 (Figure 6.6 A). As mentioned above a micro-tidal estuary is protected from the sea by a barrier complex (Boyd et al., 2006;

Dalrymple, 1992). Seismic facies analysis showed no trace of such a protective barrier complex suggesting it was removed from the geologic record during transgression and shoreface retreat.

As sea level continues to rise the palaeoshoreline continues to migrate landward. Mathys (2009) and De Clercq et al. (2018) show that the Early Eemian palaeo-Scheldt Valley estuary transforms into a tide-dominated estuary during the Middle Eemian with dimensions comparable to the present-day Westerscheldt. This demonstrates that the tidal range within the Eemian North Sea has changed from a micro-tidal range into a meso- to macro-tidal range suggesting the southern and northern North Seas have connected somewhere between the Early and Middle Eemian. When the sea reached a high-stand position the entire BCS is inundated (see De Clercq et al., 2018 for more detail on the Eemian transgression of the BCS).

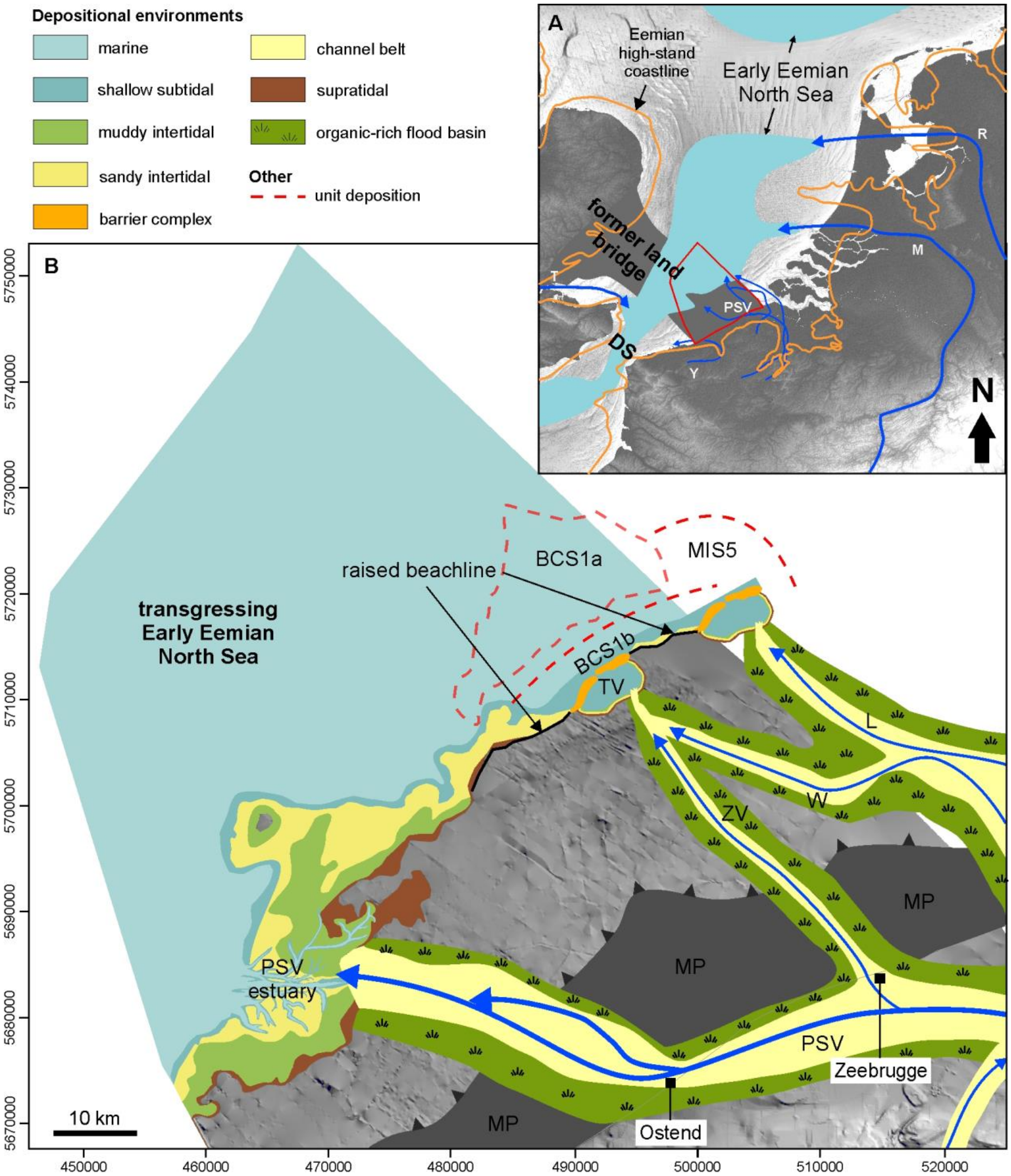


Figure 6.11 Regional (A) and local (B) scenario for the Early Eemian. The regional scenario shows the northwest European landscape during the Eemian high-stand (orange line) as a reference to the local scenario. The local scenario shows a detailed view of the environmental conditions in which units BCS1a and BCS1b were deposited (modified from De Clercq et al., 2018). Rhine and Meuse: Hijma et al. (2012) and Peeters et al. (2015, 2016). Other main river courses are derived from their current flow paths, while their offshore continuations are derived from Hijma et al. (2012). Palaeo Scheldt Valley and its tributaries: De Clercq et al. (2018). The reader is referred to Figure 6.1 B for abbreviations used.

6.5.2 Weichselian Early Glacial

During the Weichselian Early Glacial climate deteriorated and is characterised by an alternation of cold stadial conditions (Herning, Rederstall) and phases of more temperate interstadial climate (Brørup, Odderade; Caspers and Freund, 2001; Helmens, 2014). Within the cold stadial conditions the landscape changes into a treeless grassy-shrub tundra landscape, while during the warmer interstadials birch and pine forests dominate the northwest European landscape (Helmens, 2014). A global compilation of corals that develop at sea level performed by Medina-Elizalde (2013) indicates that sea level probably did not drop below -25 m for the entire period except during the Rederstall stadial when sea level dropped to -60 m.

During the Early Glacial sea-level fall freshwater to brackish water environments from the prograding Rhine-Meuse Delta dominate the southern North Sea (Busschers et al., 2007; Hijma et al., 2012; Laban, 1995; Oele, 1971; Peeters et al., 2016). This westward expansion of the Rhine-Meuse Delta was particularly important during the Brørup interstadial (Peeters et al., 2016) and presumably acted as an important barrier when the former high-stand Eemian North Sea gradually narrows and retreats to lows in the landscape between England and Belgium and the Netherlands (Figure 6.12 A).

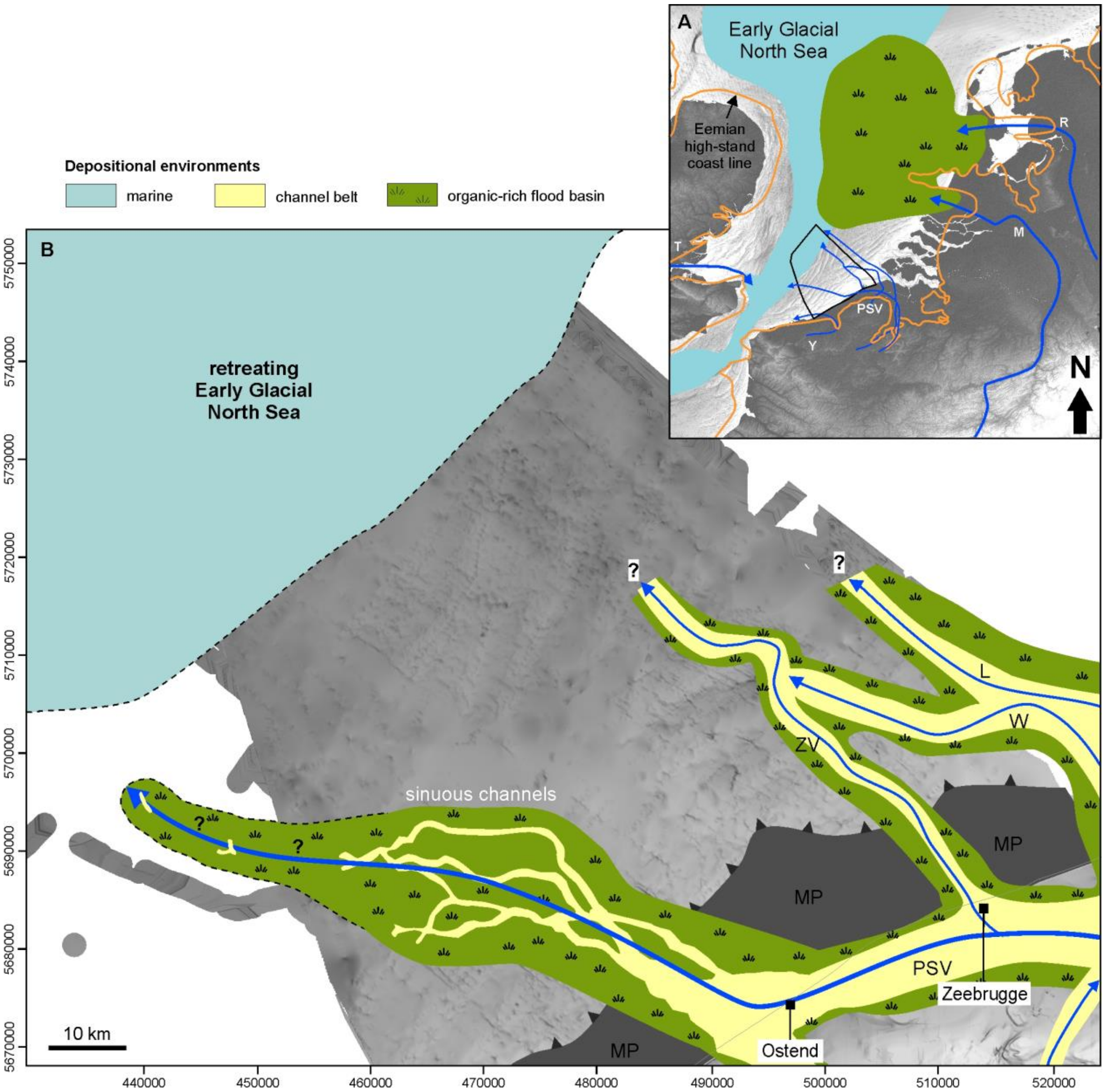


Figure 6.12 Regional (A) and local (B) scenario for the Weichselian Early Glacial. The regional scenario shows the northwest European landscape during the Eemian high-stand (orange line) as a reference to the local scenario. The local scenario shows a detailed view of the environmental conditions in which the fluvial unit SCH5 from De Clercq et al. (2018) was deposited. Rhine-Meuse delta plain after Hijma et al. (2012) and Peeters et al. (2015, 2016). Other main river courses are derived from their current flow paths, while their offshore continuations are derived from Hijma et al. (2012). The reader is referred to Figure 6.1 B for abbreviations used.

With decreasing sea level the palaeo-Scheldt Valley reclaimed its former position and extended its fluvial pathway farther offshore by scouring a series of sinuous channels on the Middle BCS (De Clercq et al., 2018; Figure 6.12 B). These channels are filled with basal medium sands overlain by an alternation of heavy consolidated peats and clays interpreted as wetland deposits (De Clercq et al., 2018). This happened most likely when sea level dropped to 30–35 m below LAT, a depth just below the Offshore Scarp, i.e. from the transition of the Brørup interstadial to the Rederstall stadial. During this period sea level dropped fast causing scouring followed by deposition of medium sands once the river base level stabilised (unit SCH5 from De Clercq et al., 2018). The sinuous channels transformed into a wetland environment in which clay and peat accumulated during the next sea-level rise of the Odderade interstadial. The Early Glacial clays identified on the FCS to the north of the Offshore Scarp (Kirby and Oele, 1975) may be contemporaneous to the clays and peat in the sinuous channels of the palaeo Scheldt Valley as they continue downstream towards this area (De Clercq et al., 2018). Based on its crosscutting relationship, we propose that the Fairy Valley has a younger Pleniglacial to Late Glacial age than these sinuous channels. No evidence of a fluvial incision is observed in the Thornton Valley related to the Brørup-Rederstall sea-level fall.

6.5.3 Weichselian Early Pleniglacial

The dynamic oscillation and unknown thickness histories of the main masses of the Early Pleniglacial ice-sheets covering British and Scandinavian bedrock further north resulted in complicated glacio-isostatic movements in northwest Europe (Lambeck et al., 2006). The actual dimensions of these major ice-sheets are poorly understood and constrained. In fact, a number of previous researchers (Carr et al., 2006; Cotterill et al., 2017; Houmark-Nielsen, 2011) have argued that the glacial history of the North Sea Basin is far more complex comprising several phases of ice-sheet lobe growth and decay; including the Ferder Episode (ca. 70 ka). According to the same authors ice-sheet limits may have been similar as during the LGM but the ice-sheet was likely to have been thinner (Lambeck et al., 2006).

As the ice masses encroached into the North Sea Basin meltwater deposited a complex sequence of glacifluvial and glaciolacustrine sediments on a laterally extensive outwash plain (Cotterill et al., 2017; Figure 6.13 A). Braided channels dispersing sediment and gravels transported by the ice from the highland areas surrounding the North Sea Basin, including the Midland Valley of Scotland and the British east coast, crisscrossed this outwash plain (Cotterill et al., 2017). Before the ice-sheets merged across the North Sea Basin drainage from the Northern European Plain and the BIIS outwash plain would have followed the northern North Sea-Atlantic Ocean route, while the Rhine, Meuse, palaeo-Scheldt and Thames, along with meltwater from the BIIS outwash plain would have followed the southern North Sea-English Channel route via the Axial Channel trunk valley. To the north, the braided channels of the BIIS outwash plain traversed the potential erratic area of Group 1 possibly already transporting clasts south via the Axial Channel.

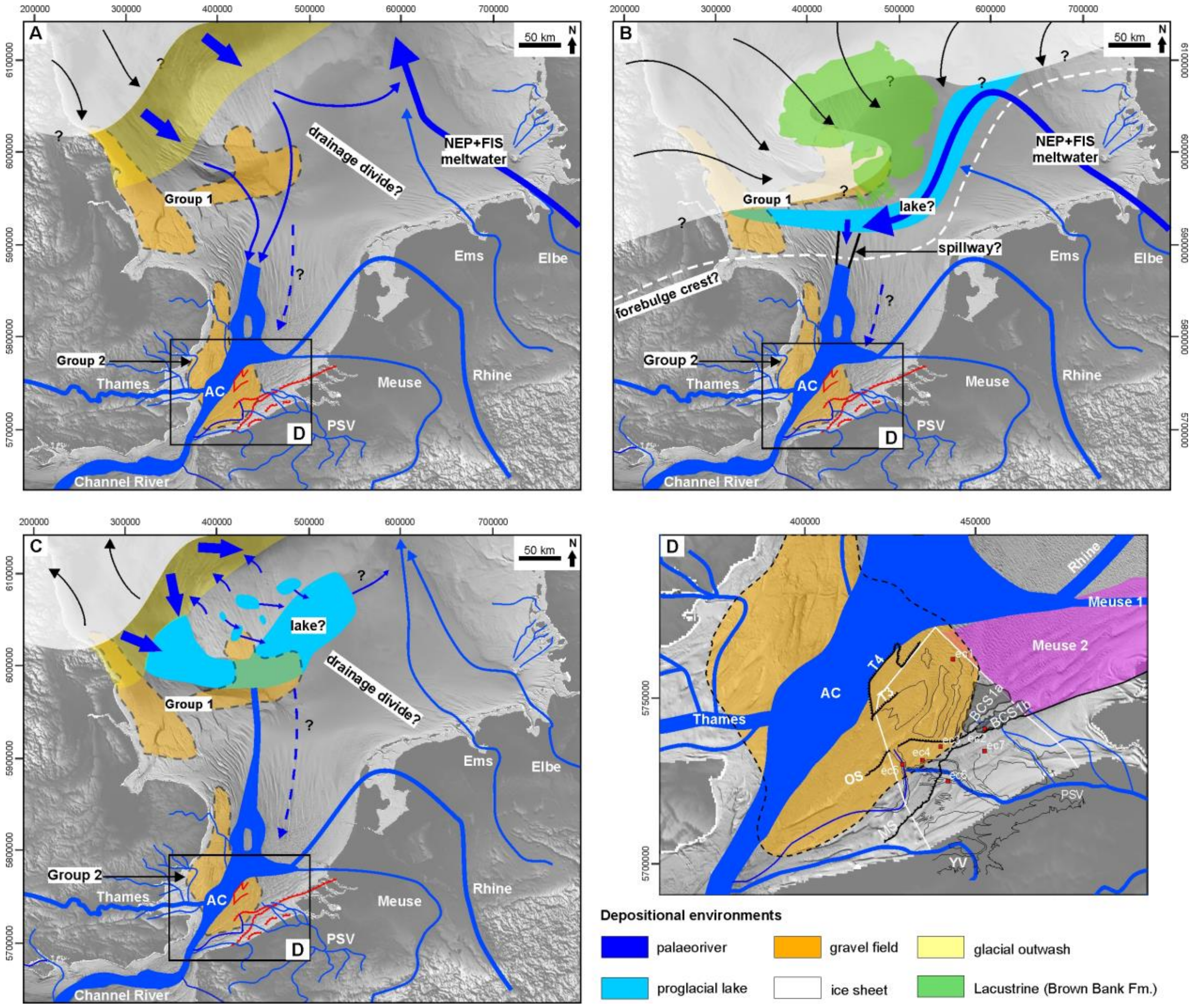


Figure 6.13 A-C) Regional scenarios for the Weichselian Early Pleniglacial at the time of the Ferder Glaciation (70 ka) based on Cotterill et al. (2017). Main river courses are derived from their current flow paths, while their offshore continuations are derived from Hijma et al. (2012). Offshore course of the Rhine from Busschers et al. (2007) and Peeters et al. (2015, 2016). Palaeo-Scheldt River derived from De Clercq et al. (2018). Axial Channel trunk valley and English Channel derived from García-Moreno (2017). NEP: Northern European Plain. D) Detailed view on the Belgian Continental Shelf with indication of the Paleogene morphology. Meuse 1: García-Moreno (2017), Meuse 2: Hijma et al. (2012) and Peteers et al. (2015, 2016). Note that Meuse 2 includes a wide braidplain compared to Meuse 1.

The growing ice masses progressively override the outwash sediments and reworked these to a clay-rich diamicton as part of the Dogger Bank Formation (Cotterill et al., 2017; Figure 6.13 B). According to Cotterill et al. (2017) the boundary of the Dogger Bank Formation represents the maximum extent of ice filling the North Sea Basin and suggest that the ice extent may have been greater than previously thought as hypothesised by Bradwell et al. (2008), Graham et al. (2007) and Sejrup et al. (2009) with the maximum ice limit being located possibly to the south of the Dogger Bank. However, subaqueous diamictons can also be produced proglacially (f.e.

Hambrey and Huddart, 1995; Lee et al., 2017), suggesting a proglacial lake may have formed in front of the merged ice-sheet. A merged ice-sheet across the North Sea Basin forced the meltwater from the FIS and Northern European Plain to join the Axial Channel trunk valley in the southern North Sea Basin potentially increasing the discharge directed to the English Channel. This major rerouting resulted in an increase in fluvial-derived terrigenous sediments in the downstream Bay of Biscay where this rerouted river system entered the Atlantic Ocean, though the amount of fluvial-derived terrigenous sediments is only half as high as during the LGM (Toucanne et al., 2009a; assuming there is an instantaneous link from source to sink and ignoring the propagation of a terrestrial signal through a coastal zone in the western English Channel to the Bay of Biscay during higher sea levels). Within this setting, the loading of the merged ice-sheet would have resulted in glacio-isostatic adjustments to the south of it potentially resulting in the formation of a proglacial lake. However, the dimensions and location of this lake remain speculative. This proglacial lake then received the meltwater from the FIS and the Northern European Plain. The lake water levels would then rise up to the limit of a forebulge crest and a spillway may be formed. Such spillway drainage of a proglacial lake may have occurred as a GLOF (a similar smaller GLOF was registered from a ca. 750 km^2 lake on top of the Dogger Bank: Cotterill et al., 2017). In the presence of a merged ice-sheet the spillway would be directed south and an unknown volume of water would then enter the Axial Channel trunk valley. The floodwaters are potentially hyper-concentrated with finer-grained sediments that are capable of transporting and depositing large volumes of material, including very large boulders, possibly including icebergs. The GLOF was also highly erosive, and consequently, deposits from such discharges would be rare within the spillways (Lord and Kehew, 1987) and would be deposited further downstream, such as on the Outer BCS (unit BCS2/BCS3).

Subsequently, the ice-sheet retreated from its maximum limit exposing the glaciated surface to periglacial activity. Braided, glacial outwash channels fed by meltwater liberated from the retreating ice masses incised this tundra-like plain, carrying glacially transported material south, possibly into a series of lakes that bordered the edges of a premature Dogger Bank within depressions or between morainic landforms (Cotterill et al., 2017;.Figure 6.13 C). Consequently the increasing volume of meltwater being released from the retreating ice mass as it retreated further north may have become dammed by these morainic landforms, leading to the formation of relatively large lakes (in some cases up to ca. 750 km^2: Cotterill et al., 2017). At some point these morainic landforms damming the lakes breached providing periodic outlets for rising meltwater levels within the lakes potentially resulting in smaller GLOFs (Cotterill et al., 2017).

The water volume from one lake may have entered into another lake resulting in a cascade of GLOFs that are able to transport sediments and gravel material. When the smaller lakes empty into the lowest and possibly largest lake (Figure 6.13 C) this process may have induced larger more catastrophic GLOFs of unknown size in the direction of the Axial Channel trunk valley, especially when a spillway was already formed during previous spillway erosion phases (Figure 6.13 B and Figure 6.14). The periodic nature of the GLOFs originating from the premature Dogger Bank may be used to suggest that when lake water levels were high enough they overtopped a pre-existing meltwater channel cut through a morainic landform, the latter forming a spillway, which drained water with suspended material from an upper lake towards the southeast into the lowest lake. This lake may then have partially drained in the direction of the Axial Channel trunk Valley but possibly also northwards draining to the Northern North Sea, where it joined the Elbe and Ems Rivers. It is not unlikely that some of these GLOFs, in the event that they were large enough, may have reached the Outer BCS. This was especially the case when the FIS and BIIS were still merged (Figure 6.13 B to C) and FIS meltwater still flowed into the Axial Channel trunk valley.

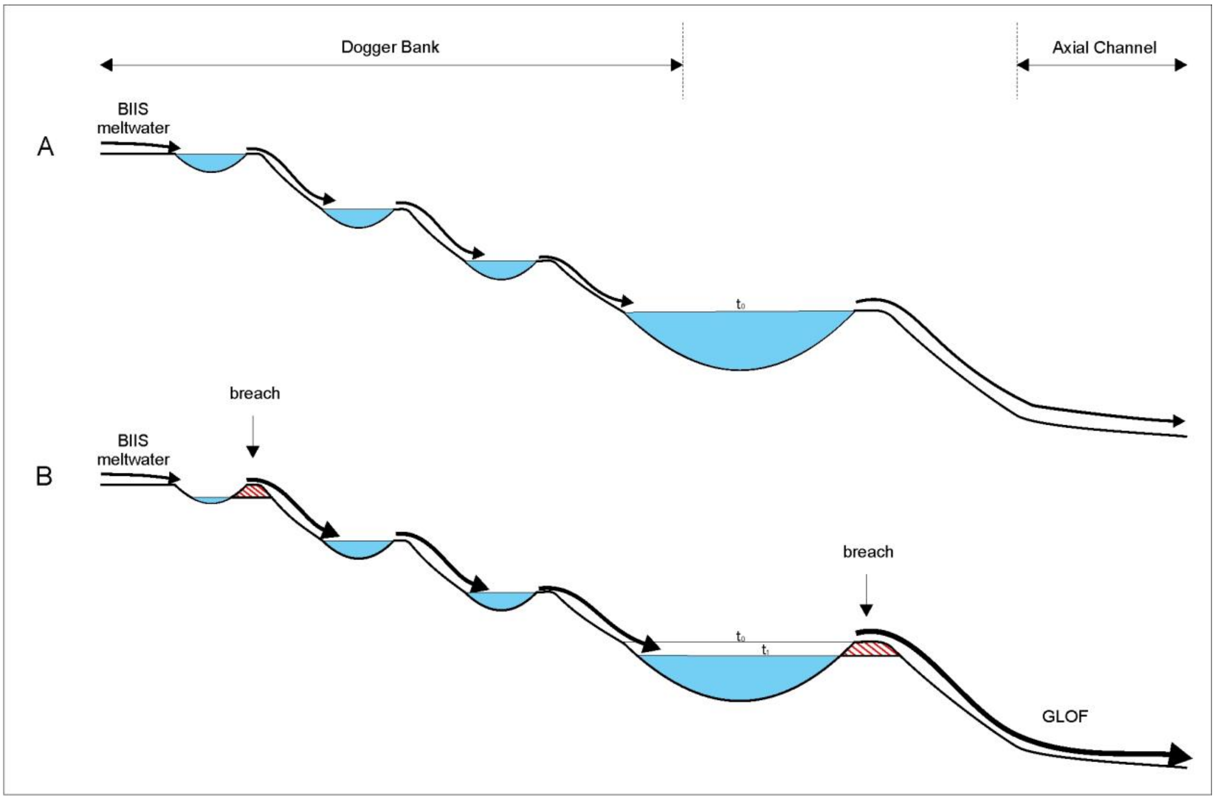


Figure 6.14 Hypothesis for multiple GLOF scenarios. A) Meltwater from the BIIS enters a series of interconnected lakes that are connected downstream to the Axial Channel. Each lake is located in a depression of damned by morainic landforms left behind in the landscape after the BIIS retreats. B) When a morainic landform fails the impounded water is released. This volume of water causes a cascade of potential overflows, possibly breaching

more morainic landforms. Downstream, by the greatest lake, this may cause a catastrophic failure of the damn (red area). Water from this lake, combined with water released from upstream is released in the Axial Channel as a GLOF. The GLOF transports material, ranging from fine-grained sand and silt to gravels and boulders in a downstream direction. Note that this scenario can occur multiple times until all damns are breached.

On the other hand, if the FIS and BIIS did not merge across the North Sea Basin the northern North Sea-Atlantic Ocean (i.e. Elbe, Weser and Northern European Plain) and southern North Sea-English Channel (Thames, Meuse, Rhine, palaeo-Scheldt) drainage routes would have been maintained. Such a drainage division alternatively explains the lower concentrations of terrigenous sediments downstream at the Bay of Biscay when compared to that of the LGM. This drainage divide may have alternatively been induced or intensified by glacio-isostatic upwarping in the North Sea Basin. Glacio-isostatic rebound models by Lambeck et al. (2006) demonstrate that an upwarped zone with a forebulge was located between Denmark and East Anglia and that the forebulge crest would be positioned somewhere at its centre. Reconstructions of the organisation and behaviour of this Early Pleniglacial river system is however largely unknown and subject to debate.

Recent studies from the Netherlands suggest that the Meuse River joined the Axial Channel trunk valley and drained towards the Dover Strait by crossing the BCS just north of the Offshore Scarp (Hijma et al., 2012; Peeters et al., 2015, 2016) or at terraces T3 and/or T4 sometime between 80 and 20 ka (García-Moreno, 2017; Hijma et al., 2012; Figure 6.13 D). However, the north to south-southwest oriented elongated palaeodepressions and ridges observed on the Outer BCS are incompatible with an east-west orientation, with the Offshore Scarp defining the southern slope of the Meuse River system. This is supported by the way subunit BCS1a is truncated in a northeast-southwest direction by unit BCS2. The current palaeogeographical scenario makes it more likely that the Axial Channel trunk valley occupied terraces T3 and T4 during episodic GLOFs or meltwater pulses during the 70 ka Ferder Glaciation.

The Early Pleniglacial is associated with a period of intense fluvial erosion across northwest Europe (Huijzer and Vandenberghe, 1998; Huissteden et al., 2001). Busschers et al. (2007) regard the Rhine and Meuse Rivers as ‘braid belts’ and ‘channel belts’, formed by synchronously operating fluvial processes of channel scour, bar deposition and reworking, and lateral migration and widening of former active channels within these belts and that the infilling units have each formed over millennia to tens of millennia. The downstream base of the Pleniglacial Meuse River sediments contains a sharp erosional boundary and consists of a gravel lag (Hijma et al., 2012). This suggests that, if the Meuse River followed a course across the Outer BCS, with the Offshore Scarp as its southern boundary, the BCS1a and BCS1b Eemian sediments would have been reworked. These observations are however incompatible with the observed north to south-southwest elongated palaeodepressions scoured in the pre-Quaternary surface of the Outer BCS and the truncation observed at the northwest edge of subunit BCS1a. Moreover, unit BCS2 infills these north to south-southwest palaeodepressions suggesting that the formation of these palaeodepressions and the deposition of unit BCS2 are related to the same event with an orientation almost perpendicular to the proposed Rhine-Meuse River projection of Hijma et al. (2012) and Peeters et al. (2015, 2016). The above summarised observations of the Rhine-Meuse River orientation this far south is not compatible with observations from our study area and suggest that a more northern and eastern pathway directed to the Axial Channel trunk valley was used as proposed by García-Moreno (2017) at least as early as from 70 ka,. The orientation of these incisions, which are oblique to the Axial Channel trunk valley, the drainage pathway towards the Dover Strait, implies large volumes of meltwater during GLOFs to be directed to the southern North Sea Basin, resulting in a wide floodplain that stretches across the Outer BCS. Evidence for such meltwater pulses is provided by Lambeck et al. (2010) who suggest that the Early Pleniglacial FIS retreat was substantial and did experience an oscillatory behaviour similar to that of the Saalian (Toucanne et al., 2009a) and LGM (Toucanne et al., 2015). It is thus possible that the formation of the observed palaeodepressions on the Outer BCS and the sedimentation of unit BCS2 initiated during the Early Pleniglacial. However, no direct evidence or dating exists to substantiate this hypothesis.

6.5.4 Weichselian Middle Pleniglacial

After the relatively cold Early Pleniglacial, climate improved and humidity increased (Helmens, 2014). This climate improvement caused the ice-sheets to shrink during the first part of the Middle Pleniglacial and even almost disappear (Brown et al., 2007; Lambeck et al., 2010). This is supported by glacio-isostatic models (Lambeck et al., 2010), showing a shrinking FIS that caused a reduction in forebulge height. These events caused the drainage network of northwest Europe to be divided in one flowing toward the northern North Sea-Atlantic Ocean and another one toward the Dover Strait-English Channel. These drainage routes are similar to those observed at the end of the Early Pleniglacial (Figure 6.13 C) and suggest that a drainage divide persisted in the landscape or that the Early Pleniglacial river systems retained their former positions downstream in the southern North Sea. The persistence of a drainage divide is also observed as a decrease in the fluvial-derived terrigenous sediments in the downstream Bay of Biscay (Toucanne et al., 2009a). The impact of these landscape changes on the BCS is unclear as no sediments or direct morphological evidence is preserved. Dutch studies suggest that the Rhine River avulsed upstream between 60 and 40 ka and that it joined the Meuse River upstream (Busschers et al., 2007). This joint Rhine-Meuse River system follows the Early Pleniglacial route of the Meuse River across the BCS (Hijma et al., 2012; Peeters et al., 2015, 2016). However, the location and truncation of unit BCS1 suggests no Rhine-Meuse River system to have flowed here. This suggests that, like during the Early Pleniglacial, the Rhine-Meuse River system might have followed a more northern and eastern route joining the Axial Channel, as proposed by García-Moreno (2017).

6.5.5 Weichselian Late Pleniglacial

Climate reconstructions indicate that during the Middle to Late Pleniglacial transition climate was extremely cold and dry in northwest Europe (Helmens, 2014). This resulted in the formation and ice mass expansion of the FIS and BIIS, which once again merged across the North Sea Basin (Clark et al., 2012; Lambeck et al., 2010; Mangerud et al., 2011; Sejrup et al., 2009). The ice-sheets reached their maximum extent between ca. 30 and 18 ka, i.e. the LGM (Clark et al., 2012; Sejrup et al., 2016; Figure 6.15 A and B). Sea level dropped to levels 130 m below present-day (Medina-Elizalde, 2013). One of the greatest impacts of this ice-sheet expansion and sea-level lowering on northwest European palaeogeography was the modification of the river network (Toucanne et al., 2015) similar to the Saalian and Early Pleniglacial. The Axial Channel trunk valley channelled significant volumes of water to the North Atlantic via the English Channel route resulting in the largest river system that drained the European continent at the time (Gibbard, 1988; Lericolais et al., 2003; Ménot et al., 2006: Figure 6.15 A and B). When the merged ice-sheet was at its maximum size at 22.73 ka (Patton et al., 2016), the total catchment of the fluvial system that drained into the Bay of Biscay covered an area of 2.56 x 10^6 km^2 (Patton et al., 2017) incorporating drainage systems from the North European Plain all the way up to Poland (Figure 6.15 A). Over half of its catchment was ice-covered (1.36 x 10^6 km^2: Patton et al., 2017), capturing subglacial meltwater drainage from the FIS and BIIS. According to Patton et al. (2017) and Lambeck et al. (2010) proglacial lakes were present along the rims of the FIS. Two of these lakes belonged to the drainage system of the Northern European Plain and drained in the direction of the southern North Sea (Figure 6.15 A).

It has been hypothesised that during the LGM a proglacial lake was present in the southern North Sea Basin (Hijma et al., 2012) and that this lake was located somewhere in the German-

Danish continental shelf (Sejrup et al., 2016; García-Moreno, 2017; Patton et al., 2017). However, the precise dimensions and position of that lake are unclear, although it has been proposed that it was located along the southern margin of the merged ice-sheet between the northern Netherlands, Germany and Denmark (Clark et al., 2012; Patton et al., 2017; Sejrup et al., 2016), where a topographic depression is present in the present seafloor. Glacio-isostatic models predict that, in case of a merged ice-sheet across the North Sea Basin, a convergence zone of maximum glacio-isostatic upwarping is positioned across the Netherlands, northeast Belgium and easternmost Germany (Peltier, 2004; Figure 6.15 B). The height and position of this upwarped zone would have affected the dimensions, position and depth of a supposed proglacial lake. To its southwest the lake was believed to be connected to the Axial Channel trunk valley through a spillway channel (García-Moreno, 2017). A similar channel is envisaged by Hijma et al. (2012) more to the east. In the absence of a proglacial lake it is proposed that the FIS meltwater was diverted southwards in the direction of the Axial Channel trunk valley through these supposed spillway channels (Hijma et al., 2012). The glacio-isostatic impacted palaeotopography combined with the glacial and fluvial systems thus exerted a major control on a supposed proglacial lake and the downstream Axial Channel trunk valley during the LGM and its early deglaciation. Deglaciation appears to have occurred in several phases of ice-sheet retreat and advance. Three phases of retreat and advance are registered for the BIIS across the Dogger Bank (Phillips et al., 2018) and five phases of ice-sheet retreat and advance are observed for the FIS (Toucanne et al., 2015). The ice-sheet oscillations for the FIS were verified through ^{14}C dating of foraminifers collected from fluvial-derived sediments at the Bay of Biscay and occur between ca. 30.7 and 28.9 ka, between ca. 25.7 and 23.4 ka, between ca. 22.5 and 21.3 ka, between ca. 20.3 and 18.7 ka and between ca. 18.2 and 16.7 ka (see Toucanne et al., 2015).

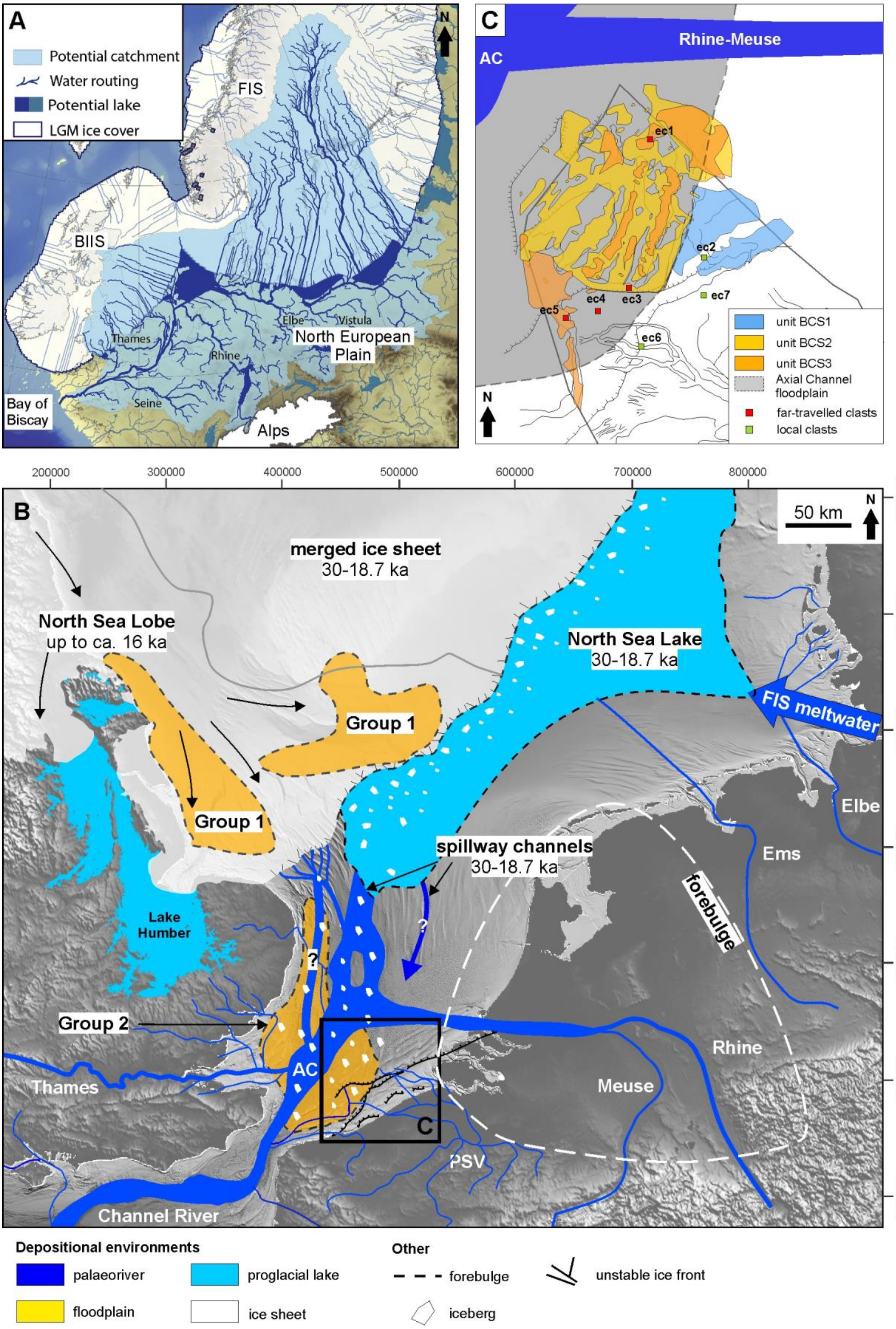

A
Potential catchment
Water routing
Potential lake
LGM ice cover
FIS
BIIS
Thames
Rhine
Elbe
Vistula
North European Plain
Seine
Bay of Biscay
Alps
C
Rhine-Meuse
AC
ec1
ec2
ec3
ec4
ec5
ec6
ec7
unit BCS1
unit BCS2
unit BCS3
Axial Channel floodplain
far-travelled clasts
local clasts
B
50 km
merged ice sheet
30-18.7 ka
North Sea Lobe
up to ca. 16 ka
North Sea Lake
30-18.7 ka
Group 1
Group 1
FIS meltwater
Elbe
Ems
spillway channels
30-18.7 ka
forebulge
Lake Humber
Group 2
AC
Thames
Meuse
Rhine
PSV
Channel River
200000
300000
400000
500000
600000
700000
800000
6200000
6100000
6000000
5900000
5800000
5700000
5600000
Depositional environments
palaeoriver
proglacial lake
floodplain
ice sheet
Other
forebulge
unstable ice front
iceberg

Figure 6.15 (previous page) A) The complete northwest European catchment with a terminus in the Bay of Biscay during the Last Glacial Maximum at 22.73 ka (modified from Patton et al., 2017). B) Palaeogeographic reconstruction for the Weichselian Early Pleniglacial and Last Glacial Maximum in northwest Europe. C) Detailed view on the Middle-Outer BCS and the relation between the landscape and the identified units and erratic clasts within this study. Lobourg-Axial Channel from García-Moreno, (2017); Rhine-Meuse River: Hijma et al. (2012) and Peeters et al. (2015, 2016). Other main river courses are derived from their current flow paths, while their offshore continuations are derived from Hijma et al. (2012) and García-Moreno, (2017). North Sea Lake: Sejrup et al., 2016 and Patton et al. (2017). British lakes: BRITICE compilation from (Clark et al., 2004; Evans et al., 2005, 2017). North Sea Lobe: Clark et al. (2012) and Davies et al. (2011). Forebulge modified from Peltier (2004). Southern North Sea floodplain is based on the topographic relief and the highest position of erratic clasts found on the BCS. Glacifluvial pathways from the North Sea Lobe are derived from Dove et al. (2017).

The youngest phase took place when the merged ice-sheet detaches and the proglacial lake subsequently catastrophically drains to the northern North Sea (Hjelstuen et al., 2017). During the first four phases, between 30.7 and 18.7 ka, significant volumes of meltwater were directed to the English Channel and sediments were deposited at the Bay of Biscay (Toucanne et al., 2015). In the event a proglacial lake was positioned in the southern North Sea, meltwater from both the BIIS and FIS would have pounded within the lake allowing its dimensions to grow dramatically, a process that would be intensified during the deglaciation. In this scenario a proglacial lake would have reached a threshold once it reached the height of its spillway channel to the south, followed by a possible breach releasing large amounts of water into the Axial Channel trunk valley. This 'breach event', also called a GLOF, could for example be induced by large meltwater pulses released from other proglacial lakes positioned along the rim of the FIS (e.g. Meinsen et al., 2011).

In the event of a GLOF or great discharge events the flow would enter the Axial Channel trunk valley and be directed towards the Dover Strait. Evidence for LGM megafloods are observed in the youngest channel incision of the Axial Channel trunk valley at the Dover Strait (see García-Moreno, 2017). There, a range of bedrock erosional features, such as amphitheatre-head channels, streamlined islands, longitudinal scours, linear ridge-and-groove bedforms, are identified that are typically found in megaflood-eroded terrains. Additional arguments for a GLOF in the southern North Sea Basin is provided by the north to south-southwest oriented palaeodepression network identified in the pre-Quaternary surface of the Outer BCS. Indeed, when large volumes of water enter into the Axial Channel trunk valley system the head of the flood pulse may have breached the confinements of the Axial Channel trunk valley and travelled south via the relatively flat southern North Sea floor on to the Outer BCS. The BCS may therefore be an important location in the Axial Channel trunk valley as it is located in the direct path of where the floods flow south and forms a bend towards the west with the Late Saalian Meuse terrace, e.g. the Offshore Scarp. The high-erosive head of a GLOF erodes large parts of the pre-Quaternary substratum and of the Eemian marine to near-coastal sediments (subunit BCS1a: Figure 6.15 C). At its western edge the Offshore Scarp is only ca. 4–6 m high and the GLOF was able to surpass the Offshore Scarp to cross parts of the Middle BCS.

During a GLOF, a melange of coarse-grained sediments mixed with shell material and erratic clasts (up to 3 m) were deposited on the Outer BCS, whereas only clasts are deposited on the Middle BCS on top of the Offshore Scarp. When the fluvio-hydraulic energy subsides finer-grained sand, silt and clay were able to settle from the remaining water body, first within the scoured palaeodepressions (e.g. facies F1 of unit BCS2) and later on top of these, when most of the channels were filled, a draped deposit (e.g. facies F2 of unit BCS2; Figure 6.4 C). When the head of a GLOF impacts the Offshore Scarp the water inundates the higher platform of the Middle BCS and is able to deposit gravels, cobbles, boulders and rocks. The strong erosive boundary between units BCS2 and BCS3 and their similar alternations of fine- and coarse-grained layers seems to suggest that at least more than one GLOF or major flood event may have taken place during the deglaciation following the LGM during each of the ice-sheet retreat events, recorded by Phillips et al. (2018) and Toucanne et al. (2015) between 30 and 18.7 ka.

6.6 Erratic clast transportation mechanism

The precise mechanism for the GLOFs in the southern North Sea Basin and the timing of deposition, ranging from Early Pleniglacial and/or LGM, cannot be determined, but several general inferences can be made based on evidence from past GLOF-eroded terrains, such as the Dover Strait (Collier et al., 2015; García-Moreno, 2017; Gupta et al., 2007, 2017), the Channeled Scabland region (Baker and Nummedal, 1978; Bretz, 1969, 1923) and in many other areas of North America (Clarke and Mathews, 1981; Kehew, 1993; Kor et al., 1991; Murton and Belshaw, 2011; Wiedmer et al., 2010). The geological record shows that GLOFs are known to cause devastating effects in their track, moving vast quantities of debris from the subglacial, terminal and proximal areas of ice masses (Menzies, 2002). During a GLOF icebergs may be left stranded in the landscape and substantial thicknesses of sediment across a wide range of grain sizes are deposited, new channels may be formed and previous ones abandoned, topographic barriers may be breached or removed and substantial areas of bedrock/substratum may be eroded. Here, we will provide arguments to the transportation mechanisms that resulted in the deposition of a gravel field, containing both boulders and rocks, observed on the Outer-Middle BCS at the foot and top of the Offshore Scarp (Figure 6.10).

The boulders of erratic clast group ec1 and ec3–ec5 comprise both sub-rounded to sub-angular clasts (Dusar, 2014; Dusar et al., 2016). Similar observations of both sub-rounded to even angular cobbles have been observed within the same gravel field by Van Lancker et al. (2005, 2007). On the other hand, the 2–3 m rocks observed on the Middle BCS, in the eastern section of the gravel field, appear to be sub-rounded to rounded as observed from the backscatter images (Figure 6.10 C).

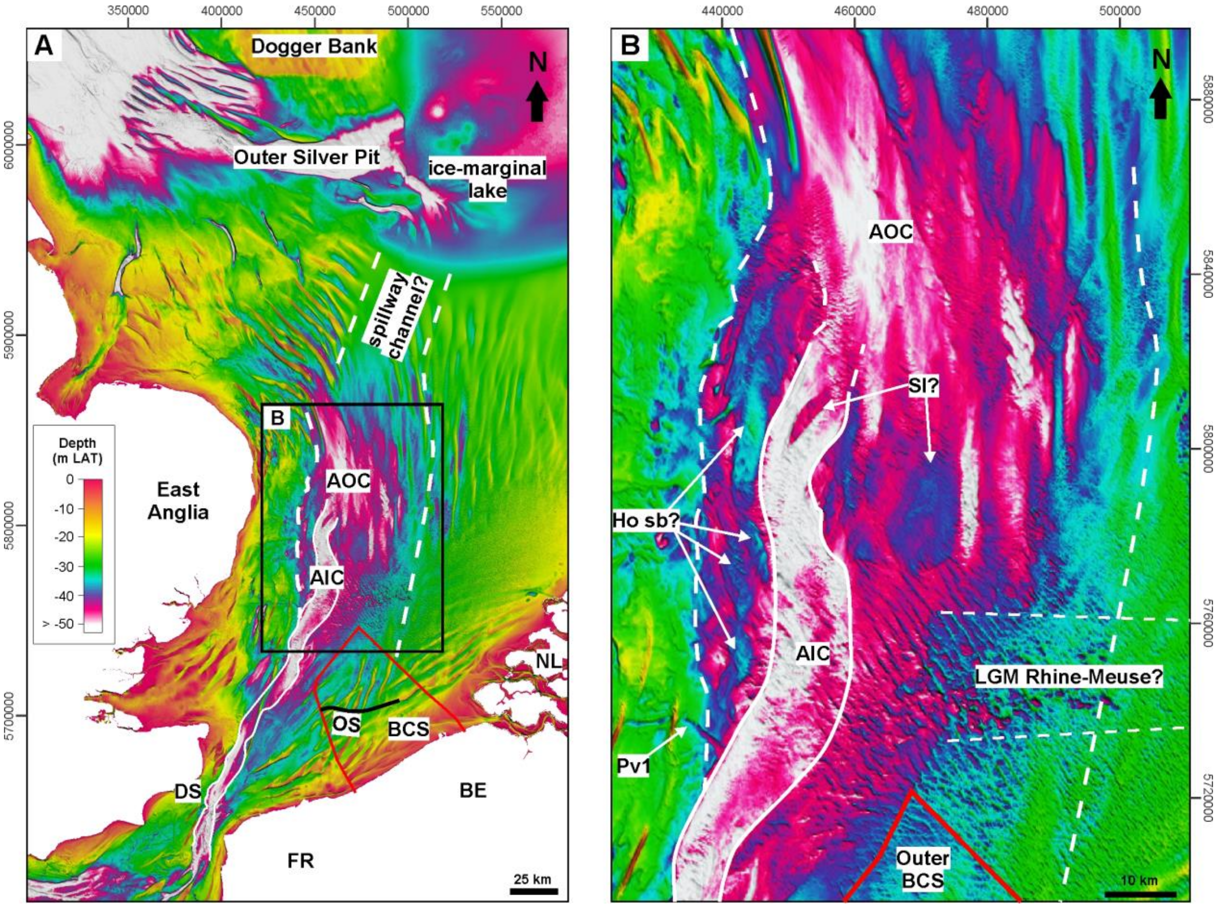


Figure 6.16 Regional (A) and detailed (B) overview of the Axial Channel trunk valley morphology in the bathymetry. Location of the spillway channel is modified from García-Moreno (2017). Si: streamlined island: sb: sandbank; Pv1: palaeovalley 1.

There are two main factors that control the morphology of clasts (Lindholm, 1987): 1) their initial morphology, which is inherited from the source rock; and 2) the mechanical alteration (abrasion and fracturing) during transport. The initial shape of gravel-sized and larger clasts depends on jointing and cleavage as well as on the rock type and the climate of the source area

(Lindholm, 1987). Gravels derived from crystalline source rocks are generally angular, while gravel-derived sandstones will inherit some degree of roundedness. During transport, abrasion and fracturing may strongly modify particle morphology. Dominantly rounded clasts have undergone transport in a fluvial medium such as a river, while angular clasts likely originate from a glacial environment, such as transport by ice masses (Boggs, 2009; Middleton et al., 2003). As a result the total clast population from the Middle-Outer BCS suggests that the more rounded clasts were transported by a fluvial medium and that the angular clasts were transported by an ice medium.

The above presented landscape evolutions of the Early Pleniglacial and the LGM presents a scenario in which ice masses merged across the North Sea Basin. Several authors demonstrated that the BIIS (Bateman et al., 2015; Busfield et al., 2015; Davies et al., 2011; Larkin et al., 2011; Roberts et al., 2013) and the braided river systems of the proglacial outwash plain (Cotterill et al., 2017; Phillips et al., 2018) transported bedrock material, originating from the Scottish Grampian Highlands and the British East Coast, south in the direction of the Dogger Bank and the area to its west, up to the British coast, and possibly extended farther east depending on the ice limit.

During the maximum glaciation of the Early Pleniglacial and LGM a proglacial lake is hypothesised in the southern North Sea (Clark et al., 2012; Mangerud et al., 2011; Sejrup et al., 2016). This lake was bounded by the ice masses to the north and glacio-isostatic upwarped palaeotopography to its southeast and was fed by meltwater from the FIS and BIIS (Figure 6.13 and Figure 6.15). From the period of maximum glaciation up to the final deglaciation the ice masses experience an oscillatory behaviour as a precursor to the final deglaciation (Phillips et al., 2018; Toucanne et al., 2015). When the ice masses retreat the total lake surface may increase, depending on the equilibrium-state between ice mass disintegration and glacio-isostatic adjustment of the landscape that depends on the position of the oscillating ice mass. During this deglaciation phase the ice front that enters the proglacial lake may remain grounded or become floating. In both cases the ice front starts to calve due to its interaction with the lake water, thus contributing to ice mass loss (Greve and Blatter, 2009). Similar to icebergs from ice shelves released into the oceans, where icebergs are able to travel hundreds of kilometres out into the open sea driven by wind and ocean currents, the icebergs in a proglacial lake may have a prolonged life span. These icebergs would then be able to carry the ice-transported sediments and clasts from Scotland and the British East Coast across the proglacial lake to its outermost edges.

During short episodes of deglaciation massive amounts of meltwater from the BIIS and FIS enter the southern North Sea proglacial lake resulting in one or more GLOFs. The initiation of a GLOF requires the failure of a natural dam, such as a moraine or during the previous Elsterian and Saalian glaciations in the southern North Sea glacio-isostatic upwarped bedrock or substratum (De Clercq et al., 2018; García-Moreno, 2017; Moreau et al., 2015). In both cases a catastrophic breach is induced and releases a large volume of meltwater into the relatively flat southern North Sea plain. This happens by vertical and lateral erosion of the spillway channel that connects the lake to the Axial Channel trunk valley. The water discharged from the proglacial lake would have, at first, contained little or no sediment, but sediment concentrations would have increased rapidly downstream as the spillway and the northern section of the Axial Channel trunk valley were entrenched (Lord and Kehew, 1987). The sediment concentrations in the water rapidly increase and turn into a hyper-concentrated flow (Lord and Kehew, 1987; Maizels, 1997; Menzies, 2002). After sediment concentrations peaked in the head of the surge, they probably gradually decreased because the rate of erosion must have decreased as the dimensions of the Axial Channel trunk valley became large enough to contain the discharges and the spillway and northern section of the Axial Channel trunk valley were incised to depths of more resistant sediments (cf. Lord and Kehew, 1987). Similar to observations in alpine areas it is expected that the sediment concentration peak typically precedes the discharge peak (Beecroft, 1983; Hewitt, 1982;) and suggests that this part of the water body has the largest potential to transport very coarse-grained material such as boulders and rocks. Indeed, ancient and present-day boulder deposits have been recorded up to hundreds of kilometres downstream spillway channels, which carried GLOF meltwater from Lake Agassiz (Matsch, 1983; Teller and Thorleifson, 1987), Lake Nipigon (Teller and Thorleifson, 1983), Lake Wisconsin (Clayton and Attig, 1987), and the Missoula and Lake Bonneville flood routeways (Baker, 1973a; O'Connor, 1993) in the United States, but also in Icelandic sandurs (Maizels, 1992).

Due to the very-high fluvio-hydraulic energy at the head of the surge, combined with the large volume of meltwater released, it is expected that potentially large parts of the southern North Sea Basin become inundated and are affected by a GLOF. This is particularly the case between East Anglia and the central Netherlands, just south of the spillway channel, where the Axial Channel trunk valley appears to widen (here named the Axial Outer Channel trunk valley). This is apparent from the seafloor bathymetry (Figure 6.16) and is supported by observations made by García-Moreno (2017) in this section of the Axial Channel trunk valley. Here, streamlined islands appear to be present, an indication of very-high fluvio-hydraulic energy similar to megafloods as observed in the Dover Strait (Collier et al., 2015; García-Moreno, 2017; Gupta et al., 2007, 2017), the Channeled Scabland region (Baker and Nummedal, 1978; Bretz, 1969, 1923) and in many other areas of North America (Clarke and Mathews, 1981; Kehew, 1993; Kor et al., 1991; Murton and Belshaw, 2011; Wiedmer et al., 2010). This Axial Outer Channel trunk valley was likely only active during GLOF events, possibly as early as the Early Pleniglacial but definitely during the LGM, possibly up to four times between 30 and 18.7 ka (Toucanne et al., 2015). The eastern boundary of this channel is most unclear, covered by younger Holocene marine sands, but based on the observations made on the Outer BCS the edge of the channel must have been located between 30 and 40 km to the east. During the LGM, this would imply that the Rhine-Meuse River mouth, which was inferred by García-Moreno (2017) to be located north of the BCS, would be overrun during a GLOF event.

When the head of the surge within the Axial Outer Channel trunk valley travels south into the BCS it will encounter the Offshore Scarp as an obstacle that impedes the southward drainage in the direction of the Dover Strait. Upon impact the hyper-concentrated flow likely overtopped the scarp and flowed onto the Middle BCS. Here, the flood velocity diminishes thereby decreasing the capacity of the flood to transport very coarse-grained material such as cobbles, boulders and rocks. These events may explain the location of a gravel field surrounding the Offshore Scarp on the BCS. It is likely that icebergs, as a result of calving, were present in the lake. During a GLOF these icebergs were likely transported south and may have grounded along the eastern and southern margin of the Axial Outer Channel trunk valley. Upon collision with the Offshore Scarp there is a possibility that icebergs are transported and deposited on the higher Middle BCS. When the icebergs finally melt they release their load consisting from fine-grained sediments up to boulders and possibly also rocks. Finally, it is possible that terraces T3 and T4 on the Outer BCS and UKCS were formed during the Early Pleniglacial (ca. 70 ka) and/or LGM (30–18.7 ka) and that this is the result of GLOFs depending on the orientation of the GLOF within the Axial Outer Channel trunk valley.

6.7 Conclusions

In this study we have integrated very-high-resolution seismic reflection data and archived core-vibrocore data of the Middle-Outer BCS and have provided a new insight on the nature of the Late Pleistocene sediments deposited in this area and what their relationship is with the northwest European landscape evolution. The main findings are:

- Early Eemian marine and wave-dominated estuary environments were formed at the Late Saalian Meuse terrace. The transition from a wave-dominated estuary on the Outer BCS to a tide-dominated estuary on the Inner BCS shows that during the Early-Middle Eemian period the southern and norther North Seas connected and that this has a profound impact on coastal development across the North Sea.
- The existence of Early Pleniglacial and/or Last Glacial Maximum glacial lake outburst floods. These are evidenced from:

  - A network of north to south-southwest incised palaeodepressions on the Outer BCS with the Offshore Scarp as its topographical southern boundary. The Offshore Scarp is thought to represent an important topographical barrier for outburst floods, breaking their momentum and diverting it westwards.

  - Two braided channel belt units (BCS2 and BCS3). Unit BCS2 represents two phases of infilling after initial incision, while unit BCS3 represents a final phase of deposition.

  - A truncation of Eemian unit BCS1a by BCS2 parallel to the palaeodepression network. This erosion phase occurred contemporaneously with the formation of the palaeodepression network.

  - Far-travelled erratic clasts, originating from the Scottish Grampian Highlands and the British East Coast, ranging in size from cobbles to rocks of 3 m in diameter were founds across the Outer and Middle BCS with a major field centred around the western part of the Offshore Scarp. Clasts of this large a size can only be transported through a fluvial medium with very high fluvio-hydraulic energy.
- A new transportation mechanism was worked out for the far-travelled erratic clasts of the BCS. They have been transported from their provenance areas through glacial scouring and fluvial outwash processes to the Dogger Bank and the area east of it. During the Early Pleniglacial and/or Last Glacial Maximum they were released into a proglacial lake and transported south through one or multiple GLOF events.

6.8 Acknowledgements

This study was carried out within the framework of the IWT SBO Project ‘SeArch’ (Archaeological heritage in the North Sea, project nr. 120003) and is funded by IWT grant project nr. 131308. We would like to thank the Captains and Crews of the RV Belgica and RV Simon Stevin for their excellent work during the geophysical campaigns. We would also like to thank Koen De Rycker for the great job he did during the data acquisition. We acknowledge Belwind, Nobelwind and C-Power for the use of geological data. This research greatly benefited from discussions with Freek Busschers at various stages of the work. We dedicate this research to Jean-Pierre Henriet who passed away at the age of 72 in April 2017.

7 PRESERVATION POTENTIAL ASSESSMENT FOR SUBMERGED HERITAGE ON THE BELGIAN CONTINENTAL SHELF

Abstract

This chapter describes the preservation potential of the Pleistocene geological record below and around the BCS, as a background to study the distribution of the preservation potential for submerged heritage. Moreover, palaeogeographical reconstructions show the distribution between the main areas of sediment accumulation and areas dominated by erosion. They provide the prerequisites for solid inferences on and predictions of the palaeontological and archaeological preservation potential. The preserved deposits on the BCS show an irregular, patchy distribution pattern with varying depths of layer bases, thus using just simple layer-based models or depth-relates-to-age expectation models will not suffice. Such depositional records are hard to unravel and both the lithology and fossil content often show many elements reworked from earlier phases and require palaeogeographical reconstruction to stratify and order them, as attempted in this chapter.

7.1 Introduction

During the Pleistocene the North Sea and its coastal plains evolved between long-term dry terrestrial environments and short-lived interglacial shallow seas. In between these glacial-interglacial transitions dramatic landscape changes took place that are related to sea-level changes, ice-sheet oscillations and glacio-isostasy. These regional-scale processes control the preservation potential of the depositional environments that once were part of these landscapes. Many of these now buried landscapes were able to sustain large mammalian herds and the animals that preyed upon them, including early humans that migrated to the area presently known as northwest Europe from 0.8–1 Ma onwards (Parfitt et al., 2010). As a result the relatively shallow southern North Sea is one of the richest marine fossil-bearing localities worldwide, with tons of Pleistocene bone material brought ashore every year (Bicket and Tizzard, 2015; Mol et al., 2006; Van Kolfschoten and van Essen, 2004). This bone material is primarily retrieved during fishing activities, aggregate extraction and during dredging operations related to the maintenance of shipping lanes and navigation channels. However, on the BCS the number of finds, compared to the Netherlands (Mol et al., 2006; Van Kolfschoten and van Essen, 2004) and the United Kingdom (Bicket and Tizzard, 2015), is very low. With the recent SeArch project (Missiaen et al., 2017) a first ever overview is provided of all fossil finds on the BCS (Vermeersch et al., 2015). The majority of the finds are located within 5 km of the coastline and are found near/in navigation channels. This sampling bias is mostly the result of industrial activities within these areas (maintenance dredging and aggregate extraction). Almost all the finds are located within the palaeo-Scheldt Valley system (composed of the Coastal-Ostend Valley and Zeebrugge-Waardamme-Lys Valley). This now buried incised-valley system connects landward to the Flemish Valley (Figure 7.1). Although the majority of the finds on the BCS have been identified (Vermeersch et al., 2015) they are still lacking a geological context. Moreover, they have an ambiguous broad age determination that often spans a full glacial-interglacial cycle.

This study will use the spatio-temporal distribution of the BCS deposits and their relationship to the landscape evolution to provide, for the first time, a geological context for past and future archaeological and palaeontological (i.e. submerged heritage) finds on the BCS. This is important as more infrastructural works are planned at sea in the near future, in particular in the near-coastal zone (e.g. in the framework of the ‘Complex Project Kustvisie’, formerly known as ‘Vlaamse Baaien’: www.kustvisie.be). These construction works may not only disturb the geological layers underneath but often also require large volumes of sediments to be extracted from other locations on the BCS. Finds recovered during such activities need to be situated in their proper geological setting in order to translate them into relevant (and correct) information about past landscapes and their inhabitants.

The southern North Sea Basin is a critical area to answer some important questions in the field of palaeoanthropology as northwest Europe was at the edge of hominin expansion during major parts of the Pleistocene (e.g. Parfitt et al., 2005, 2010; Roebroeks, 2006). Moreover, establishing the patterns of presence and absence of submerged heritage is of crucial importance to the offshore industry as a better knowledge framework may assist them in pro-active counter measurements that help them save time and money.

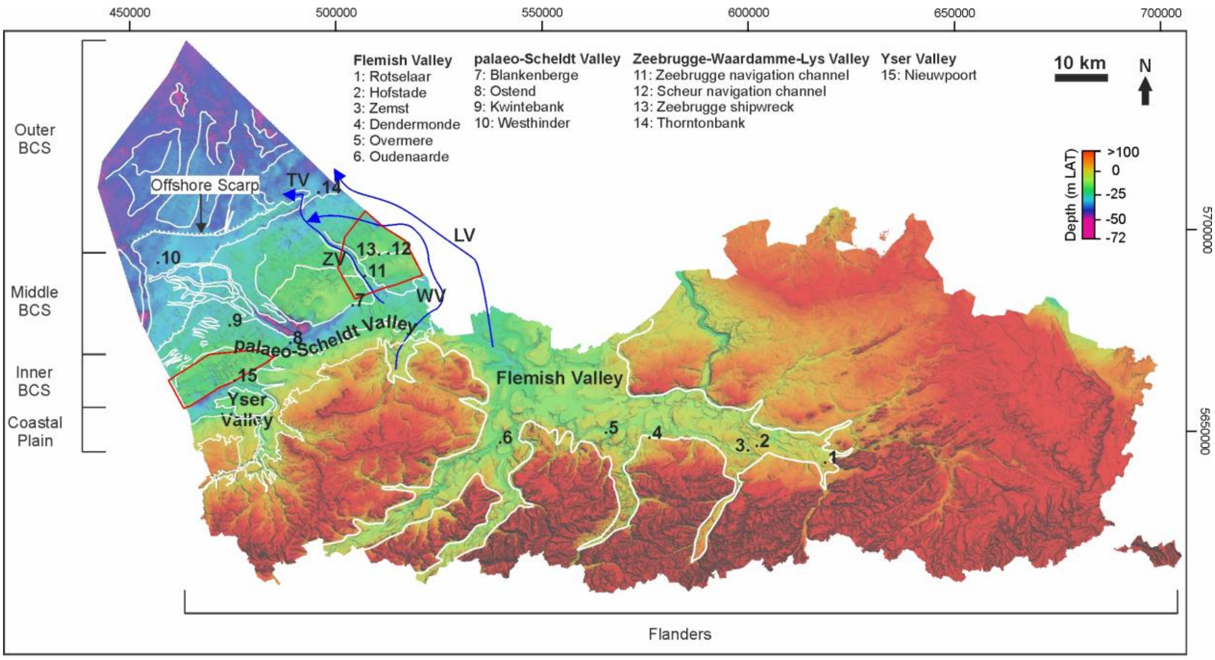


Figure 7.1 The Belgian Continental Shelf and Flanders projected on top of the Paleogene substratum with indication of onshore (Gautier, 1995; Germonpré et al., 1993b) and offshore mammalian find sites (Vermeersch et al., 2015). ZV: Zeebrugge Valley; WV: Waardamme Valley; LV: Lys Valley; TV: Thornton Valley.

7.2 Methodology

The Middle-Late Pleistocene archaeological and palaeontological record of Flanders has been studied in great detail in the past few decades. Sedimentary units and landforms have been discriminated, dated and (lithogenetically) interpreted (for an overview see Demoulin, 2017). Within these sedimentary environments or on top of landforms numerous archaeological and palaeontological discoveries have been made and extensively studied.

The Flanders Heritage Agency provides a state-of-knowledge overview for each archaeological period in Flanders (www.onderzoeksbalans.onroerenderfgoed.be). We provide a summary overview of their findings, where archaeological sites are located, what their context may be (when provided) and why they are found at these locations. This overview is based on several pre-defined geomorphological regions in Flanders (Figure 7.2): South Flanders, Sandy Flanders, the Campine Area, the Polders, and the Flemish and Meuse Valleys. Except for the latter valleys, these areas are predefined as agricultural areas based on differences in soil composition.

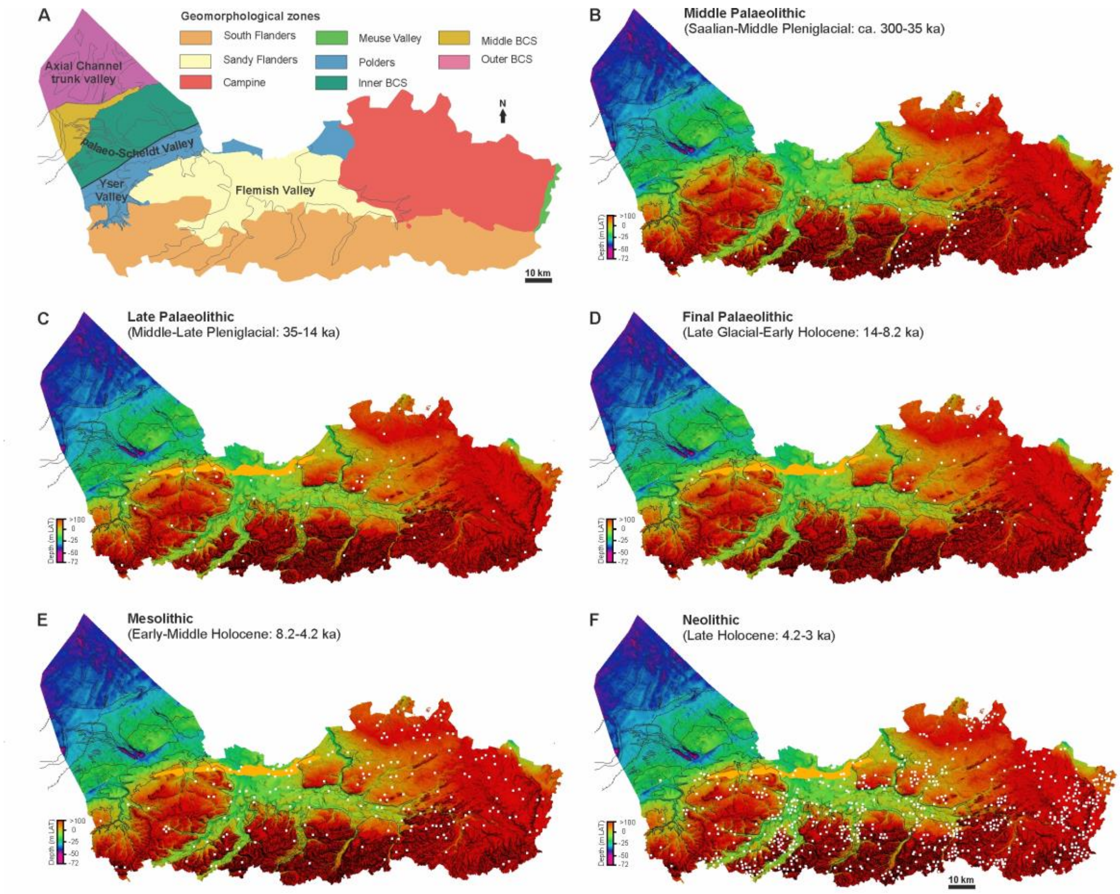


Figure 7.2 Overview of archived archaeological sites in Flanders in relationship to the geomorphological landscapes and the Paleogene topography according to the Central Archaeological Inventory (www.onderzoeksbalans.be) for the Middle Palaeolithic to the Neolithic. The orange area represents the Maldegem-Stekene cover sand ridge that blocked the palaeo-Scheldt River ca. 17–15.5 ka ago (Chapter 5).

In the past decades large quantities of fossil bones were recovered in the Flemish Valley (Figure 7.1). Germonpré (1993b) analysed the bone material and identified two basic Late Pleistocene faunas: a cold-adapted mammoth and a warm-adapted elephant fauna (Figure 7.3). Each fauna may deviate from its standard species composition as certain species show adaptations towards climate change. Next to these two climate-dependant faunas also climate independent species are present called the background fauna (cave lion, hyena, bear, red deer and wild horse: von Koenigswald, 2007). The Holocene fauna is unlike any previous interglacial ‘elephant’ fauna. The continental-scale losses of large mammal populations during the Late Pleistocene-Holocene transition took place over millennial-scale time periods in response to pre-modern anthropogenic environmental pressures and technologies (Crees et al., 2016) in combination with climate change (Stanturf et al., 2006). By the Early Holocene (ca. 9 ka), the forests of northwest Europe had been recolonized by species such as red and roe deer, but also by moose, aurochs, bear, wolf, lynx, beaver, and others (Aaris-Sørensen, 1998; Yalden, 1999). However, the species richness was poor compared to previous interglacials, specifically in large herbivores and carnivores (e.g. no straight-tusked elephant, rhinoceros, hippo, cave lion).

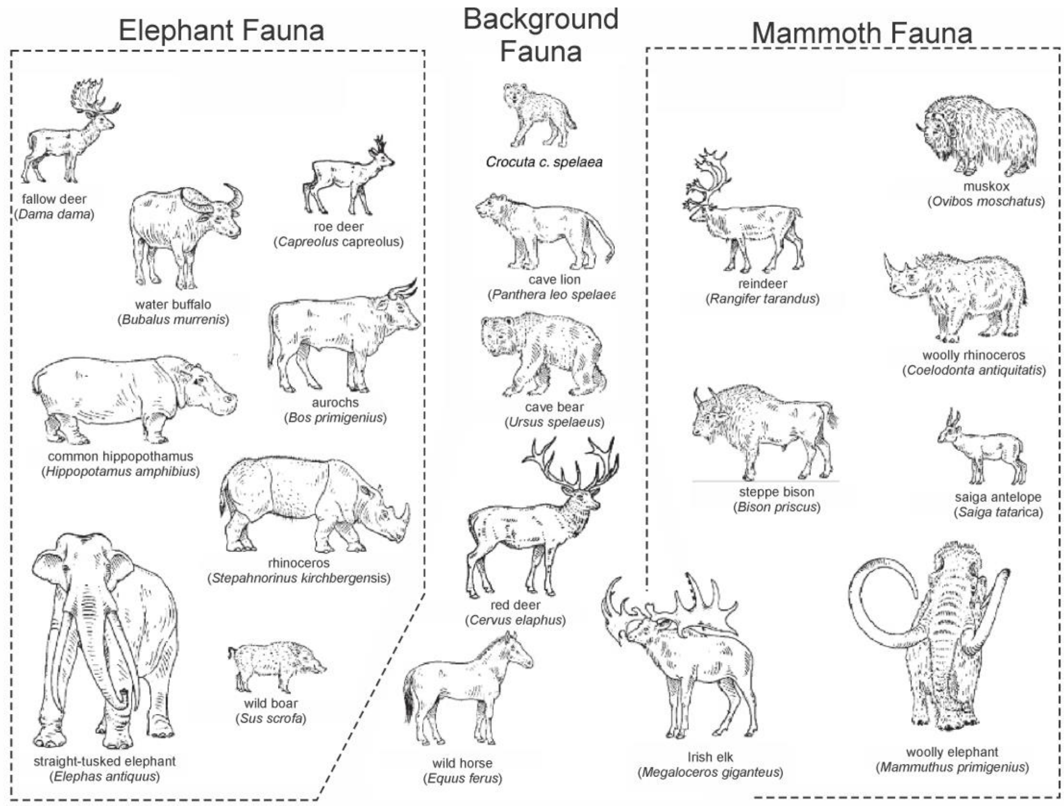


Figure 7.3 The typical fauna encountered in northwest and central Europe according to von Koenigswald (2002). During interglacials a typical elephant fauna dominates, while during glacials a mammoth fauna replaces it. A background fauna is always present during glacial-interglacial cycles.

The unit-descriptive framework and geomorphological descriptions of landforms of Chapter 4 to 6 demonstrate that the BCS shows a strong affinity with Flanders (e.g. Cohen et al., 2014), such as similar landforms (platforms, escarpments, hills) and substratum (Paleogene and Neogene clays, sands and silts), but that they also share an incised-valley system (e.g. the

Flemish-palaeo-Scheldt Valley system). The results from Chapter 4 to 6 are therefore used as the starting point for assessing and mapping the offshore preservation potential of archaeological and palaeontological material.

For the offshore and intertidal palaeontological finds Vermeersch et al. (2015) provided a starting point overview. Their list of find locations and age determination will be expanded with on-going research performed at the Scheur navigation channel led by the Flanders Marine Institute (VLIZ) and by research performed by Post et al. (2017). The fossils found at each offshore location will be compared to the faunas defined by Germonpré (1993b) in the Flemish Valley (Table 7.2). The faunas will then be linked to the landscape evolution and the stratigraphy of the area in order to provide a better age estimation of the fossils and where they may originate. No archaeological material has been reported from the offshore area.

For each archaeological period we will discuss the formation processes of the BCS landscape and what impact this had on the preservation potential for archaeological material. To determine the preservation rating we apply the system of Ward and Larcombe (2008) whom determined the preservation rating of submerged archaeology based on a first-order geomorphological approach. Their results are summarised in Table 7.1.

Table 7.1 General preservation rating for different landform elements in the southern North Sea, derived from a combination of 1) the probability of occurrence of archaeological and palaeoenvironmental remains and 2) the potential risk to these from natural and anthropogenic processes (from Ward and Larcombe, 2008).

Depositional environment	Primary context	Preservation rating	Secondary context	Preservation rating
		Glacial		
Glaciofluvial	-	-	Human occupation along channel margins. Artefacts incorporated into glaciofluvial and debris-flow deposts	Low
Glaciolacustrine	-	-	Human occupation of lake margins (glacial) and abandoned beaches (post-glacial)	High
Tunnel valley	-	-	Abandoned artefacts within sand and gravel deposits	Medium
		Lake		
Basin	Abandoned artefacts incorporated into muds (possible lake dwellings)	Medium-high	-	-
Shoreline	Human occupation of lake margins Gravelly sediments	Medium-high	-	-
		Fenland		
Fenland	Human occupation of fenland, preservation in sily-mud/peat	High	-	-

	Possible indicatior of transgression/regression			
		Riverine		
Channel/channel belt	-	-	Sand and gravel associated with bars channel lag and channel fill may preserved material particularly in aggradational systems	Low
Overbank (inc. crevasse splay and abandoned channel)	Sand, silt deposits provide potential for human occupation and preservation	Medium	Abandoned channels with low-energy sediments ideal for preservation. Potential for re-erosion in meandering system, artefacts reworked downstream	Medium-high
Alluvial plain	Sand, gravel	Low	-	-
Floodplain	Human occupation of floodplain with deposits buried in fine sand associated with flood deposits and/or peat deposits in shallow depressions or infilled palaeochannels	High	-	-
Terrace	Many examples of terraces as focus for settlement and preservatin as focus for settlement and preservation of archaeological sites	Medium	-	-
		Estuarine		
Intertidal sand	-	-	Abandoned artefacts only	Low
Intertidal mud	Abandoned artefacts incorporated into silty muds/peat	Medium	-	-
Salt-marsh	Potential for temporary occupation sites	Medium	-	-
Lagoon/barrier	Sheltered environments with low-energy silty-sand/muds and marginal peat	High	-	-
		Open coast		
Sandy beach	Significant reworking	Low	-	-
Dune ridge	Potential for indicating transgression/regression phases, preservation of underlying deposits	Medium-high	-	-
Tidal channel	Significant reworking	Low	-	-
Tidal sand bar	Significant reworking	Low	-	-

Discussing the landscape evolution and indicating areas suitable for human occupation helps assessing the preservation rating of the stratigraphical units and geomorphological features. This means that areas where no sediments are preserved are indicated as possible locations for human occupation (i.e. on geomorphic grounds). The preservation rating (low, medium or high) is meant as a quality indicator for the suitability of the sedimentary unit or geomorphological feature to preserve archaeological material. The context of archaeological material within sedimentary units is expressed as primary or secondary, while those on geomorphological grounds are expressed as secondary or tertiary. A primary context refers to material preserved in its original location after burial, while a secondary context refers to material that was reworked from its primary burial location, transported and than buried at another location. A tertiary context refers to material that has been removed from its secondary context to a new location through marine transgression (Dix et al., 2008).

7.3 Setting and preserved sediments of the Belgian Continental Shelf

This section provides a brief overview of the late Middle and Late Pleistocene geological history and the preserved sediments of the BCS (see Chapters 5 and 6). This information is used to assess the preservation potential of the preserved depositional environments on the BCS.

The BCS is situated between the depositional southern North Sea Basin and the erosional Dover Strait area: a critical region to link both areas by tracing of depositional and erosional stratigraphical contacts. The relatively low accommodation space outside incised-valley systems resulted in a relatively thin and fragmented sediment cover on top of the pre-Quaternary surface (Mathys, 2009; Chapters 5 and 6). The Paleogene substratum strongly influenced the Pleistocene landscape evolution of the BCS and can be divided into three separate geomorphological regions separated from each other by relatively gentle escarpments: the Inner, Middle and Outer BCS (Chapters 5 and 6; Figure 7.1). The Inner BCS is characterised by the incised-valley system of the palaeo-Scheldt River that extends from the Flemish Valley to the Middle BCS. The Outer BCS is carved by north to south-southwest elongated palaeodepressions that form an extension of the Axial Channel trunk valley system (Chapter 6). The latter formed the major drainage artery of northwest Europe during glaciations when ice-sheets merged across the central North Sea (García-Moreno, 2017). The Middle BCS was influenced by both these river systems during glacial-interglacial cycles (Chapters 5 and 6). An overview of the depositional units and when they were formed is provided in Figure 7.2, where a correlation scheme is provided with the northwest European chronostratigraphy and the archaeological timescale.

Up to 160 ka ago (Late Saalian) the rivers of the Flemish Valley flowed in a northern direction, possibly towards the Meuse River or a proglacial lake (Chapter 5). During the Late Saalian ice mass expansion across the North Sea resulted in glacio-isostatic upwarping and the formation of a topographic ridge between East Anglia and Belgium extending farther into the continent. This resulted in the deflection of river systems and the formation of the palaeo-Scheldt Valley system on the Inner BCS. The merged ice sheet across the North Sea and the topographic ridge to the south pounded all meltwater forming a proglacial lake (Gibbard, 1995; Hijma et al., 2012; Murton and Murton, 2012; Toucanne et al., 2009) . Approximately 150 ka ago the topographic ridge was ‘breached’ and resulted in a GLOF in the direction of the Dover Strait through the Axial Channel trunk valley (Cohen et al., 2014; Gibbard and Cohen, 2015; Hijma et al., 2012). As a result up to 10 m of the pre-Quaternary sediments of the Outer BCS were removed north of the Offshore Scarp (Hijma et al., 2012). At its southern edge the Meuse River entered this Axial Channel trunk valley and its boundary is defined by the edge of the Offshore Scarp. The only sediments from this period are deeply buried in the palaeo-Scheldt Valley (fluvial unit SCH1) and their deposition may have continued into the Early Eemian interglacial.

On-going glacio-isostatic relaxation following Saalian ice-sheet retreat resulted in a fast marine transgression of the BCS that lagged behind global eustatic sea level. During this climate amelioration rivers change their style from braided to meandering. Soil-formation processes intensified and erosion and reworking were less widespread. Few interglacial terrestrial deposits survive in the long term as they are easily and preferentially eroded during subsequent glacials, making interglacial (and interstadial) palaeolandscapes and in situ archaeological remnants rare in comparison to their periglacial counterparts, but exceptions exist (Cohen et al., 2014). Sediments from this transgression are preserved near the Belgian-Dutch border north and south of the Offshore Scarp as shallow marine (subunit BCS1a) and fluvial to estuarine and shallow marine deposits (subunit BCS1b) and are equivalent to deposits found on the DCS.

Other deposits from the Eemian are found in the palaeo-Scheldt Valley and consist of a succession of tidal-influenced, estuarine and shallow marine sediments (units SCH2 to SCH4). Eemian high-stand was achieved very late, possibly somewhere around 110–115 ka. When sea level dropped during the Weichselian Early Glacial incision occurred in the palaeo-Scheldt Valley system and fluvial sediments were deposited (likely during the Rederstall stadial when sea level dropped to levels below the Middle BCS). Other Weichselian deposits on the BCS are found on the Outer BCS and consist of fluvial sediments and cobbles and boulders deposited during GLOF events (units BCS2 and BCS3).

Possibly more Pleistocene sediments are preserved on the BCS, specifically on the extreme eastern and western areas of the Inner BCS. These areas have not yet received much attention and need further geological investigation. This is mainly the result of a combination of shallow gas in the subsurface (eastern nearshore zone) and shallow depths (western nearshore zone).

The gas obliterates the seismic signal not allowing penetration below the gas horizon, while shallow depths result in a strong multiple that often overlaps with the layers of interest. In the east, in the framework of Complex Project Kustvisie, a large geotechnical campaign near Zeebrugge is planned for spring/summer 2018, which will provide crucial information on preserved Quaternary sediments and the landscape evolution of this area.

7.4 Palaeontological finds of the Flemish Valley

Within the Flemish Valley, large quantities of mammalian remains have been found in several localities. The Flemish Valley was first established during the Middle Pleistocene and has since then experienced a complex infilling history (Heyse and Demoulin, 2018). Several rich assemblages have been uncovered and studied in detail, all located on or close to the central axis of the Flemish Valley (Figure 7.1: Rotselaar, Hofstade, Zemst, Dendermonde, Overmere and Oudenaarde; Gautier, 1985, 1995; Gautier and Schumann, 1973; Germonpré, 1993a, 1993b; Germonpré et al., 1993; Germonpré and Ervynck, 1978). All six sites contain Pleistocene assemblages that date from the Weichselian (mammoth fauna) and were recovered from fluvial deposits linked to braided river systems. Two of these sites (Zemst and Overmere) also contain mammalian remains from the temperate Eemian interglacial (elephant fauna). An overview of the mammoth and elephant faunas in the Flemish Valley is presented in Table 7.2.

Two distinct faunas have been identified within the Flemish Valley: a mammoth and elephant fauna. The mammoth fauna is typical for cold glacial periods such as the Weichselian, while the elephant fauna is typical for temperate interglacial periods such as the Eemian (Figure 7.4).

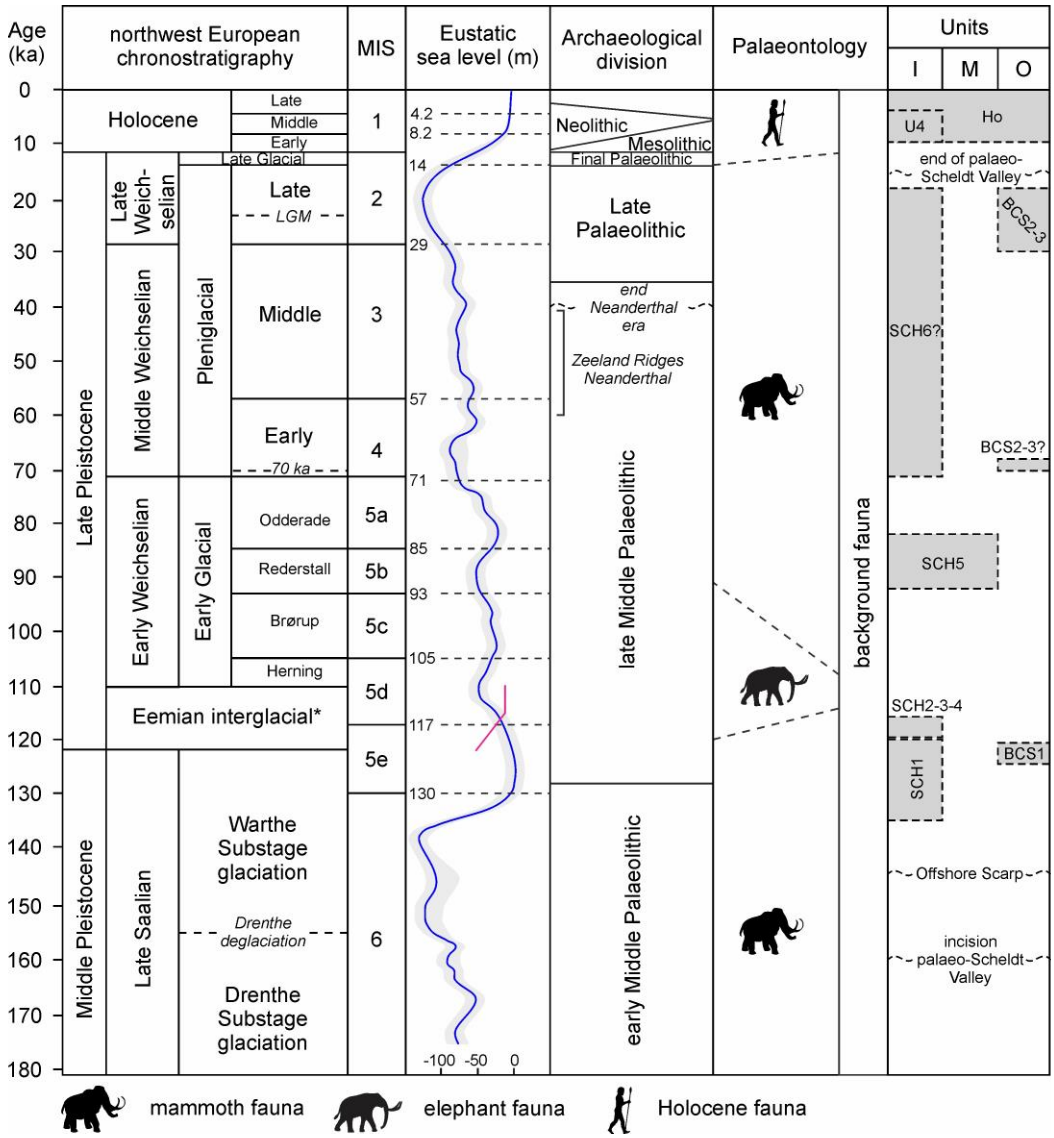


Figure 7.4 Overview of the northwest European chronostratigraphic subdivision of the late Middle and Late Pleistocene correlated to the marine isotope stage (MIS) division (after Cohen and Gibbard, 2016) and architectural units. The eustatic sea-level curve (black line) and associated confidence interval (light grey area) is from Medina-Elizalde (2013). The local Eemian relative sea-level curve for Belgium (red line) is from De Clercq et al., (2018). *Recently, new insights from Sier et al. (2015) placed the onset of the Eemian in northwest Europe at ca. 121 ka, thereby post-dating the MIS 6/5e transition by ca. 10 ka. I: Inner BCS; M: Middle BCS; O: Outer BCS.

7.5 Archaeological finds/sites of Flanders

The archaeological record of Flanders is best explained through its different geomorphologic regions (De Bie et al., 2008; Van Gils et al., 2010; Vanmontfort, 2011; Figure 7.2). These geomorphologic regions consist of the sandy loam area of South Flanders, the sandy areas of the Campine Area and Sandy Flanders, the Flemish and Meuse Valleys and the Polders in the coastal plain and the present Scheldt River. Across these different terrestrial and fluvial landforms various taphonomic scenarios occurred that are related to burial depth, lithology, topography, erosion, and hydrography. These processes resulted in different preservation conditions across these regions since the Palaeolithic period. Here we provide a short overview of these landscapes. For a detailed overview, the reader is referred to the website of Flanders Heritage Agency:

https://onderzoeksbalans.onroerenderfgoed.be/onderzoeksbalans/archeologie.

7.5.1 South Flanders

The earliest sites in this region date from the Middle Palaeolithic and are mostly located along the eastern tributaries of the Flemish Valley (Figure 7.2 B). From the Late Palaeolithic to the Mesolithic sites are rare and mostly concentrated along the edges of the Flemish Valley tributaries (Figure 7.2 C–E). From the Neolithic onwards this location contains one of the highest site densities in the whole region of Flanders (Figure 7.2 F). A large part of the Neolithic sites are located on the Paleogene hills between the Flemish Valley tributaries, however sites or finds are also known from within the tributary valleys.

The loess-derived soils from South Flanders belong to the most erodible soils in the world (Poesen, 1993). As a result the hilly landscape of South Flanders experienced large-scale erosion of its hilltops and slopes resulting in thick colluvium deposits at their bases. Most archaeological finds from this area are therefore known from a secondary context (Moreau, 1986), though the recent discovery of a Final Palaeolithic site in primary context offers new perspectives (Dijkstra et al., 2007). In contrast to Palaeolithic and Mesolithic sites those dating from the Neolithic comprise larger sites and artefacts (e.g. floor plans and pits for waste and preservation) that offer a better preservation potential (Vanmontfort, 2011).

7.5.2 Sandy Flanders and the Campine Area

The combined area of Sandy Flanders and the Campine Area comprise respectively a large part of the incised Flemish valley system and a platform located northeast of it. The oldest sites date from the Middle Palaeolithic of which most are located in the Campine Area. Here, site distribution remains fairly constant up to the Final Palaeolithic. For Sandy Flanders an increase in find/site concentrations is observed during the Late Palaeolithic but is followed by a decline for the Final Palaeolithic. Most sites are concentrated on the edges of the Flemish Valley system. Other finds/sites are located within the valley system possibly located close to active rivers. From the Mesolithic onwards the find/site concentrations increase.

The sandy soils of Sandy Flanders and the Campine Area are located on the higher Paleogene substratum. The Campine Area is located away from major valley systems. This setting resulted in a relatively flat landscape that has experienced high aeolian erosion and relatively little water erosion and water-related sedimentation (De Bie et al., 2008; Van Gils et al., 2010; Vanmontfort, 2011). Within these boundary conditions archaeological sites since the Palaeolithic are buried close to the surface and are easily accessible to research. On the other hand this shallow position makes them vulnerable to post-depositional processes such as bioturbation or anthropogenic activities. An unforeseen side-effect of such shallow burial depths is the mixing of archaeological material from different periods (Van Noten, 1978). The dry and acid soils of this landscape make the preservation of organic materials impossible. This makes the dating of these sites virtually impossible (De Bie et al., 2008). The southern edge of the Campine Area, a terrace system, offers good preservation conditions (Van Peer et al., 1984).

7.5.3 Meuse and Flemish Valleys

The oldest archaeological site in Flanders may date to 300 ka ago and defines the boundary of the Lower-Middle Palaeolithic and is located in the Meuse Valley (Van Baelen et al., 2007). Most sites in the area are located near the edge of the valley with the Campine Area. Find/site distributions are low through the Palaeolithic and Mesolithic and increase exponentially during the Neolithic.

In the Meuse Valley Late Glacial dunes, riverbanks and levees offer high preservation potentials (De Bie and Caspar, 2000). Good preservation here depends on post-depositional aeolian and alluvial sedimentation and recent use of the land.

The Flemish Valley is located in South and Sandy Flanders and the southeast edge of the Campine Area. Find/site distribution within the valley is quite rare in the Palaeolithic compared to the Mesolithic and Neolithic period. This may be caused by a deeper burial depth of these sites (see Heyse and Demoulin, 2018 for an overview of the infill history and possible depths of these contemporary finds/sites). The edges of the morphological features of the Flemish Valley are also known for numerous archaeological discoveries (Figure 7.2 B–F; Cauwe et al., 2001; Crombé and Van Der Haegen, 1994; Van Peer and Smith, 1990). Many of these areas experienced differential post-depositional processes and offer a wide range in preservation potential (i.e. primary and secondary). The best preservation conditions in primary context are preserved at the base of the sandy infill and for areas where thick loess sequences were deposited in the east and southwest areas of the Flemish Valley. Elsewhere archaeological remains are known from secondary context and are the result of post-depositional erosive processes, such as fluvial incision.

7.5.4 Polders

The Polders are situated in the present coastal plain and in the north along the present Scheldt River. These areas offer good preservation conditions that is the result of post-depositional peat formation and alluvial to perimarine clays. Most often these archaeological sites are buried at greater depths beneath the surface and they are discovered as chance finds during industrial activities. This is for example the case in the Harbour of Antwerp where Final Palaeolithic sites where discovered (Crombé et al., 1999, 2000; Perdaen and Ryssaert, 2002).

7.6 Palaeontological finds of the BCS

In this section we present an overview of the presently known paleontological find locations with remains from mammalian terrestrial and marine faunas that have been reported with certainty as coming from the intertidal or subtidal part of the BCS (Table 7.2; Figure 7.5). This overview is largely based on intensive inventory work done by Flanders Heritage Agency (Demerre et al., 2008) and on an unpublished report from the SeArch project (Vermeersch et al., 2015), supplemented with additional information from www.fossiel.net (a knowledge base and communication platform providing information about geology and fossils to amateurs and the general public) as well as from on-going research at the Flanders Marine Institute. In total eight find locations are reported for the BCS and the intertidal zone: seven of these are situated within the wider palaeo-Scheldt Valley system, while one belongs to the offshore Yser Valley system located in the western part of the BCS.

7.6.1 palaeo-Scheldt Valley system

7.6.1.1 Blankenberge (Figure 7.5, nr. 1)

Blankenberge lies on the boundary between the incised palaeo-Scheldt Valley and the Central Marginal Platform (De Clercq et al., 2016). The local Quaternary stratigraphy consists of Late Eemian shallow marine sediments of subunit SCH4a overlain by Holocene sediments (Chapter 5).

Here, a single broken humerus of an aurochs was found as a stray beach find in a secondary or tertiary context (Figure 7.6 A). Expert analysis demonstrated that this bone fragment belongs to a young aurochs, younger than 4 year, and could be interpreted as a human worked bone, as the humerus was broken on the right spot to extract the marrow (Vermeersch et al., 2015). Two successive trials for ^{14}C dating did not produce a reliable result. However, the bone is estimated to belong to the Weichselian-Early Holocene period (110–8 ka; Vermeersch et al., 2015).

7.6.1.2 Ostend (Figure 7.5, nr. 2)

Ostend is located on the central axis of the palaeo-Scheldt Valley within a bend of the valley system. Pollen analyses from a continuous core 5 km offshore Ostend (core GR1; see Chapter 5, Figure 5.9) shows that at this location a ca. 45 m Late Pleistocene to Holocene sedimentary sequence is preserved. It comprises of Late Saalian to Early Eemian fluvial sediments at its base, overlain by a transgressive sequence of tidal-influenced fluvial, estuarine and shallow marine sediments. On top of this transgressive sequence lies a Late Eemian to Early Glacial fluvial deposit overlain by Preboreal to Subboreal Holocene sediments, with in between, an important stratigraphic hiatus.

From the beach one tusk of a walrus and a bone fragment from ray-finned fish (Figure 7.6 B and C) are known together with six other bone finds from terrestrial mammals (Table 7.2). These include mammoth, giant deer, reindeer, red deer, roe deer (Figure 7.6 D) and boar. The walrus and mammoth have been estimated to date from the Weichselian (110–10.5 ka: Vermeersch et al., 2015). The red deer is dated from the Weichselian-Early Holocene period (110–8 ka: Vermeersch et al., 2015) while the roe deer is thought to date from the Early Holocene (11.5–8 ka: Vermeersch et al., 2015). No age estimation is provided for the ray-finned fish (source: www.fossiel.net) or the boar but they both have been interpreted to have a Pleistocene age. All palaeontological finds are found in a secondary or tertiary context.

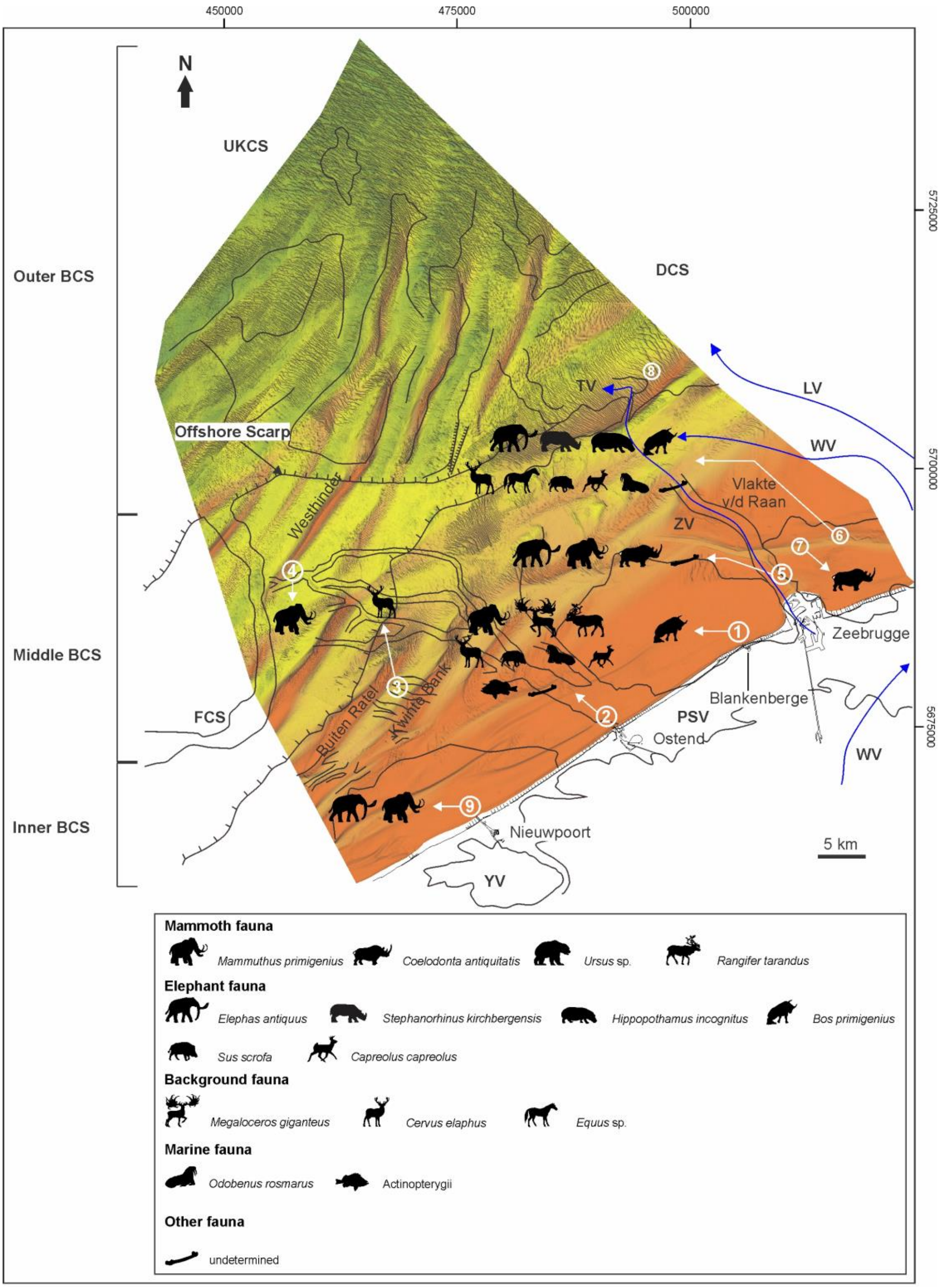


Figure 7.5 Overview of offshore fossil find locations projected on the bathymetry. The find locations are a compilation from Vermeersch et al. (2015), Post et al. (2017), www.fossiel.net, and on-going research at the Scheur navigation channel (Flemish Marine Institute). For abbreviations the reader is referred to Figure 7.1.

7.6.1.3 Kwintebank (Figure 7.5, nr. 3)

The Kwintebank is a tidal sandbank located downstream the palaeo-Scheldt Valley. The core of the sandbank consists of Early Holocene back-barrier sediments (Mathys, 2009). Underlying

Pleistocene sediments are of Late Eemian marine origin. At this location a red deer antler was found during sand extraction activities (Figure 7.6 E; Wessex Archaeology, 2008). There are no tool marks on the antler and it seems to have been shed naturally. The combination of natural shedding and extraction make it unclear if the antler was found in a primary context or not. Analysis by Vermeersch et al. (2015) dates the find to the Early Holocene period.

7.6.1.4 Westhinder (Figure 7.5, nr. 4)

The Westhinder is a tidal sandbank that extends from the Middle to the Outer BCS. On the Outer BCS the core of the sandbank is composed of Pleniglacial fluvial sediments. At its southwest end, on the Middle BCS, the tidal sandbank lies on tope of the Paleogene substratum. Here, a mammoth skull of Weichselian age was dredged (Vermeersch et al., 2015). The skull was remarkable due to its association with puparia of the subarctic blowfly (*Protophormia terraenovae*) that were found within the skull and the horns (Gautier, 1995, 1998).

Figure 7.6 A) Humerus of aurochs from Blankenberge, B) walrus tusk from Ostend, C) vertebrae from Actinopterygii from Ostend, D) vertebra of roe deer from Ostend, antler from red deer from the Kwintebank, F) sternum vertebra from mammoth or elephant from the Zeebrugge navigation channel, G) vertebrae from Archaeocetes from Scheur, H) shark tooth from *Carcharocles* sp., I) piece of upperjaw with tusk and teeth and lumbar vertebra from walrus from Scheur.

7.6.2 Zeebrugge-Waardamme-Lys Valley

7.6.2.1 Zeebrugge navigation channel (Figure 7.5, nr. 5)

The navigation channel towards the Zeebrugge harbour is regularly dredged (to depths of 6 m). These maintenance activities have most likely exposed Late Pleistocene and Paleogene strata but the precise stratigraphy remains unknown and as a consequence several fossil discoveries from mammoth are reported from the area. A scapulae and sternum from the navigation channel are determined to come from either mammoth or straight-tusked elephant (Figure 7.6 F). This suggests that the bones may have had a primary context, through a secondary or tertiary context cannot be excluded. Preliminary age estimations indicate the uncertain mammoth/elephant material date to the Eemian-Weichselian period while the mammoth dates from the Weichselian period (Vermeersch et al., 2015).

7.6.2.2 Scheur navigation channel (Figure 7.5, nr. 6)

The Scheur is a navigation channel directed to the Westerscheldt. Maintenance dredging of the navigation channel (depths of 10 to 12 m) has exposed older Late Pleistocene and even Paleogene strata. Here stratigraphic information is poor due to local presence of shallow gas prohibiting seismic penetration. A part of the Late Pleistocene layers may belong to the Eem Formation (Du Four et al., 2006; Ebbing et al., 1992) while the Paleogene strata are correlated with the Eocene Maldegem and Zelzate Formations (Le Bot et al., 2003). Dutch fishing activities over the past few decades have uncovered large amounts of mammalian bone material (see Post et al., 2017). This observation only recently came to attention of Belgian scientists. As a result two targeted surveys in July 2017 and a recent survey in January 2018 produced more finds and made clear that remains from four fairly distinct faunas from different geological periods are present at the Scheur: 1) a marine Eocene/Early Oligocene fauna (primitive whale and possibly also shark: Figure 7.6 G and H; Post et al., 2017; pers. comm. B. Langeveld); 2) a Late Pleistocene interglacial fauna (wild horse, aurochs, red deer, roe deer, rhinoceros, boar and hippo); a Late Pleistocene estuarine/marine fauna composed of only walrus (Figure 7.6 I); and 4) a Late Pleistocene glacial fauna (bear; wild horse; red deer). Based on the current finds the site appears to be one of the largest and southernmost find locations of walrus bone material in the world (pers. comm. K. Post and D. Mol). The bone material is most likely preserved *in situ* as the fragile teeth of the walrus are found within the jawbone, derived from layers exposed at the seafloor (pers. comm. K. Post). A second indicator for *in situ* preservation (primary context) is the presence of a clean side and an overgrown side (by active fauna and flora). Determination and dating of the fossils are still on going at this moment. Further research of the site, such as coring and additional seismic, is planned for May 2018 and is coordinated by the Flanders Marine Institute.

7.6.2.3 Zeebrugge shipwreck (Figure 7.5, nr. 7)

The Zeebrugge shipwreck is an early 16th century wreck located several kilometres northeast of the Zeebrugge harbour and lies in between the Zeebrugge and Scheur navigation channels. From this location hundreds of archaeological finds are known. Geomorphologically the wreck is positioned between the Zeebrugge and Waardamme Valleys. From this location remains from woolly rhinoceros are known that are believed to date from the Weichselian period (Vermeersch et al., 2015). Not much is known about the local geology similar to the Scheur find location. However, it can be assumed that based on its close proximity a similar stratigraphy is to be expected.

7.6.2.4 Thorntonbank (Figure 7.5, nr. 8)

Van Kolfschoten and Van Essen (2004) mention fossil finds from the Early Pleistocene terrestrial association I, associated with the Tiglian (ca. 2.4–1.8 Ma), found close to the Thornton sandbank. This tidal sandbank stretches across the BCS and the DCS suggesting these fossils could also originate from Dutch waters as no stratigraphic association of this age is found on the BCS (Mathys, 2009; Chapters 5 and 6). The authors mention elephant, southern elephant, bush-antlered deer and horse. Mol (2008) also mentions a molar from an Early Pleistocene elephant found next to the Thornton sandbank at the boundary between the BCS and DCS.

7.6.3 Yser Valley system

7.6.3.1 Nieuwpoort (Figure 7.5, nr. 9)

Nieuwpoort is located in the western part of the Belgian coast where the Yser Valley enters the North Sea. From the navigation channel towards the harbour an Eemian-Weichselian molar of a mammoth or elephant was dredged up. The offshore stratigraphy of this location is not fully understood and is the combined result of strong multiples in seismic data and locally also shallow gas in the subsurface. The onshore Holocene geological evolution has been well studied over the past few decades (Baeteman, 1981, 1999, 2005a, 2008; Baeteman and Declercq, 2002; Bertrand and Baeteman, 2005; Denys and Baeteman, 1995), the Pleistocene sequences of the valley infill have not received much attention until recently. Here, Bogemans et al. (2016)

determined that the pre-Holocene infill of the valley near Nieuwpoort contains sediments belonging to the Saalian, Eemian and Weichselian periods (Figure 7.7).

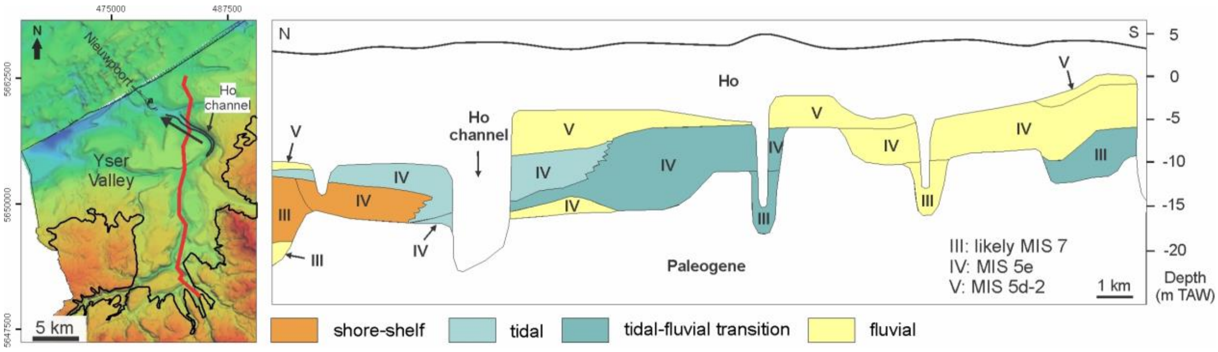


Figure 7.7 North-south geological cross-section from the Yser Valley located in the western coastal plain (modified from Bogemans et al., 2016).

7.7 Landscape context

Table 7.2 clearly indicates that the find locations off the Belgian coast have a much lower concentration in bone material and species compared to the onshore Flemish Valley sites but show a similar bias towards larger species (cf. Gautier and Schumann, 1973; Germonpré, 1993b). This may be caused by the secondary or even tertiary contexts in which most of the BCS bone finds are encountered. As a result a clear correlation with the local stratigraphy is often difficult, preventing a reliable age determination. However, with the available stratigraphic information from literature tentative similarities can be drawn with the identified faunas of the Flemish Valley. Here, the offshore find locations will be discussed in relationship to the geomorphology of the Paleogene substratum as this surface contains the landforms that have persisted to the Late Pleistocene period. Finally, the identification of the Flemish Valley faunas can help assess which species can be expected at various find locations on the BCS based on the present stratigraphic framework.

7.7.1 Palaeo-Scheldt Valley

The main branch of the palaeo-Scheldt Valley comprises four find locations: from upstream to downstream the find locations are Blankenberge, Ostend, Kwintebank and Westhinder (Figure 7.5).

7.7.1.1 Blankenberge (Figure 7.9, nr. 1)

The single aurochs fossil at Blankenberge is difficult to associate with any known fauna association from the Flemish Valley. The Flemish Valley fauna associations (Table 7.2) show that aurochs occur in both mammoth and elephant faunas ranging from the Eemian to the Early Holocene, but that they prefer more temperate conditions (Germonpré, 1993b).

Palaeogeographic reconstructions of the Eemian landscape show that the area of Blankenberge belonged to a rapidly drowning landscape (as a result of post-Saalian glacio-isostatic relaxation) and was positioned on the northern edge of the Flemish Valley, more precise on the Central Marginal Platform (CMP Figure 7.9 A–D). With rising sea level this Paleogene platform transformed into an island of decreasing size. Cut off from the mainland, such an island was likely unable to sustain large mammals, such as aurochs. Furthermore the availability of freshwater must have decreased during the marine transgression. During the Early Glacial (Figure 7.9 E), when climate was still not too cold, the area was located again on the northwest edge of the palaeo-Scheldt River (a situation similar to the Early Eemian). During the Pleniglacial climate was likely too cold for aurochs (Figure 7.9 F–G; see also Table 7.2). During the climate improvement of the Holocene aurochs entered the area once more (Montgomery et al., 2014; Figure 7.9 H).

The aurochs bone from Blankenberge was a beach find in secondary or tertiary context. This suggests that the bone likely originates from layers exposed at the seafloor not far away from its find location. Onshore no fluvial Eemian sediments are recognised at this location and the Eemian and Holocene marine sediments are buried beneath Holocene back-barrier sediments (Figure 7.8 A and D). More so these marine sediments are unlikely to have hosted aurochs unless they have been reworked during the Eemian and/or Holocene transgression. Our best guess is that the aurochs find could be related to the Early Holocene as part of a Holocene fauna (Aaris-Sørensen, 1998; Yalden, 1999). This corroborates the Weichselian-Early Holocene age determination of Vermeersch et al. (2015), though based on the fauna assemblages of Table 7.2 and the local stratigraphy an Early Holocene age seems more likely. Additionally, if the interpretation as human reworked bone is correct the find indicates the nearby presence of Mesolithic human activity.

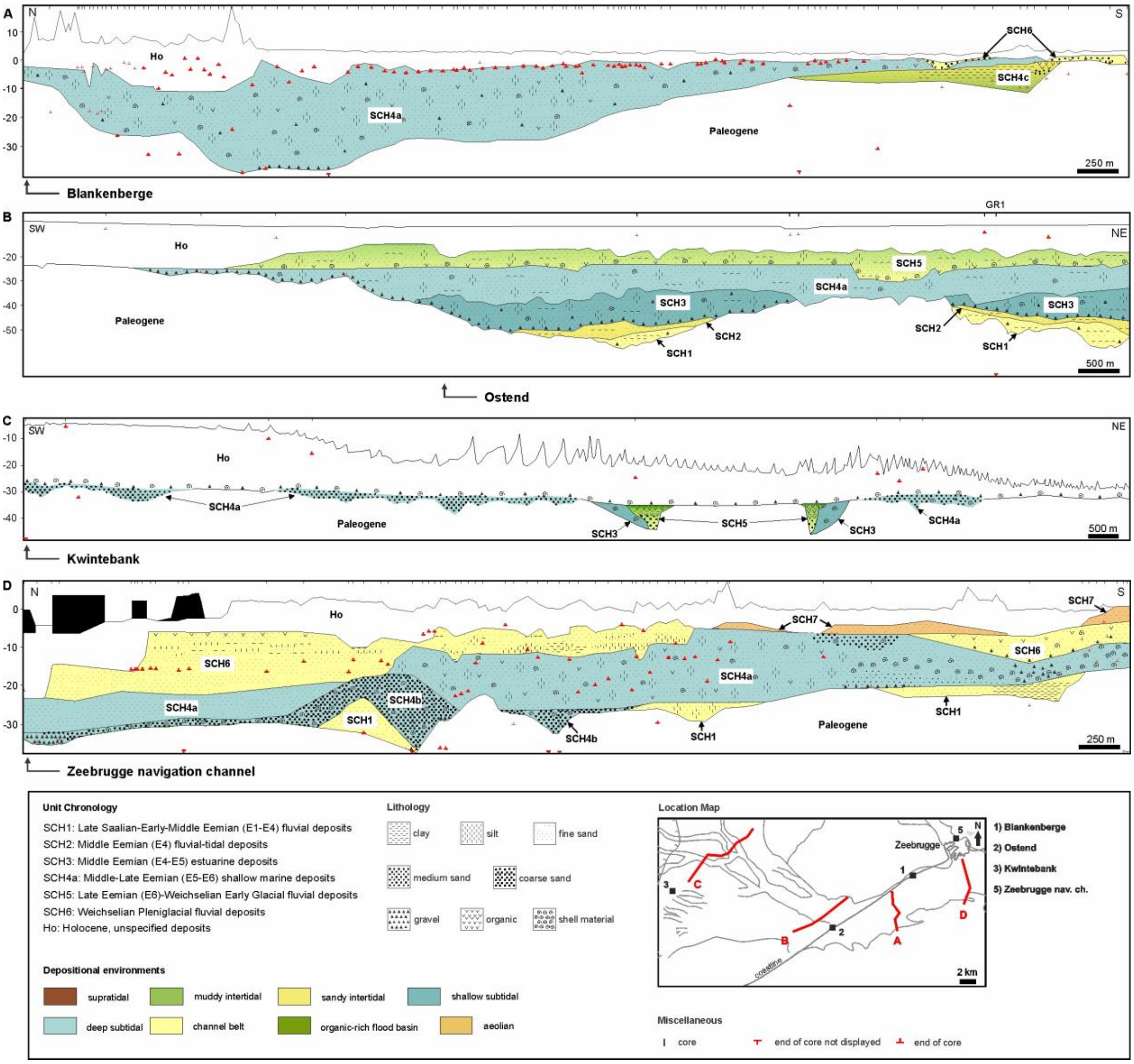


Figure 7.8 Geological cross-sections from the palaeo-Scheldt Valley locate in the eastern coastal plain and Inner BCS (modified from De Clercq et al., 2018).

7.7.1.2 Ostend (Figure 7.9, nr. 2)

The species composition at Ostend is by far the largest encountered in the palaeo-Scheldt Valley and comprises both a marine and elephant fauna (Table 7.2), where the marine fauna comprises walrus and remnants of ray-finned fish.

7.7.1.2.1 Walrus

Present-day walrus inhabit the coastlines of the North Pole region. For the study area this observation creates a paradox as conventional literature states that arctic conditions in this area were achieved not earlier than the Early Glacial (Helmens, 2014) when sea level drops drastically and the coastline is far beyond the reach of Ostend.

Stratigraphic analysis shows that the marine sediments at this location date from the Late Eemian-Early Glacial and Holocene (Figure 7.8 B; see Chapter 5). It is unlikely that the walrus tusk is related to the Early Holocene period as the palaeoshoreline already reached Ostend somewhere between 9 and 8 ka (Mathys, 2009) and climate conditions rapidly improved after the Last Glacial in our regions (Missiaen et al., 2016). This seems to suggest that the find originates from the Late Eemian-Early Glacial. The palaeogeographic map series of Chapter 5 suggest that the Eemian transgression of the BCS took place when global eustatic sea level was already falling (see Figure 5.11 Chapter 5) and that high-stand conditions were achieved between 115 and 110 ka. According to the marine isotope chronostratigraphy this high-stand takes place at the transition of MIS 5e–d when climate was already deteriorating. Recently, Vansteenberge et al. (2016) demonstrated that between 120 and 117.3 ka climate started to deteriorate in northwest Europe and that these colder conditions were rapidly followed by stadial-interstadial conditions between 117.3 and 97 ka. It therefore seems likely that the high-stand conditions of the Eemian interglacial occurred during these deteriorating or even colder climate conditions of MIS 5e–d. Within this timeframe the palaeo-Scheldt Valley was a rapidly drowning estuary that transformed into a shallow seaway (Figure 7.9 B–D). This period is represented in the stratigraphy by subunit SCH4a (Chapter 5). These events are in accordance with the stratigraphic evidence suggesting that the walrus tooth originates from the shallow marine subunit SCH4a. However, at Ostend this unit is buried 20–30 m beneath the surface indicating a secondary or tertiary context. Eemian marine sediments may have been removed from their original location during the Holocene transgression of the area.

7.7.1.2.2 Ray-finned fish

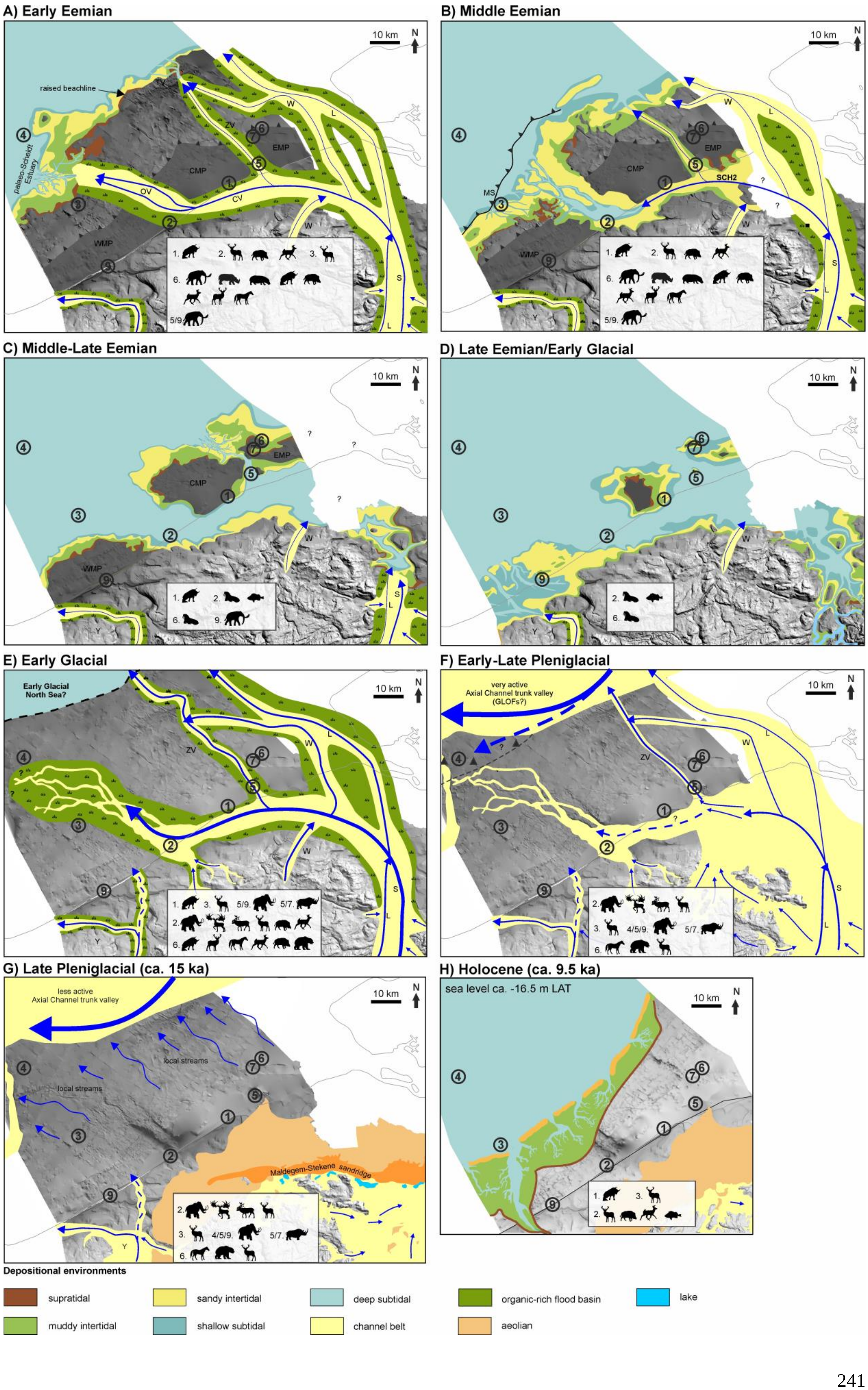
Ray-finned fish occur in all aquatic environments ranging from marine to freshwater. It is unclear if the find originates from the Late Eemian shallow marine subunit SCH4a, the Early Glacial fluvial unit SCH5 or even from Holocene fluvial to marine environments (unit Ho). The bones were found in a secondary or tertiary context suggesting they were reworked.

7.7.1.2.3 Elephant fauna

The remaining bone material from Ostend comprises the mammalian species mammoth, giant deer, reindeer, red deer, roe deer and boar. The presence of mammoth and reindeer may typify a typical mammoth fauna (Table 7.2) as identified in the Flemish Valley. The presence of red deer, roe deer and boar suggest milder climate conditions (Germonpré, 1993b) and shows large affinities with the Early Glacial Zemst IIB fauna of the Flemish Valley (Table 7.2). On the other hand, red deer, roe deer and boar can also point towards a Holocene deprived elephant fauna.

The stratigraphy offshore Ostend shows that Early Glacial sediments overlie Holocene back-barrier and marine sediments (Figure 7.8 B). The Early Glacial sediments are preserved between 5 and 10 m depth. The Holocene sediments consist of terrestrial (Preboreal and Boreal) and marine environments (Atlantic and Subboreal: see Figure 5.9, Chapter 5). With the current state of knowledge and the secondary or tertiary context of the finds we have no way of discriminating between both periods. A mammoth fauna comprising mammoth, giant deer and reindeer from the Early Glacial unit SCH5 can be determined with certainty, while red deer, roe deer and boar may originate from the same fauna or from the Holocene fauna. With this interpretation the age determination of Vermeersch et al. (2015) may be better constrained. The mammoth and reindeer likely date from 110–71 ka (instead of 110–10.5 ka), the giant deer dates from the same period (instead of 11.5–8 ka), red deer and roe deer may date from 110–71 ka or 11.7–9.2 ka (instead of 11.5–8 ka). The boar did not have a specific age determination and has the same age as red deer and roe deer.

Figure 7.9 (next page) Palaeogeographic maps from the Early Eemian to the Holocene showing the fossil find locations on the BCS. Each map shows for each find locations, which species may have occurred at that time in the landscape. The map series A-G are modified from De Clercq et al. (2018; cf. Chapter 5), while map H is modified from Mathys (2009).



A) Early Eemian
B) Middle Eemian
C) Middle-Late Eemian
D) Late Eemian/Early Glacial
E) Early Glacial
F) Early-Late Pleniglacial
G) Late Pleniglacial (ca. 15 ka)
H) Holocene (ca. 9.5 ka)
raised beachline
palaeo-Scheldt Estuary
10 km
N
CMP
EMP
WMP
ZV
OV
CV
SCH2
MS
Early Glacial North Sea?
very active Axial Channel trunk valley (GLOFs?)
less active Axial Channel trunk valley
local streams
Maldegem-Stekene sandridge
sea level ca. -16.5 m LAT
Depositional environments
supratidal
sandy intertidal
deep subtidal
organic-rich flood basin
lake
muddy intertidal
shallow subtidal
channel belt
aeolian

7.7.1.3 Kwinte sandbank (Figure 7.9, nr. 3)

The red deer antler recovered from the Kwinte sandbank alone cannot be directly correlated to any fauna association of the Flemish Valley, as it is a species from the background fauna (see Figure 7.4). This means that it may have been present from the Early Eemian to the Holocene when sea level was low enough. However, according to Vermeersch et al. (2015) the antler dates from 110–8 ka, i.e. Early Glacial to Early Holocene. Stratigraphic evidence from the Kwinte sandbank (Figure 7.8 C) shows that only two depositional units are preserved from this location: a shallow marine Late Eemian subunit SCH4a and the Holocene unit Ho. Palaeogeographic reconstructions from the Holocene by Mathys (2009) demonstrate that the Kwinte sandbank area was already inundated by ca. 10 ka. This makes it more likely that the Holocene sediments in this area are of a marine origin. As both units are marine it is very likely that the find was already reworked (secondary context) and that its recovery may even suggest a tertiary context (a Holocene marine reworking from Eemian marine sediments). Hence, the find may originate from Early Eemian to Early Holocene sediments.

7.7.1.4 Westhinder sandbank (Figure 7.9, nr. 4)

The most offshore find location from the palaeo-Scheldt Valley is a skull from a juvenile mammoth south of the Westhinder sandbank on the Middle BCS. Mammoths only occur during glacial periods and the transitioning towards interglacial conditions (Germonpré, 1993b). This means that the mammoth from this location may date from as early as the Late Saalian, when this location was at the very edge of the Axial Channel trunk valley. However, Vermeersch et al. (2015) attributed a Weichselian age (110–10.5 ka) to the find. Palaeolandscape reconstructions of this area of the BCS show that after the Late Saalian fluvial action sea-level cycles affected this location making a Saalian age unlikely (see Chapters 5 and 6). Furthermore, no Late Saalian sediments are preserved at this location.

The palaeo-Scheldt River was also active in this location during the Early Glacial (Figure 7.9 E), most likely during the Rederstall-Odderade interval (93–71 ka), when eustatic sea level was positioned at levels lower than the incised valley system. Fauna associations Zemst IIB, Hofstade I, Overmere II and Dendermonde I of the Flemish Valley (Table 7.2) show that mammoths did indeed occur in this region during the Early Glacial. A second possible date for the mammoth skull is the Early and Late Pleniglacial (ca. 70 ka and 30–18.7 ka), when GLOFs and megafloods were channelled south through the Axial Channel trunk valley (Figure 7.9; Chapter 6). On the BCS evidence of these events are preserved as unit BCS2, unit BCS3 and as far-travelled erratic clasts. The mammoth skull location lies in the projected path of these floods and very close to erratic clast locations that belong to a larger gravel field (Chapter 6, Figure 6.10). Because the skull was found during sand extraction activities this suggests that the gravels may be partly incorporated within the sandbank, most likely at its base as part of a sandbank core. We therefore suggest that the mammoth skull originates from the Early or Late Pleniglacial at 70 ka and 30–18.7 ka when these catastrophic events occurred, although an Early Glacial age (Odderade-Rederstall) cannot be entirely excluded. The geological context would, in both cases, likely be secondary.

7.7.2 Zeebrugge-Waardamme-Lys Valley

The Zeebrugge-Waardamme-Lys Valley system is a side branch of the palaeo-Scheldt Valley system that was active from the Late Saalian to the Late Pleniglacial (ca. 15 ka) (Figure 7.9 A–F). It consists of the Zeebrugge Valley that extends from the Zeebrugge harbour to the Thornton Valley and the Waardamme and Lys Valleys, which are mostly located near or east off the Belgian-Dutch border (Figure 7.9).

7.7.2.1 Zeebrugge navigation channel (Figure 7.9, nr. 5)

The Zeebrugge harbour and its navigation channel are located near the bifurcation of the Zeebrugge Valley. The fossil bones that were uncovered here have mostly remained undetermined (Vermeersch et al., 2015) except for one mammoth and a possible mammoth/elephant find, which make it impossible to correlate with the Flemish Valley faunas of Germonpré (1993b). The finds were discovered as a result of dredging activities of the navigation channel suggesting a primary or secondary context. According to Vermeersch et al. (2015) the finds are dated to the Eemian-Weichselian period (130–10.5 ka).

No stratigraphic evidence exists of the navigation channel itself due to shallow gas in the subsurface. The nearest evidence comes from the onshore part of the palaeo-Scheldt Valley several kilometres from the find location (Figure 7.8 D). Here, a sequence of Late Eemian, Weichselian and Holocene sediments is found. The Late Eemian sediments comprise estuarine sediments deposited within a very-high fluvio-hydraulic energy environment that have reworked all older material (subunit SCH4b). The Weichselian sediments belong to the braided fluvial unit SCH6, while the Holocene sediments are unspecified. Considering the Holocene fauna, where no large mammals are present (Aaris-Sørensen, 1998; Yalden, 1999), this latter unit Ho can be excluded as the possible origin of the mammoth/elephant find. The estuarine unit that includes older reworked terrestrial sediments of the Middle-Late Eemian (Figure 7.9 A–D) seems to be a possible origin but these Eemian deposits are most likely located too deep (between 20 and 30 m below the seafloor: Figure 7.8 D) for any dredging activity to reach. We therefore interpret the mammoth remains to likely originate from unit SCH6 that was deposited during the Pleniglacial (Figure 7.9 F). The uncertain determination of the mammoth/elephant also fits with this interpretation.

7.7.2.2 Scheur navigation channel (Figure 7.9, nr. 6)

The Scheur is an exceptional find location where it is believed that most of the remains are preserved in primary context. Unfortunately no stratigraphic information from this location is available due to gas in the subsurface. The ancient whale and shark fauna from the Lutetian-Bartonian Maldegem Formation (ca. 43–38 Ma) is out of the scope of this study and will not be further discussed here. The numerous walrus fragments suggest that this site once harboured a large walrus colony comprising both youngsters and adults (pers. comm. Klaas Post, Bram Langeveld, Dick Mol). Palaeogeographic reconstructions of the Eemian transgression (Figure 7.9 A–D) suggest that during the Middle-Late and Late Eemian islands with intertidal flats were present at both sides of the Zeebrugge Valley estuary. This suggests that the walrus finds here most likely relate to the contemporaneous walrus tusk found at Ostend.

A large part of the remaining fossils from the Scheur show strong resemblance with the elephant faunas of Zemst A and Overmere I (Table 7.2). Both faunas are associated with the Eemian interglacial. If this is true than this bone material is slightly older than the cold walrus fauna and dates from the Early-Middle Eemian (Figure 7.9 A–B). The presence of bear in association with red reed and wild horse on the other hand suggest an Early Glacial-Late Pleniglacial origin. The terrestrial Early-Middle Eemian and Weichselian mammalian remains suggests that not only marine Eemian sediments are preserved at this location as stated by Ebbing et al. (1992) and Du Four et al. (2006), but likely also fluvial or fluvial-related sediments from these periods and that they have been preserved in very good conditions.

7.7.2.3 Zeebrugge shipwreck site (Figure 7.9, nr. 7)

Between the Zeebrugge and the Scheur navigation channels lies the Zeebrugge shipwreck site. Here, a single find of a woolly rhinoceros is found during diving excavations suggesting the bone was uncovered from the uppermost sediments (secondary or tertiary context?). The find was dated to the Weichselian period (110–10.5 ka) by Vermeersch et al. (2015). Woolly rhinoceros is associated with the cold mammoth faunas of the Flemish Valley described in

Table 7.2. As for the Zeebrugge and Scheur navigation channels, no stratigraphic information is available from this location but the proximity to these find locations suggests a similar stratigraphy may be present here. Moreover, the woolly rhinoceros may be correlated to the mammoth fauna found at the Scheur and Zeebrugge navigation channels. From the Early Glacial to the Late Pleniglacial (ca. 15 ka; Figure 7.9 E-G) the location was situated between the Zeebrugge and Waardamme Valleys. This suggests that the woolly rhinoceros find is likely related to Early Glacial sediments from this valley system and has an approximate age of 110–71 ka, though an Early and/or Late Pleniglacial age (ca. 70 ka and 29–11.7 ka) is also possible. A primary context is likely but a secondary context by fluvial transport cannot be excluded.

7.7.2.4 Thornton sandbank (Figure 7.9, nr. 8)

The fossil finds from the Thornton area were found in secondary or tertiary context. The finds date from the Early Pleistocene period terrestrial fauna association I and are associated with the Winterton Shoal and Ijmuiden Ground Formations (Van Kolfschoten and van Essen, 2004). Geological cross-sections of the Dutch Continental Shelf by Balson et al. (1992) and Hijma et al. (2012) demonstrate that these formations are present on the DCS and that they pinch out ca. 20 km northeast of the Thornton sandbank. On top of these Early Pleistocene units lie the marine Eemian and fluvial Kreftenheye Formations. The Kreftenheye Formation was deposited in the southern North Sea during the Late Pleistocene (Rijsdijk et al., 2005) under primarily glacial conditions (Busschers et al., 2007; Hijma et al., 2012). During this period the Meuse River and later also the Rhine River flowed in a southwest direction in the direction of the so-called Axial Channel trunk valley not far from the present-day Thornton sandbank (Hijma et al., 2012). Sediments of the Kreftenheye Formation suggest that fluvial reworking during the late Middle and Late Pleistocene reworked the Winterton Shoal and IJmuiden Ground sediments and may have incorporated them. The Thornon sandbank find location is not further discussed in the text as its fauna association predates the Late Pleistocene and no landscape information is available from that time.

7.7.3 Yser Valley

7.7.3.1 Nieuwpoort (Figure 7.9, nr. 9)

The mammoth/elephant fossil from Nieuwpoort was found in a secondary or tertiary context from the intertidal area and was dated to the Eemian-Weichselian period (130–10.5 ka). This suggests that the fossil originates from older stratigraphic layers of a glacial or interglacial origin and does not originate from the Holocene fauna. Projection of a geological cross-section from Bogemans et al. (2016) on top of the pre-Quaternary surface map shows that a Holocene-filled channel is directed to Nieuwpoort. This channel cuts through Eemian (MIS 5e) fluvial and coastal to marine sediments, while its top comprises Weichselian (MIS 5d–2) fluvial sediments. It is very likely that the fossil find originates from one of these fluvial environments. However, until a conclusive determination is provided it is impossible to distinguish between the Eemian and Weichselian fluvial environments as a possible source for the mammalian bone.

7.7.4 A taphonomic scenario similar to the Flemish Valley?

The studied mammalian fauna of the BCS shows a mix of marine, mammoth, elephant and Holocene faunas. The mammoth and elephant faunas show a strong resemblance to the faunas identified in the Flemish Valley by Germonpré (1993b). Similar to the Flemish Valley faunas the palaeo-Scheldt Valley faunas appear to be largely influenced by the deteriorating climate conditions of the Weichselian Early Glacial. From this moment onwards, large mammals dominate the fauna as temperate species disappear.

The correlation with the Flemish Valley faunas is not evident, as most of the offshore and intertidal fossil finds have lost their primary geological context. This means that a best estimate can only be provided based on the landscape evolution and the local stratigraphy as done in Section 7.7.1. In most cases an approximation of the geological context could be provided (Table 7.2) and a better age was estimated through a better understanding of the landscape evolution.

A specific taphonomic scenario described for the Flemish Valley may help understand why the offshore and intertidal finds have lost their primary geological context. According to Gautier (1974, 1985) and Germonpré (1993b), herbivores migrated to the valleys during the cold winter for shelter and food, however mortality rates remained relatively high. Because of the cold climate the carcasses were better preserved during the winter. During the following spring, when the carcasses thawed the blowfly *Protomorphia terraenovae* led their eggs in the decaying flesh (similar to the juvenile mammoth find from the Westhinder sandbank). During spring the carcasses started to decay and were transported downstream during peak flash floods. This gradual, long-term accumulation of medium-sized to very large mammal skeletal remains was subject to the hydraulic properties of the transporting medium and the remains and the prevailing conditions. This means that in sheltered areas of the river, during decreased fluvio-hydraulic conditions the carcasses were deposited and rapidly covered by fine-grained sediments. It is therefore assumed that the transport of the bones is not very far away from the original site of deposition, perhaps at most a few kilometres (a secondary context).

The palaeo-Scheldt Valley is defined as the downstream section of the Flemish valley up to ca. 15 ka. This suggests that a similar taphonomic scenario likely took place in the downstream part of the valley system and may explain why so few primary geological context locations exist (with exception of the Scheur navigation channel). This observation is biased towards locations where industrial activities take place such as dredging of navigation channels.

7.8 Archaeological preservation assessment of the BCS

At present no archaeological sites or human artefacts from the Palaeolithic-Neolithic period have been found on the BCS. This is in stark contrast to the large number of archaeological sites/finds in Flanders (Figure 7.2). It is safe to assume that similar preservation conditions as described onshore also apply to the offshore region with the exception of the marine transgressions that affect the shelf area and most likely reworked the upper part of the archaeological record. In this section we discuss the possible human presence during previous glacial-interglacial cycles in relation to the palaeolandscape, preserved and buried landforms and their depositional environments in light of their overall palaeogeographic settings. We compare these environments to those of Flanders and apply the preservation rating of Ward and Larcombe (2008) for preserved sedimentary units and geomorphological grounds. Based on this we also assess the geological context, which is partially supported by the fossil finds described in section 7.7.

7.8.1.1 Early Middle Palaeolithic (Saalian: 300–130 ka)

7.8.1.1.1 Saalian

Not much of the BCS landscape is known before ca. 160 ka. From 160 ka onwards the BCS became incised by the palaeo-Scheldt Valley system (Figure 7.10). In Chapter 5 it is discussed that this valley system removed large parts of the underlying substratum on the Inner BCS and that well-defined terraces bound the river system. Not much later, approximately 10 to 20 ka, a GLOF resulted in the formation of a major incised trunk valley system that removed large parts (up to 10 m) of the Outer BCS substratum and resulted in the Offshore Scarp terrace that is between 4 and 14 m high. To its east this terrace consisted of a Paleogene hill, perhaps comparable to the hills in South Flanders, that likely contributed to this high elevation of the Offshore Scarp.

In Flanders early Middle Palaeolithic archaeological sites are best known from the valley edges of Flemish Valley tributaries, other sites are known from the Campine Area, an area comparable to the Middle and Inner BCS platforms, while some sites are also known from the Flemish Valley itself (Figure 7.2 B).

Within the Saalian landscape terraces and glaciofluvial environments may have been used as temporary occupation sites (Ward and Larcombe, 2008). Terraces are attributed a medium preservation rating when sediments in primary context are preserved, however on the BCS only rarely sediments on top of terraces are preserved. When these sediments are removed from the record no preservation rating is assigned, which we have assigned in this study as a geomorphological ground with an unknown preservation rating. Glaciofluvial sediments are known from the palaeo-Scheldt Valley as unit SCH1. This fluvial unit is composed of floodplain and channel sediments (see Chapter 5, Figure 5.7). The channel sediments may contain artefacts in secondary position (Table 7.1) that were incorporated through erosion of valley margins and former terraces. The floodplain deposits on the other hand, have a negligible reworking and thus have high preservation rating in primary context (Table 7.1).

Not much is known about the depositional environments of the Inner and Middle BCS outside incised-valley systems. The thin to almost absent sedimentary cover and the presence of a river system on these higher platforms make them comparable to the Flanders Campine Area. This suggests that the Inner and Middle BCS may have been used as temporary occupation sites, though no exact preservation rating can be given for these areas. The Campine Area contains several archaeological sites/finds indicating preservation on higher platforms with a thin protective cover is possible and suggests that the comparable Inner and Middle BCS may have preserved similar archaeological sites. However, the Eemian and Holocene marine transgressions may have affected these landscapes and their preserved archaeological sites to an unknown extent. The final result of the preservation rating of the Saalian landscape is presented in Figure 7.10 B.

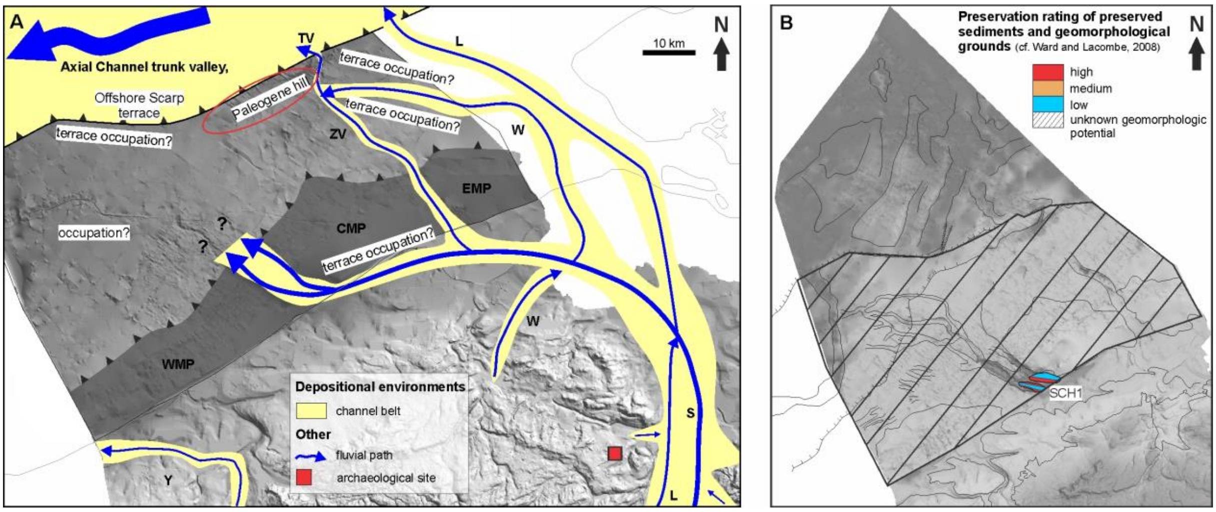


Figure 7.10 A) Palaeogeographic map for the Late Saalian (modified from De Clercq et al., 2018) with indication of actual preserved landforms. B) Indication of the final preservation potential for the landforms and the units.

7.8.1.2 Late Middle Palaeolithic (130–35 ka)

7.8.1.2.1 Early Eemian

By the Early Eemian the rising sea transformed the Outer BCS into a shallow marine sea where the Offshore Scarp was its palaeoshoreline up to the Middle Eemian transition (Figure 7.11 A). According to the landscape model explained in Chapter 5 this palaeoshoreline consisted of an open coast with sandy and muddy intertidal flats and local supratidal marshes. The sedimentary characteristics of the shallow marine subunit BCS1a (Figure 7.11 B) suggest that prolonged wave action occurred in front of the Offshore Scarp palaeoshoreline. This process likely decreased the preservation potential of coastal sediments in front of it (Table 7.1). These are therefore attributed a low preservation ranking (Figure 7.11 B). Supratidal marshes may preserve material in primary context and are given a medium preservation ranking (Table 7.1). However, up to now no supratidal marsh sediments of this Early Eemian palaeoshoreline have been found.

Along the Early Eemian palaeoshoreline two estuaries were present: the palaeo-Scheldt estuary and the Thornton Valley estuary (Figure 7.11 A). Estuaries belong to the richest coastal habitats in terms of readily available nutrition and early humans may well have exploited them since relatively early in their evolutionary history (Cohen et al., 2012, and sources therein). Each estuary consists of various depositional environments (e.g. intertidal sand and mud, supratidal marshes, etc.) each having their own characteristic preservation properties (Table 7.1). No sediments have been preserved from the Early Eemian palaeo-Scheldt estuary, but in the Thornton Valley estuary a stacked facies sequence of fluvial, estuarine and shallow marine deposits are preserved (subunit BCS1b; see Figure 6.6). The fluvial facies at the base of the sequence are likely channel deposits and have low preservation ranking and a secondary context. The overlying estuarine facies comprise a central basin mud. This environment is always submerged and protected by a coastal barrier. Preservation conditions are likely suitable but preservation of archaeological material is likely in secondary position. No evidence of a coastal barrier exists in the sedimentary record as it was likely removed during the Early-Middle Eemian transgression. A barrier environment would have had a high preservation rating in primary position (Table 7.1). The uppermost facies of the sequence in the Thornton Valley estuary consists of transgressive sand deposits. During the transgression the coastal barrier and the upper part of the central basin mud facies were removed. If archaeological material were preserved in the coastal barrier and central basin muds these may have been reworked in the lower parts of the transgressive sand layers. They are given a low preservation rating (Table 7.1). Intertidal sands, muds and supratidal peats formed along the edges of the Thornton Valley estuary. Each of these has a different preservation ranking (Table 7.1) ranging from absent to medium for the intertidal environments (primary and secondary contexts) to medium for the supratidal marshes (primary context). It is clear that the Thornton Valley estuary is a complex environment that comprises various depositional environments with a mix of preservation ratings and primary to secondary preservation context. To this end the overall environment as an estuary is considered as a suitable location for human occupation, especially during the climatic warm Early Eemian interval. A medium preservation ranking is given to the overall environment (Figure 7.11 B).

More inland the landscape is composed of meandering rivers comprising fluvial channels and floodplains with in between forested and/or open areas (Inner BCS and the Marginal Platforms WMP, CMP and EMP). Fluvial channels have a low preservation ranking in a secondary context; while the floodplains or organic-rich flood basins have a high preservation ranking in primary context (Table 7.1). At these locations human occupation may have been possible. Unfortunately, no evidence of these organic-rich flood basins has been preserved. The only location where Early Eemian sediments are preserved is in the palaeo-Scheldt Valley as fluvial unit SCH1 (Figure 7.11 B).

The primary geological context of mammalian bone material in the Scheur navigation channel, located on the northern edge of Marginal Platform EMP, suggests that at this location Early Eemian sediments are preserved. The high-ground position of the platform EMP is similar to that of the CMP and WMP and suggests that, when sediments are preserved, similar good preservation conditions may have prevailed at these locations. This remains largely unknown due to the presence of shallow gas and strong multiples in the seismic data. All other locations, outside the platforms are considered as geomorphological grounds with an uncertain preservation rating.

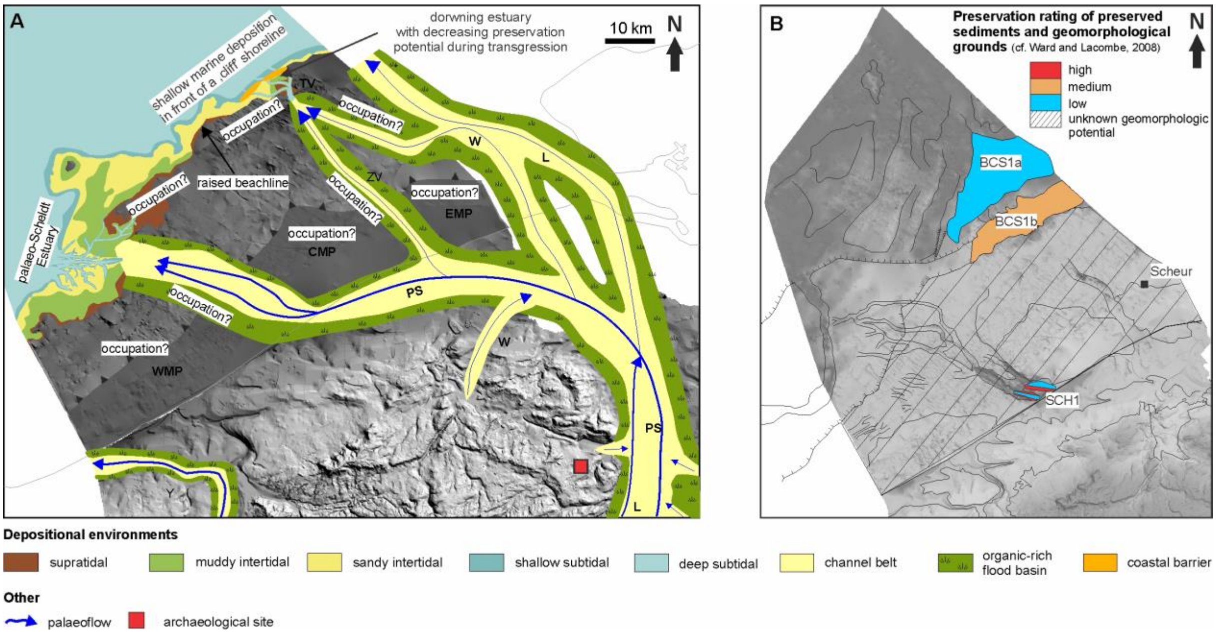


Figure 7.11 A) Palaeogeographic map for the Early Eemian (modified from De Clercq et al., 2018) with indication of actual preserved landforms. B) Indication of the final preservation potential for the landforms and the units.

7.8.1.2.2 Middle Eemian

Early Middle Eemian

During the first part of the Middle Eemian the invading sea transgresses the Offshore Scarp palaeoshoreline and Thornton Valley estuary, while the palaeo-Scheldt Valley transforms into a tide-dominated estuary (Figure 7.12 A). Upstream migrating tidal forces likely removed/reworked large parts of the former fluvial deposits (possible transition from primary to secondary contexts; e.g. red deer antler from Kwinte sandbank). This process also resulted in the isolation of the Marginal platforms CMP and EMP.

The palaeoshoreline east of the palaeo-Scheldt estuary likely consists of wide intertidal flats where the Zeebrugge, Waardamme and Lys river branches entered the sea, likely forming smaller estuaries. These areas may have been suitable for occupation, but no sediments are preserved at this location. The palaeo-Scheldt estuary is comparable to the present Westerscheldt estuary and comprises tidal channels, intertidal sands and muds and supratidal marshes. Tidal channels within the estuary are not expected to have any preservation potential due to their high fluvio-hydraulic energy and erosive power, while the intertidal sands have a low preservation ranking to preserve abandoned/reworked artefacts in a secondary context. Intertidal muds on the edges of the estuary may similarly contain abandoned artefacts. These may be preserved in a primary context due to lower reworking and have a medium preservation ranking. The supratidal salt marshes of the palaeo-Scheldt estuary may have been abundant on the downstream southwest edge where, according to the landscape model, a supratidal marsh landscape comparable to the present 'Land van Saeftinghe' may have been present. This part of the landscape, especially next to a nutrient-rich estuary may have been suitable for occupation during the Eemian climatic optimum. Sediments of this environment may have been incorporated into subunit SCH4a (see Chapter 5). This may be supported by the presence of shallow gas in this area, which may originate from peat layers related to this environment. If so, than these peat layers have a positive influence on the preservation rating of underlying sediments.

For the Scheur area on the Marginal Platform EMP the same conditions apply as during the Early Eemian, however, as the area is now more isolated by tidally influenced rivers it is likely that larger mammals, such as rhinoceros and straight-tusk elephant, disappear from this area. On the other hand the area may now have been more suitable for hippopotamus to thrive in, when the cooling climate still allowed this. The same high preservation ranking is maintained for this area (Figure 7.12 B).

Late Middle Eemian

During the second part of the Middle Eemian the former estuaries are now all submerged (compare Figure 7.12 A to C) and the climate is gradually cooling down (Vansteenberge et al., 2016). These locations have now become unsuitable for occupation. More so, during the marine transgression large sections of the former fluvial en fluvial-tidal sediments within the palaeo-Scheldt Valley have been reworked/removed. Sediments from this period are preserved as estuarine unit SCH3 and the shallow marine subunit SCH4a (Figure 7.12 D). These units are given a low preservation ranking (secondary and/or tertiary context).

The Scheur area is now located in an intertidal area on the northern edge of the Marginal Platform island EMP. The fossils preserved in primary context suggest that reworking in this area during transgression was small or negligible and that the remains of animals that lived in this environment may have been preserved here. If climate was already cold enough, though this is still speculation, walrus may have already resided in this area.

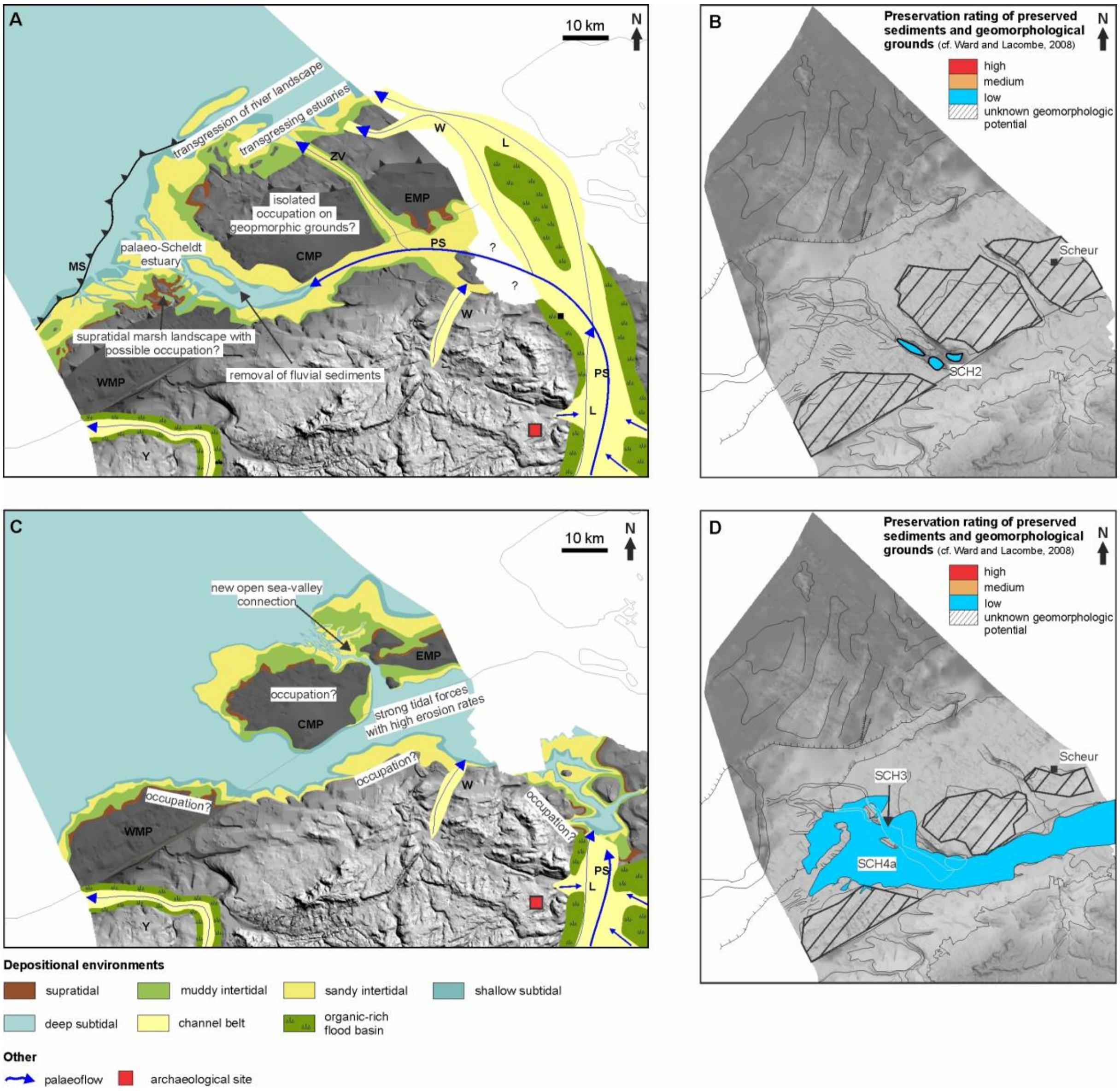


Figure 7.12 A and C) Palaeogeographic map for the Middle and Middle-Late Eemian (modified from De Clercq et al., 2018) with indication of actual preserved landforms. B and D) Indication of the final preservation potential for the landforms and the units.

7.8.1.2.3 Late Eemian - Early Glacial

By the Late Eemian the BCS was almost completely drowned (Figure 7.13 A). The coastline now consists of intertidal sand flats offshore the Yser Valley and intertidal/supratidal islands of the Marginal Platforms CMP and EMP. The Scheur area is still likely located on the last remnants of the island EMP. The deteriorating climate conditions and the presence of walrus bone indicate the island may have been used as a colony with a high preservation rating. Not shortly after the islands were drowned and the BCS was completely covered by the sea. It is expected that preservation potential for the BCS is negligible for this last part of the Late Eemian (Figure 7.13 B).

This situation likely lasted until the second part of the Early Glacial, the Rederstall-Odderade interval (ca. 93–71 ka). When sea levels lowered the rivers reclaimed their former Late Saalian-Middle Eemian positions and removed the upper parts of the former sandy infill. The position of the palaeoshoreline is not known for this period (Figure 7.13 C). The process of sea level lowering caused the rivers to cut down removing or altering the preservation rating of older deposits to an unknown extent. The fluvial meandering river systems were comparable to those of the Early-Middle Eemian, though the climate was likely colder. Most likely comparable preservation conditions to this period can be expected. Sediments from the Early Glacial are preserved as the fluvial unit SCH5 in the offshore part of the palaeo-Scheldt Valley. The fluvial meandering system is described as a low energy fluvial system with short pulses of increased fluvio-hydraulic energy. Unit SCH5 comprises channels deposits, floodplains and (downstream) a fluvial wetland. The channels have a low preservation ranking in secondary position, while the floodplains have a high preservation ranking in primary context (Figure 7.13 D). In a downstream direction the energy of the channels decreases suggesting a higher preservation ranking. The fluvial wetland consists of peat and clay layers that have been given a high preservation ranking in a primary and possibly secondary condition. From this fluvial unit it is suggested that many bone fossils originate (see Section 7.7.1.2) supporting the good preservation conditions for this environment

The Scheur area contains mammalian finds that indicate colder climate conditions. The presence of bear, wild horse, boar, red deer, reindeer and giant deer indicate sediments from this period may be preserved in this area. The primary context suggests a high preservation ranking in this area.

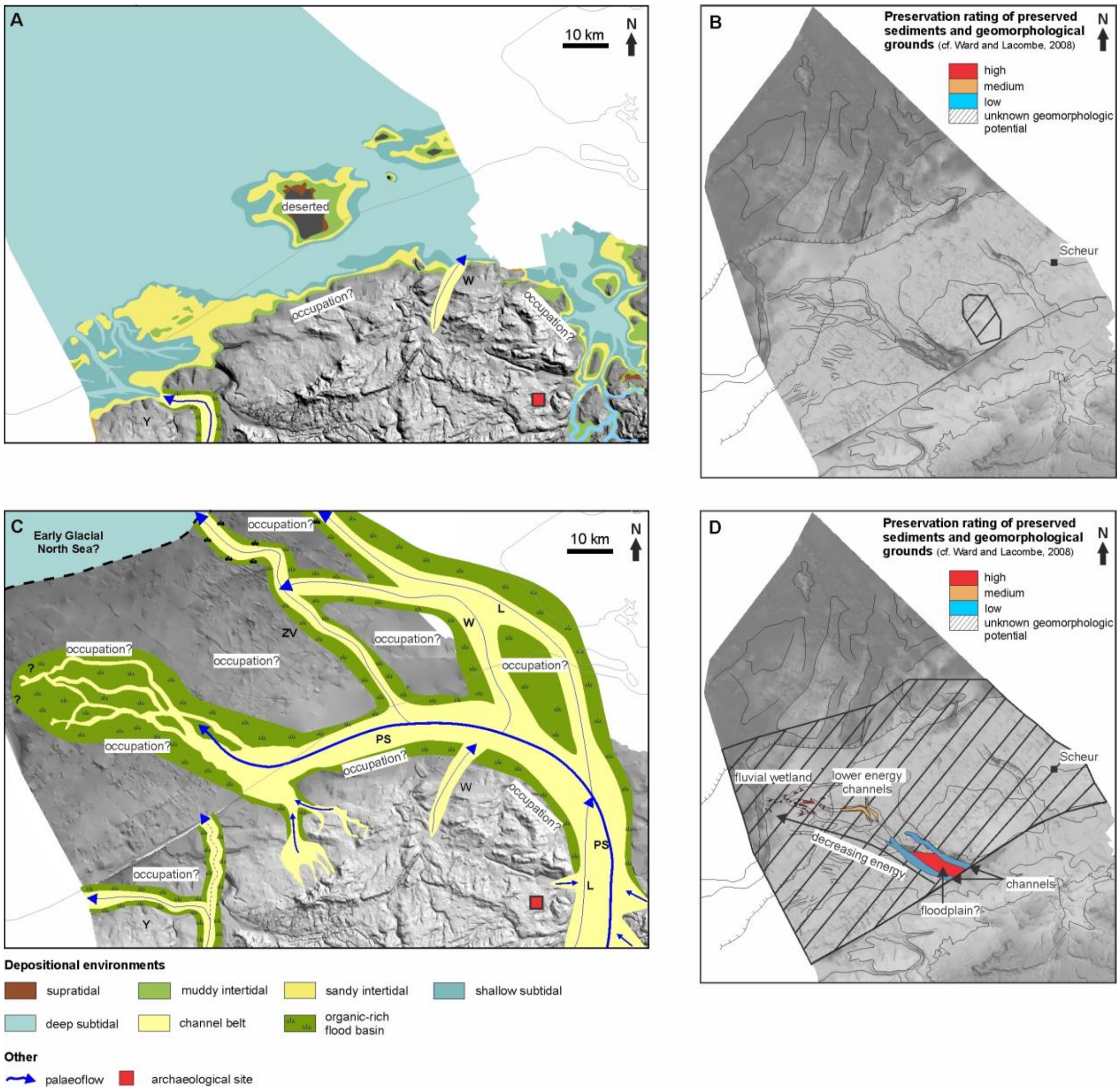


Figure 7.13 A and C) Palaeogeographic map for the Late Eemian and Early Glacial (modified from De Clercq et al., 2018, submitted) with indication of actual preserved landforms. B and D) Indication of the final preservation potential for the landforms and the units.

7.8.1.2.4 Pleniglacial

Climate deterioration during the Pleniglacial became more extreme as mixed boreal forests continued to degrade into a tundra landscape with permafrost development across northwest Europe (Huijzer and Vandenberghe, 1998). Lateral erosion may have been efficient due to the easily erodible Eemian and Early Glacial sandy deposits removing/reworking previous sediments with a high preservation ranking.

On the Inner-Middle BCS the palaeo-Scheldt Valley displaced its main activity from the Ostend Valley to the Zeebrugge Valley (Figure 7.14 A). The abandoned Ostend Valley likely hosted shallow pools that started to silt up. Similar environments are observed upstream in the palaeo-Scheldt Valley (Chapter 5) from channel abandonment (e.g. De Moor and Heyse, 1974; Heyse, 1979). Such low energy environments may have a high preservation potential for archaeological material (Ward and Larcombe, 2008) and may have been suitable for temporary occupation with a high preservation rating (Table 7.1). However here we also attribute a primary context to these environments instead of only a secondary as proposed by Ward and Larcombe (2008) as a result of the lack of water movement.

Palaeontological finds from the mammoth fauna in the Scheur and the Zeebrugge navigation channels indicate sediments from this period may be preserved in the area. However, for the Scheur it is unlikely that they are related to high fluvio-hydraulic episodes within the Zeebrugge-Waardamme Valley system, as they would rework/remove older deposits with primary preservation conditions. In other words, if the Scheur area was occupied then archaeological material is preserved here in primary context with a high preservation ranking. For the Zeebrugge navigation channel on the other hand the bone material may be the result of accumulation through fluvial transport. This may suggest that archaeological artefacts in the Zeebrugge-Waardamme Valley may be preserved in a secondary context.

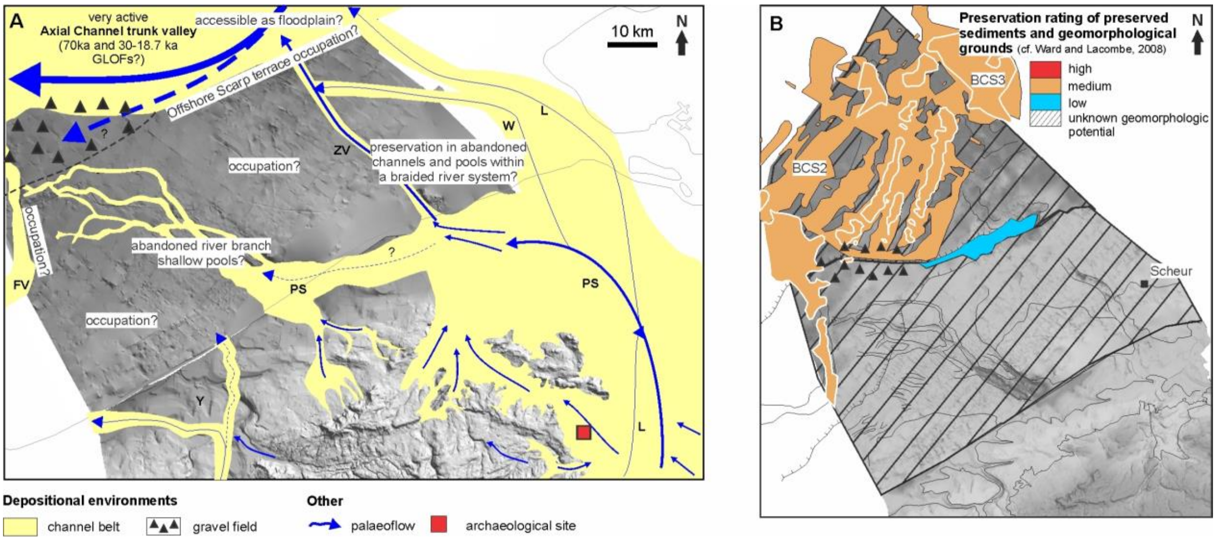


Figure 7.14 A) Palaeogeographic map for the Pleniglacial and Last Glacial Maximum (modified from De Clercq et al., 2018, submitted) with indication of actual preserved landforms. B) Indication of the final preservation potential for the landforms and the units.

During the Early Pleniglacial the Outer BCS belonged to the southern edge of the Axial Channel trunk valley (Figure 7.14 A). Approximately 70 ka ago this valley system may have channelled large volumes of water during GLOFs, with the Offshore Scarp terrace as the boundary of the river system. During these very-high fluvio-hydraulic events older sediments are removed/reworked and blocks as large as 2–3 m were deposited up to the northern part of the Middle BCS. When these GLOFs subsided a large floodplain filled with abandoned pools and streams remained present in the landscape with preservation conditions similar as those described in the palaeo-Scheldt and Zeebrugge-Waardamme Valley systems. However, the number of GLOFs that affected the area is uncertain, suggesting multiple episodes of erosion and deposition (represented as unit BCS2 and BCS3: Figure 7.14 B). The various depositional environments within this large floodplain may rank from low to high in primary and secondary context. For this reason we apply a medium preservation ranking to unit BCS2 and BCS3 ranging from a primary to secondary context (Figure 7.14 B). The secondary context is supported by the mammoth skull that was found near the Westhinder in the location of the gravel field on the Middle BCS. The Offshore Scarp terrace is attributed a low preservation ranking (Table 7.1) comparable to the Late Saalian period because no sediments are preserved on it.

During the Middle Pleniglacial the Outer BCS was likely an abandoned floodplain due the subsided fluvial activity in the Axial Channel trunk valley (see Chapters 5 and 6). This abandoned floodplain or geomorphic ground (as no sediments are preserved for the Middle Pleniglacial) may have been used for temporary occupation/migration. Archaeological material on the Outer BCS may have a medium preservation ranking in primary to secondary context.

7.8.1.3 Late Palaeolithic

7.8.1.3.1 Late Middle and Late Pleniglacial

Evidence for Late Palaeolithic human occupation is found along the edges of the Flemish Valley and its sandy infill but also on the higher plateaus of the Campine Area, as well as in the coastal Polders (Figure 7.2). This suggests that evidence for human occupation may still be preserved within similar environments of the BCS. It is assumed that climate conditions for the Early-Middle Pleniglacial continued into the LGM (e.g. Figure 7.15 A). From that moment on several major landscape changes occurred in northwest Europe that also impacted the BCS.

In Flanders the palaeo-Scheldt Valley system becomes gradually blocked by the formation of cover sands. These cover sand ridges were important Late Palaeolithic occupation sites in Flanders (Crombé et al., 2013a; Crombé and Robinson, 2017), however up to now no such cover sands have been recognised on the BCS. The palaeo-Scheldt blockade was finalised between 17 and 15.5 ka and from that point onwards the palaeo-Scheldt, Zeebrugge, Waardamme and Lys Rivers became inactive on the Inner BCS. This river abandonment may have left pools and small lakes in the landscape and it is assumed that local streams remained present in the landscape (Figure 7.14 A). It is not well understood how the landscape evolved from the LGM into the Holocene.

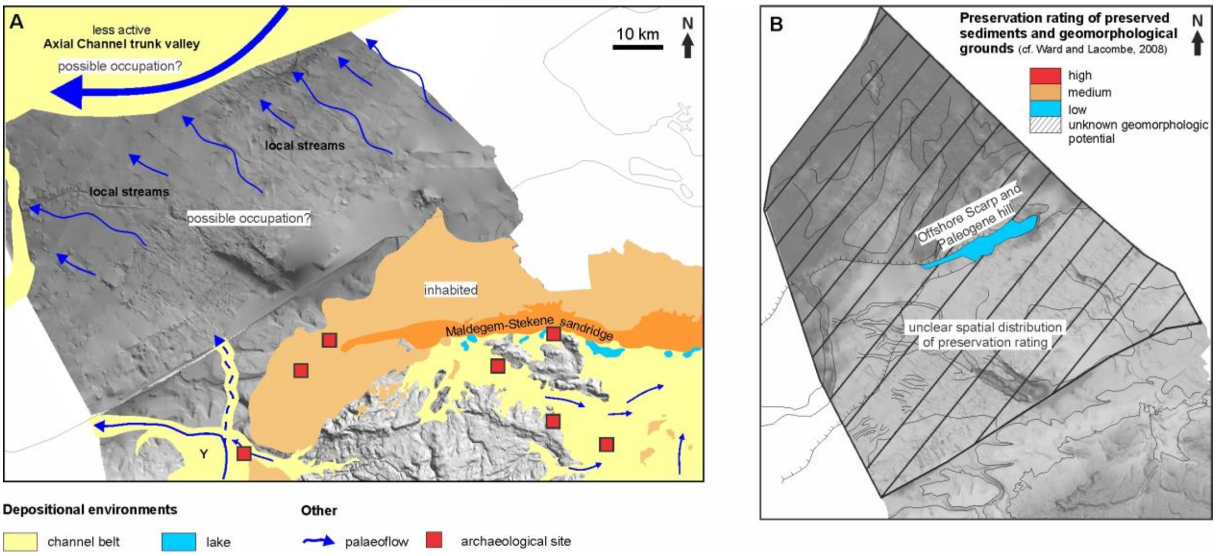


Figure 7.15 A) Palaeogeographic map for the final Late Pleniglacial post ca. 15 ka) and Late Glacial (modified from De Clercq et al., 2018, submitted) with indication of actual preserved landforms. B) Indication of the final preservation potential for the landforms and the units.

The Outer BCS one again became an inactive floodplain after 18.7 ka. The Offshore Scarp terrace and its Paleogene hill remained important features in the landscape that persisted into the Holocene. Perhaps these landforms that were important landmarks may have experienced occupation opportunities similar to cover sands in Flanders.

7.8.1.4 Final Palaeolithic (Late Glacial-Early Holocene; Figure 7.15 A and Figure 7.16 A–B)

Archaeological finds show that numerous occupation sites were present in Flanders during the Late Palaeolithic. Some of these sites are positioned on top of the former LGM cover sand ridge. Others are positioned in the lower valleys of minor streams. In the Yser Valley another occupation site was encountered. It is inferred that similar minor streams remained present in the landscape after the palaeo-Scheldt deflection post 15.5 ka. These occupation sites at or near smaller streams suggest that occupation of the Inner BCS was likely possible, though no sites/finds have been reported yet.

During the late Glacial-Early Holocene transition rising sea levels, up to 7 m ka^{-1}, rapidly transgressed the BCS. By 10 ka large parts of the BCS were already inundated. This Early

Holocene palaeoshoreline likely consists of extensive intertidal sand- and mudflats and a thin rim of supratidal marshes (Mathys, 2009). The erosive power of the transgressing palaeoshoreline is represented by a shell and gravel lag at the base of the Holocene sediments and indicates that older sediments were prone to reworking (Mathys, 2009; Chapter 5). This likely impacted the preservation potential of the upper part of Late Pleistocene sediments in a negative way resulting in a low preservation rating in primary position (Table 7.1).

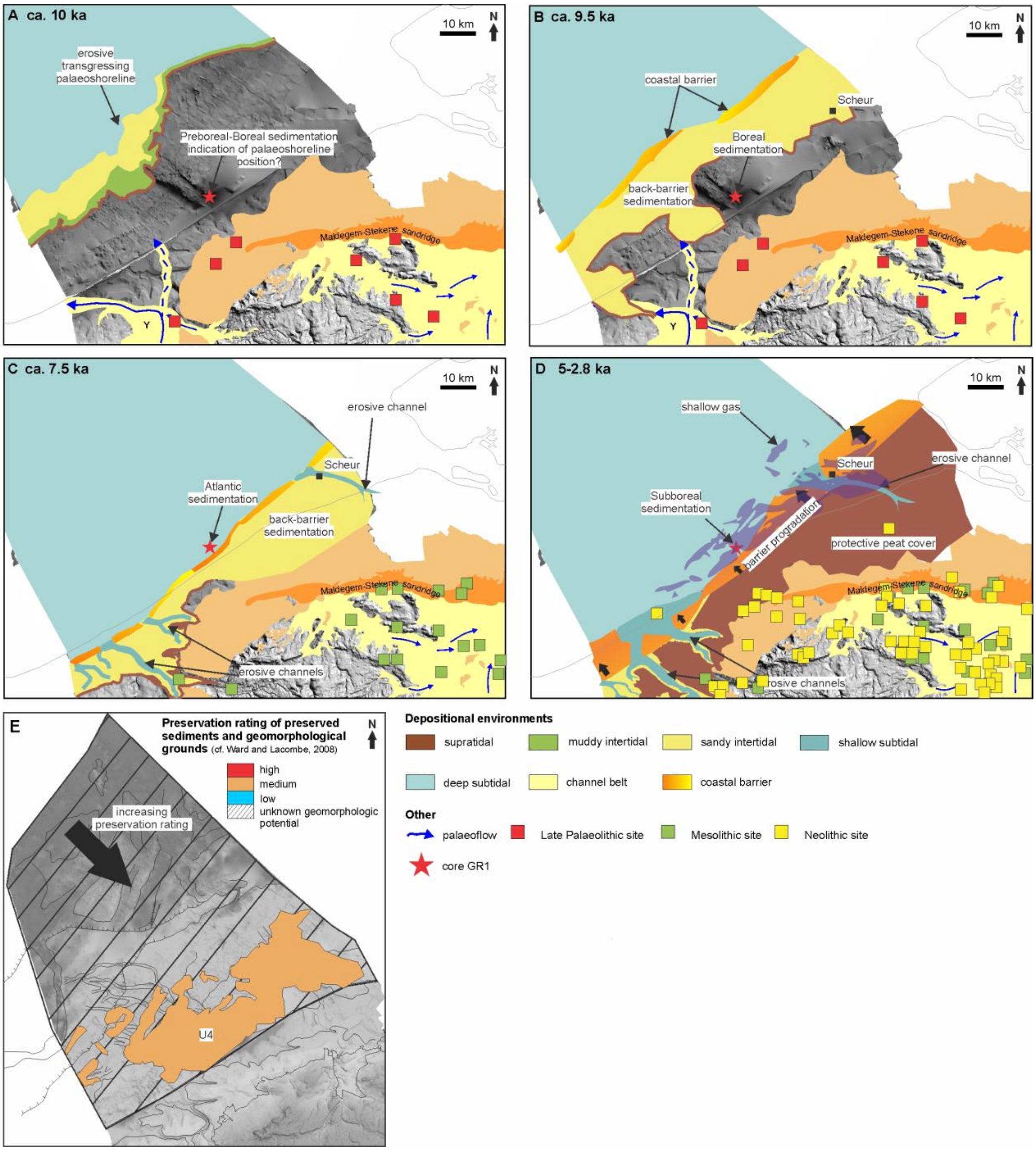


Figure 7.16 A-D) Palaeogeographic map for the Holocene (modified from Mathys, 2009) with indication of actual preserved landforms. E) Indication of the final preservation potential for the landforms and the units.

Pollen analysis of core GR1 (see Chapter 5 Figure 5.9), demonstrate that the Early Holocene palaeoshoreline was likely already located offshore Ostend as early as the Preboreal-Boreal (11.5–9.2 ka; Figure 7.16 A). This implies that large parts of the BCS may have already have been inaccessible to humans much earlier than previously assumed and that the palaeoshoreline reached much farther inland.

By 9.5 ka a coastal barrier complex had developed that protects the back-barrier environment, which comprises of tidal channels, intertidal flats and supratidal marshes (represented as unit U4 of Mathys, 2009). Tidal channels are high-energy systems prone to vertical and lateral erosion thereby removing and reworking older sediments. Such environments have a low preservation potential. The adjacent intertidal flats have a medium preservation potential and are likely to only contain abandoned material, whereas the salt marshes have a high preservation potential. The coastal barrier itself has a medium to high preservation potential (Ward and Larcombe, 2008) and may function as a protective cover for underlying archaeological material (Gonzales et al., 1996, 2000) but are themselves unlikely to preserve organic components, unless, in this case, when they were rapidly inundated and preserved under wet conditions. Because of this mixed and highly dynamic environment a medium preservation rating is applied to the coastal barrier and back-barrier palaeoenvironment, though it is unclear which parts of this environment are preserved in the sedimentary record.

7.8.1.5 Mesolithic and Neolithic (Middle-Late Holocene; Figure 7.15 C–D)

During the Mesolithic (8.2–4.2 ka) a slower sea-level rise (up to 4 m ka^{-1}) combined with a lower sediment supply over accommodation space ratio resulted in the landward migration of the protective coastal barrier. About 7.5 ka ago (Atlantic: Figure 7.16 C) it was already located near the present-day shoreline at Ostend (Mathys, 2009) where the sediments are buried ca. 3 m below the seafloor. To the east, the coastal barrier was located farther offshore. In front of Zeebrugge it was located 5–6 km outside the present coastline and at the Belgian-Dutch border ca. 10 km. The Scheur was positioned at the location of the coastal barrier complex and may have been protected this way from erosion. According to Mathys (2009) a large tidal channel, a precursor to the so-called Zwin Channel, was present at that location. The erosive action of such a large tidal channel seems to be in contradiction to the fossil finds from the Scheur and suggests that such a large tidal channel may not have been present at this location.

Between 6.8 and 5 ka coastal barrier progradation took place as a result of a decreased sea-level rise (Denys and Baeteman, 1995; Mathys, 2009; Figure 7.16 D). By 5.5 ka the barrier complex stabilised (Mathys, 2009) which likely had a positive impact on the preservation potential of lower Late Pleistocene and Early Holocene sediments. This may explain the large amounts of fossil bones that are preserved in the lower Early Glacial sediments like for example in front of Ostend in the palaeo-Scheldt Valley. Since 6.4 ka freshwater marshes developed behind the barrier complex and by 5.5–4.5 ka almost the entire area behind the coastal barrier was transformed into a marsh landscape (Figure 7.16 D; Mathys, 2009). Similar to a coastal barrier, peat formation is also a good way to preserve older sediments and its archaeological remnants (De Bie et al., 2008; Van Gils et al., 2010; Vanmontfort, 2011; Ward and Larcombe, 2008). This suggests that where peat is preserved in the sedimentary record a higher preservation potential exists for Mesolithic and Neolithic sites (see the purple shaded area in Figure 7.16 D).

Archaeological sites in Flanders during the period 7.5–2.8 ka are numerous and suggest that occupation may occur in a supratidal landscape (peat) and may have extended onto the coastal barrier system as this belonged to one accessible landscape (Figure 7.16 C and D).

The lateral extension of the sediments that belong to these Holocene landscapes, of which unit U4 of Mathys (2009) is the best approximation, shows that this unit may preserve landforms from a coastal barrier system, back barrier environment and supratidal marshes. These landscapes have a wide range of preservation ratings that cannot be exactly determined in this study. For this reason a medium preservation rating is given to this part of unit U4 of Mathys (2009).

7.9 Geological context for Palaeolithic-Neolithic archaeology: zooming out

For the Palaeolithic archaeological potential of the BCS area, one clear conclusion is that the late Middle Pleistocene sedimentary record (the Neanderthals) is very poorly preserved, whereas the Late Pleistocene record (Neanderthals and modern humans) is much better preserved. This is in stark contrast to the site distribution of Flanders (Figure 7.2). This is due to the tectonic ‘basin shoulder’ setting (Hijma et al., 2012), episodes of glacio-isostatic adjustments (Chapter 5), and enigmatic drainage network changes during the Saalian and Weichselian glaciations (Chapter 5 and 6). These long- and short-term processes resulted in progressive reworking and net erosion across the BCS. Only in deeply incised-valleys a stacked depositional record was able to accumulate as a result of the creation of accommodation space, with higher preservation potential for Pleistocene glacial-interglacial cycles (e.g the palaeo-Scheldt Valley). Throughout the BCS fluvial deposits (mapped or inferred) from the Late

Pleistocene are present at intermediate to shallow depths and are known to contain numerous mammalian remains at shallow depths. Many of these deposits are still within reach of commercial-industrial activities (extraction, dredging, fishing, wind farm, pipelines, cables, etc.) and future planned construction works such as the Complex Project Kustvisie in the near-coastal area.

The Late Saalian-Early Eemian fluvial unit SCH1 is the oldest deposit expected to contain traces of hominin presence. The Eemian interglacial units BCS1, SCH2, SCH3 and subunit SCH4a, from near-coastal, estuarine and shallow marine environments, have been preserved more extensively. Underneath the present coastline and further offshore these environments are preserved to a large extent at deep to shallow depths. From the Weichselian glacial period, both extensive and more local deposits, including river terraces, have been preserved along, amongst others, further reworking braidplains (unit BCS2 and BCS3). Close to the present shoreline and further offshore, surfaces of such terraces are preserved (e.g. the Offshore Scarp). However, most of the offshore area was superficially reworked during the Holocene marine transgression.

In the wider study area, the oldest traces of hominin occupation in northwest Europe (late Early to Middle Pleistocene) are reported from East Anglia and originate from an interglacial coastal plain setting with smaller rivers and estuaries north of the mouth of the Thames on the southern rim of the ancient North Sea coast (for an overview see Rose, 2009). The hominins who produced the artefacts recorded at the site of Happisburgh 3 (Parfitt et al., 2010) were living along a relatively small local river that may have become a tributary of the Rhine (Hijma et al., 2012). Given the palaeogeographical reconstructions presented in Chapter 5 and 6, and the amount of fossils retrieved in the near-coastal area it seems worthwhile to focus on the drowning Eemian palaeolandscape and the Early Glacial unit SCH5 in the palaeo-Scheldt Valley system. Furthermore it would be opportune to establish a good stratigraphy of the area around the Scheur. During the Eemian sea-level rise these landscapes shared a coastal plain similar as the older sites in East Anglia. The current distribution pattern can be read as pointing to a preference for near-coastal habitats of hominins, although sites in East Anglia lie relatively inland (Cohen et al., 2012; Hijma et al., 2012). A same observation can be made for Flanders. In contrast, later hominins such as Neanderthals have their traces preserved in a wider range of geographical settings, a.o. along more upstream river stretches and their interfluves (see also Figure 7.2) and the Ardennes in southeast Belgium (Di Modica et al., 2016). It is far too early to say whether the current distribution of very early Palaeolithic sites outside the BCS reflects a hominin preference or results from large-scale processes that control taphonomy. The major palaeogeographical changes described in Chapters 5 and 6 not only influenced hominin migration through the former landscapes but were also crucial for site taphonomy: the transition of a river-free to a fluvial-dominated glacio-isostatically upwarped land bridge (Late Saalian), that connects continental Europe with East Anglia, to a breached land bridge that subsequently is transgressed (Eemian) and regressed (Weichselian) to a final situation of a high-stand sea (Holocene) changed the landscape significantly and ought to have impacted migration routes of mammalian herds and the humans following them.

The Eemian interglacial sea-level rise resulted in favourable deposition conditions (syn-depositional preservation) through the accumulation of thick sedimentary sequences in Late Saalian incised-valley systems (e.g. the palaeo-Scheldt Valley system), but also areas in between these river systems (e.g. the Scheur). To this end the Eemian river and coastal landscapes could be of major importance. Of other relevance are the favourable habitats of the North Sea Basin in times of lowered sea level during the Weichselian, a period with favourable preservation condition (post-depositional preservation). For this period, parts of the study area are identified as being of large archaeological and palaeontological potential. In particular unit SCH5 must have been an ideal lowland habitat for Neanderthal hunter-gatherers in the earlier parts of the Weichselian when climate was still not too cold. The possibility of retrieving traces of hominin presence are limited by the fact that the areas where this unit lies close to the surface are, for the time being, not scheduled for aggregate extraction or other commercial-industrial activities.

7.10 Conclusions

Palaeogeographical reconstructions and landform-to-landscape identification provide the necessary background framework to the study of the natural environments of large mammals and early humans (including their spatio-temporal distributions), and for the assessment of taphonomic bias (most finds are the result of industrial activity). They are ideally based on a solid stratigraphic framework and knowledge of sedimentary environments, whereas offshore geophysical data (seismics and cores/vibrocores) are essential to constrain these reconstructions. Major palaeogeographical changes in the study area and northwest Europe occurred over the course of the Late Saalian-Holocene period (late Middle Palaeolithic-Neolithic) suggesting very distinct patterns of migration routes and occupation possibilities on a local to regional scale.

These local to regional landscape changes affected (removal or preservation) previous sediments (post-depositional processes) and provided opportunity to deposit new ones (syn-depositional processes) as is evident from the BCS stratigraphic record. The lack of submerged heritage compelled us to extrapolate taphonomic observations from onshore Flanders to the offshore area. This exercise was especially useful to understand the palaeontological bone material recovered from the palaeo-Scheldt Valley. However, caution is needed using such extrapolations because the study area has survived multiple phases of marine transgression (Eemian, Early Glacial, Holocene) and regression (Early Glacial, Pleniglacial) that affected taphonomy in a very different way compared to regions that have not been influenced by sea-level cycles onshore.

This study has also demonstrated that: 1) better age constraints can be provided to submerged heritage when the geological context of a find is better understood even when absolute dating of the material itself is lacking; 2) BCS submerged landscapes can preserve *in situ* palaeontological bones but has yet to prove this for archaeological artefacts; 3) palaeontological bone material is not always a ‘chance’ find, but indicate clear relationships to the submerged and the buried geomorphological features, palaeolandscapes that, although complex, can be examined in detail using a variety of existing field and analytical methods.

In the future, a close collaboration between scientists, policy makers and industry should make it possible to go beyond an assessment of potential submerged prehistory and identification of buried geomorphological features, and investigate the submerged heritage and its wider palaeogeographical context.

Table 7.2 (next pages) Overview of the faunal assemblages in the Flemish Valley of Flanders and the palaeo-Scheldt, Zeebrugge-Waardamme and Yser Valley systems of the BCS.

Location	Species		Fauna type	Fauna Association	Stratigraphy	Environment	Chronostratigraphy	Age	Context	Reference
Flemish Valley										
Rotselaar	*Mammuthus primigenius*	Mammoth	Mammoth	Hf I; HF II	NA	Cold open, steppe-like environment	Middle Weichselian	52–43.5	Secondary	Germonpré et al. (1993)
	Coelodonta antiquitatis	Woolly rhinoceros								
	Bison priscus	Steppe bison								
	Rangifer tarandus	Reindeer								
	Megaloceros giganteus	Giant deer								
	Canis lupus	Wolf								
	Panthera leo	Cave lion								
	Crocuta crocuta	Hyena								
	Cervus elaphus	Red deer								
	Equus sp.	Wild horse								
Hofstade	*Coelodonta antiquitatis*	Woolly rhinoceros	Mammoth	Hf I	Lembeke Mb	- Braided river with peak discharges - Gradual aridification from HF I to III	Early-Middle Pleniglacial	71–29	Secondary	Germonpré et al. (1993); Germonpré (2003)
	Alopex/Vulpe s	Arctic fox		Hf I						
	Ursus arctos	Brown bear		Hf I						
	Crocuta crocuta spelaea	Cave hyena		HF I; HF II						
	Panthera leo spelaea	Cave lion		HF III						
	Mammuthus primigenius	Mammoth		HF I; HF II; HF III						
	Equus sp.	Wild horse		HF I; HF II; HF III						
	Rangifer tarandus	Reindeer		HF I; HF III						
	Megaloceros giganteus	Giant deer		HF I; HF II; HF III						
	Bison priscus/ primigenius	Bison/aurochs		HF I; HF II; HF III						
Zemst	*Elephant antiquus*	Straight-tusked elephant	Elephant	Zemst A	Grimbergen Mb	Meandering river in forested landscape	Eemian	121 – 110	Secondary	Germonpré et al. (1993); Germonpré (2003)
	Equus sp.	Wild horse								
	Rhinocerotidae	Rhinoceros								
	Sus scrofa	Boar								
	Cervus elaphus	Red deer								

	Capreolus capreolus	Roe deer								
	Bison priscus/ primigenius	Bison/aurochs								
	Castor fiber	Beaver								
	Lepus timidus/Lepus capensis	Mountain or cape hare	Mammoth	Zemst IIB and IIC	Bos van AA Mb	Braided river system (Zemst IIB) that gradually shallowed in a cold and humid climate (Zemst IIC). The river system of Zemst IIB probably had water the whole year round, while the Zemst IIC was characterised by strong fluctuating hydraulic conditions	Early Weichselian	110 –71	Secondar y	
	Canis lupus	Wolf								
	Alopex/Vulpes	Arctic fox								
	Ursus arctos	Brown bear								
	Mustela putorius	Polecat								
	Meles meles	Badger								
	Crocuta crocutas spelaea	Cave hyena								
	Panthera leo spelaea	Cave lion								
	Mammuthus primigenius	Mammoth								
	Equus sp.	Wild horse								
	Coelodonta antiquitatis	Woolly rhinoceros								
	Sus scrofa	Boar								
	Cervus elaphus	Red deer								
	Rangifer tarandus	Reindeer								
	Capreolus capreolus	Roe deer								
	Megaloceros giganteus	Giant deer								
	Bison priscus/ primigenius	Bison/aurochs								
	Bos primigenius	Aurochs								
	Castor fiber	Beaver								
Dendermon de	*Mammuthus primigenius*	Mammoth	Mammoth	Dendermon de I	Dendermon de Mb	Gravelly braided river in an open, steppe environment	Early Weichselian	110 –71	Secondar y	Germonp ré et al. (1993)
	Coelodonta antiquitatis	Woolly rhinoceros	Mammoth	Dendermon de II	?	Open steppe environment	Middle Weichselian	ca. 30		
	Equus sp.	Wild horse								

	Bison priscus/ primigenius	Bison/aurochs								
	Megaloceros giganteus	Giant deer								
	Cervus elaphus	Red deer								
	Rangifer tarandus	Reindeer								
Overmere	*Elephant namadicus*	Asian straight-tusked elephant	Elephant	Overmere I	Oostwinkel Mb	Fluvial meandering system	Eemian	121 – 110	Secondary	Germonpré and Ervynck (1988); Germonpré (1993)
	Rhinocerotidae	Rhinoceros								
	Equus sp.	Wild horse								
	Bos primigenius	Aurochs								
	Cervus elaphus	Red deer								
	Capreolus capreolus	Roe deer								
	Sus scrofa	Boar								
	Carnivora	Carnivores								
	Castor fiber	Beaver								
	Mammuthus primigenius	Mammoth	Mammoth	Overmere II	Dendermonde Mb	Gravelly braided river in an open, steppe environment with locally wood areas or parkland	Early Weichselian (no extreme cold conditions)	110 –71	Secondary	
	Coelodonta antiquitatis	Woolly rhinoceros								
	Equus sp.	Wild horse								
	Bos primigenius/Bison priscus	Bison/aurochs								
	Megaloceros giganteus	Giant deer								
	Cervus elaphus	Red deer								
	Cervus elaphus/Rangifer tarandus	Red deer/reindeer								
	Capreolus capreolus	Roe deer								
	Carnivora	Carnivores								
	Sus scrofa	Boar								
	Mammuthus primigenius	Mammoth	Mammoth	Overmere III	Dendermonde Mb	Gravelly braided river in an open, steppe environment	Last Glacial (extreme climate)	29–14	Secondary	
	Coelodonta antiquitatis	Woolly rhinoceros								
	Equus sp.	Wild horse								

	Bos primigenius/ priscus	Bison/aurochs								
	Megaloceros giganteus	Giant deer								
	Cervus elaphus/Rangifer tarandus	Red deer/reindeer								
	Carnivora	Crnivores								
Oudenaarde	*Ursus arctos*	Brown bear	Mammoth	Zemst IIB; Hofstade I	NA	Unspecified fluvial environment	Early Weichselian	110 –48	Secondar y	Guatier (1995)
	Crocuta crocuta spelaea	Cave hyena								
	Felis silvestris	Wild cat								
	Mammuthus primigenius	Mammoth								
	Coelodonta antiquitatis	Woolly rhinoceros								
	Equus cf. germanius	Wild horse								
	Sus scrofa	Boar								
	Cervus elaphus	Red deer								
	Rangifer tarandus	Reindeer								
	Megaloceros giganteus	Giant deer								
	Bison priscus/ primigenius	Bison/aurochs								
palaeo-Scheldt Valley										
Blankenberg e	*Bos primigenius* (juvenile)	Aurochs	Holocene/ elephant / mammoth	NA	Unit Ho; Unit SCH6?	Fluvial?	Eemian; Weichselian; Holocene		Secondar y	This study
Ostend	*Odobenus rosmarus*	Walrus	Marine	NA	Unit SCH4	Cold marine	Eemian		Secondar y/ tertiary	This study
	Actinopterygii	Ray-finned fish	Marine (fresh?)	NA	Unit SCH4; Unit SCH5; Unit Ho	Fluvial/marine	Eemian; Early Glacial; Holocene		Secondar y/ tertiary	
	Mammuthus *primigenius*	Mammoth	Mammoth	Zemst IIB	Unit SCH5	Fluvial	Early Glacial		Secondar y/ tertiary	

	Megaloceros giganteus	Giant deer	Mammoth	Zemst IIB	Unit SCH5	Fluvial	Early Glacial	Secondary/ tertiary	
	Rangifer tarandus	Reindeer	Mammoth	Zemst IIB	Unit SCH5	Fluvial	Early Glacial	Secondary/ tertiary	
	Cervus elaphus	Red deer	Mammoth / Holocene	Zemst IIB/ Ho	Unit SCH5/ Unit Ho	Fluvial/terrestrial	Early Glacial/Holocene	Secondary/ tertiary	
	Capreolus capreolus	Roe deer	Mammoth / Holocene	Zemst IIB/ Ho	Unit SCH5/ Unit Ho	Fluvial/terrestrial	Early Glacial/Holocene	Secondary/ tertiary	
	Sus scrofa	Boar	Mammoth / Holocene	Zemst IIB/ Ho	Unit SCH5/ Unit Ho	Fluvial/terrestrial	Early Glacial/Holocene	Secondary/ tertiary	
Kwintebank	*Cervus elaphus*	Red deer (antler)	Elephant/ Holocene		Unit SCH4a/Unit Ho	Reworked marine	Early Eemian-Early Holocene	Secondary/ tertiary	This study
Westhinder	*Mammuthus primigenius*	Mammoth	Mammoth	NA	Unit SCH5; Unit BCS2; Unit BCS3	Fluvial	Early Glacial or Late Pleniglacial	Secondary/ tertiary	This study
Zeebrugge-Waardamme-Lys Valley									
Zeebrugge	*Mammuthus primigenius*	Mammoth	Mammoth		Unknown	Fluvial	Early Glacial; Pleniglacial; Late Glacial	Primary/ secondary	This study
	Mammuthus primigenius/Elephant antiquus	Mammoth/straight-tusked elephant	Mammoth/elephant		Unknown	Fluvial	Eemian; Early Glacial; Pleniglacial; Late Glacial	Primary/ secondary	
	undetermined vertebra	NA	Unknown		Unknown	Unknown	Unknown	Primary/ secondary	
	undetermined bone fragment	NA	Unknown		Unknown	Unknown	Unknown	Primary/ secondary	

	undetermined vertebra	NA	Unknown		Unknown	Unknown	Unknown		Primary/ secondary	
	undetermined vertebra	NA	Unknown		Unknown	Unknown	Unknown		Primary/ secondary	
	undetermined vertebra	NA	Unknown		Unknown	Unknown	Unknown		Primary/ secondary	
	undetermined piece of tusk or rib	NA	Unknown		Unknown	Unknown	Unknown		Primary/ secondary	
	undetermined bone	NA	Unknown		Unknown	Unknown	Unknown		Primary/ secondary	
Scheur	*Archaeoceti*	Ancient whale	Marine	NA	Maldegem Formation	Marine	Middle Lutetian	43 Ma	Primary	Post et al. (2017);
	Odobenus rosmarus	Walrus	Marine	NA	Unit SCH4?	Cold marine	Eemian		Primary	this study
	Equus sp.	Wild horse	Mammoth/elephant		Unknown	Fluvial	Eemian		Primary	
	Bos primigenius	Aurochs	Elephant	Zemst A/ Overmere I	Unknown	Fluvial	Eemian		Primary	
	Cervus elaphus	Red deer	Mammoth/elephant	Zemst A/ Overmere I	Unknown	Fluvial	Eemian		Primary	
	Ursus sp.	Bear	Mammoth	Zemst IIB; Hofstade I		Fluvial?	Early Glacial/Pleniglacial			
	Capreolus capreolus	Roe deer	Elephant	Zemst A/ Overmere I	Unknown	Fluvial	Eemian		Primary	
	Stephanorinus kirchbergensis	Rhinoceros	Elephant	Zemst A/ Overmere I	Unknown	Fluvial	Eemian		Primary	
	Sus scrofa	Boar	Elephant	Zemst A/ Overmere I	Unknown	Fluvial	Eemian		Primary	
	Hippopotamus incognitus	Hippo	Elephant	Zemst A/ Overmere I	Unknown	Fluvial	Eemian		Unknown	
Zeebrugge shipwreck	*Coelodon ta antiquitatis*	Woolly rhinoceros	Mammoth				Early Glacial-Late Glacial		Secondary/ tertiary	This study
Thornton	*Anancus Avernensis*	Elephant	NA	faunal association I	IJmuiden Ground and	Low energy open-marine	Early Pleistocene (Tiglian)		Secondary/	van Kolfshote

	Mammuthus meridionalis *Eucladoceros sp.* *Equus sp.*	Southern elephant Bush-antlered deer Wild horse		Winterton Shoal Fm	delta deposits from the Rhine and Meuse		tertiary	n and van Essen (2004)
Yser Valley								
Nieuwpoort	*Mammuthus primigenius/Elephant antiquus*	Mammoth/straight-tusked elephant	Mammoth/elephant	Cycle IV and V (Bogemans et al., 2016)	Tidal; fluvial	Eemian-Late Glacial	secondary	This study

8 PROFILE TYPE MAPPING: ARCHAEOLOGICAL AND PALAEONTOLOGICAL POTENTIAL MAPS

Abstract

A geological map of the Quaternary deposits of the Belgian Continental Shelf is presented. The map differs from a conventional geological map, which is of little use for the specific geological architecture of the continental shelf because of the high lateral and vertical variability in lithology in the subsoil. The method of profile type maps was chosen because these maps allow the representation of the entire vertical sequence of the Quaternary deposits. A profile type map is constructed on the basis of profile types consisting of the vertical succession of the various mapping units. This user-friendly map is based on the units defined in Chapters 5 and 6. The application of the system is demonstrated with three different profile type maps: a general profile type map, and two applied profile type maps showing the assessment where archaeological and palaeontological material may be encountered.

8.1 Introduction

The architecture of the Quaternary sediments of Flanders is highly complex. They have a great diversity in thickness and are both laterally and vertically heterogeneous. Most of the Quaternary deposits have a terrestrial origin (e.g. fluvial and dune deposits), and often do not reach thicknesses over 30 m. In many cases they are even less than 1 meter thick. The complexity of the geological architecture makes it difficult to visualise the different depositional environments in high detail on geological maps. In 1993 the Flemish Government ordered a series of maps to be developed that are able to map the complex Quaternary geological architecture of Flanders. These maps needed to be application-oriented and had to be used as policy supporting, educational, scientific and spatial planning documents. To this end profile type maps were created (see www.dov.vlaanderen.be).

Geological investigations of the Quaternary deposits of the BCS (Mathys, 2009; Chapters 5 and 6) demonstrated that the nature and architecture of these deposits are very similar as those in onshore Flanders. Sediments outside incised-river valleys often do not exceed 10 m in thickness (Figure 8.1). Only where sandbanks have formed the sediments may accumulate up to 30 m. Incised-valley systems are one of the few locations where enough accommodation space is developed for sediments to accumulate. There sediment sequences may stack up to 30 m with exceptions up to ca. 50 m. Compared to Flanders the sediments of the BCS are more difficult to map and access. Moreover, they are non-replenishable resources that are very important to the local economy and policy makers. A sustainable management of these natural resources is therefore very important. The last 10 years the socio-economic demand for marine resources has increased at an unprecedented pace for coastal protection works, harbour extensions and industrial land purposes (www.blauwecluster.be). Future demand for sand will continue to increase for two reasons: the present sea-level rise and protection from high-impact storms require more sand for sand suppletion to maintain its protection purposes; and larger volumes of resources are required for large-scale planned infrastructural works that form the key to future sustainable coastal development (e.g. Complex Project Kustvisie).

The layers that form the target of aggregate extraction are often quite thin thereby enhancing the chance of possibly exposing older, in this case Late Pleistocene layers, that have a specific preservation potential for submerged heritage (e.g. archaeological and palaeontological material; see Chapter 7). Similar exposures are created by other economic activities such as maintenance dredging of navigation channels, and the construction of cable connections such as the Nemo Link Interconnector (www.nemo-link.com/nl/het-project/de-keuze-van-hetkabeltraject). It is therefore necessary to understand where extraction of natural resources or other economic activities make contact with these archaeological and palaeontological valuable layers. A practical tool to visualise the complex nature of the sediments is a profile type map, similar to that used onshore Flanders. The map provides a connection to the geological architecture and helps understand the formation histories of the separate depositional environments. More importantly, it helps to highlight areas where potentially valuable layers are located. In this chapter a profile type map is created of the BCS that delineates specific zones of depositional environments as stacked sequences. From this map and the results of Chapter 7 an archaeological and palaeontological potential map is created. The archaeological potential map delineates zones where archaeology of specific periods can be expected while the palaeontological profile type map indicates where specific faunas can be expected. Each of these maps visualises these results irrespective of depth.

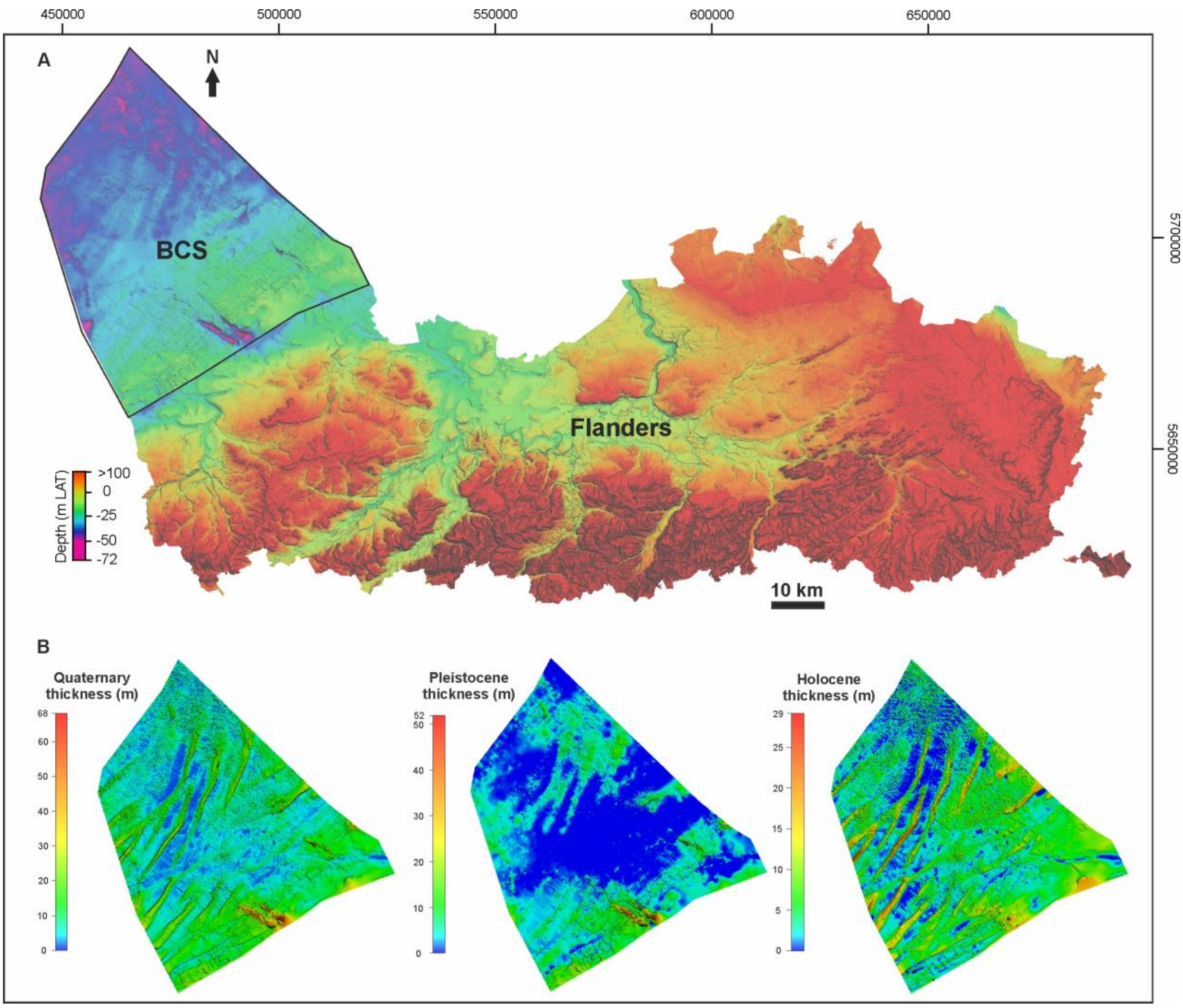


Figure 8.1 A) The pre-Quaternary surface of Flanders and the Belgian Continental Shelf. B) Isopach maps of the Quaternary, Holocene and Pleistocene sediments of the Belgian Continental Shelf. The Holocene isopach map is also equivalent to unit Ho.

8.2 Methodology

8.2.1 Profile Type Map

The wide variety of depositional environments identified on the BCS requires a practical tool to visualise them together in a two-dimensional plane. Profile type maps allow to image both the geographical distribution as well as the vertical stacking, or chronology, of the different depositional environments (Figure 8.2). By visualising the vertical architecture a third dimension is added to a two-dimensional map. To visualise this third dimension, on a two-dimensional plane, profile types are created. A single profile type represents the different geological layers in a mapped area. Every profile type will form a polygon with a specific geological layering, in the case for the BCS from the late Saalian to the Holocene. The nature and reliability of the mapped profile types is validated through cores and seismic reflection data within the first step prior to sequencing (cf. Chapters 5 and 6). This means no need for adaptations was necessary. For most of the study area this approach is very reliable due to the fragmented deposits and the clear boundaries of the units and the variable geological histories of the Inner, Middle and Outer BCS. It should be noted however that the geological architecture of the nearshore western and eastern Inner BCS is not well understood and that interpretations made about these areas depend mainly on inferred palaeogeographic mapping. Further research is needed to obtain reliable profile types of these areas.

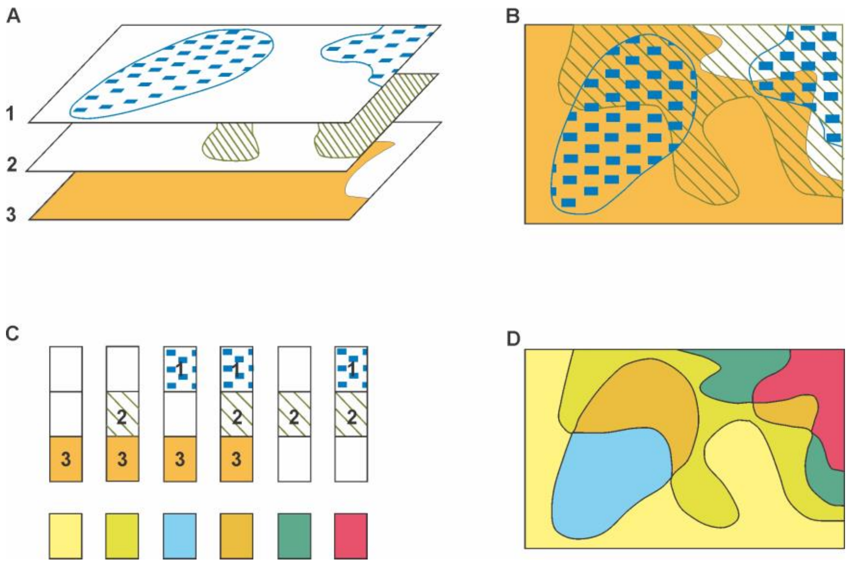


Figure 8.2 Principle of profile type mapping: A) all geological layers within a certain area are characterised, mapped and vertically stacked; B) the stacking results in certain sequences of layers that varies from region to region; C) from B these sequences of profile types can be identified and are given a code; D) the final result is a profile type map that represents vertically stacked geological layers in a two-dimensional plain as cartographic layers. Modified from www.dov.vlaanderen.be.

In the study area 18 profile types are distinguished and no generalisation of profile types was made. No distinction is made for the preservation ranking (absent, low, medium, high) or the geological context (primary, secondary or tertiary). This was tested, however each time the final map lost the information of each individual unit, as it is unfeasible to extrapolate these values or designated qualities to a two-dimensional plane. Units that were identified within this study but that are located outside the BCS, e.g. subunits SCH4b and SCH4c and units SCH6 and SCH7, are not included within the profile type maps and will not be further discussed.

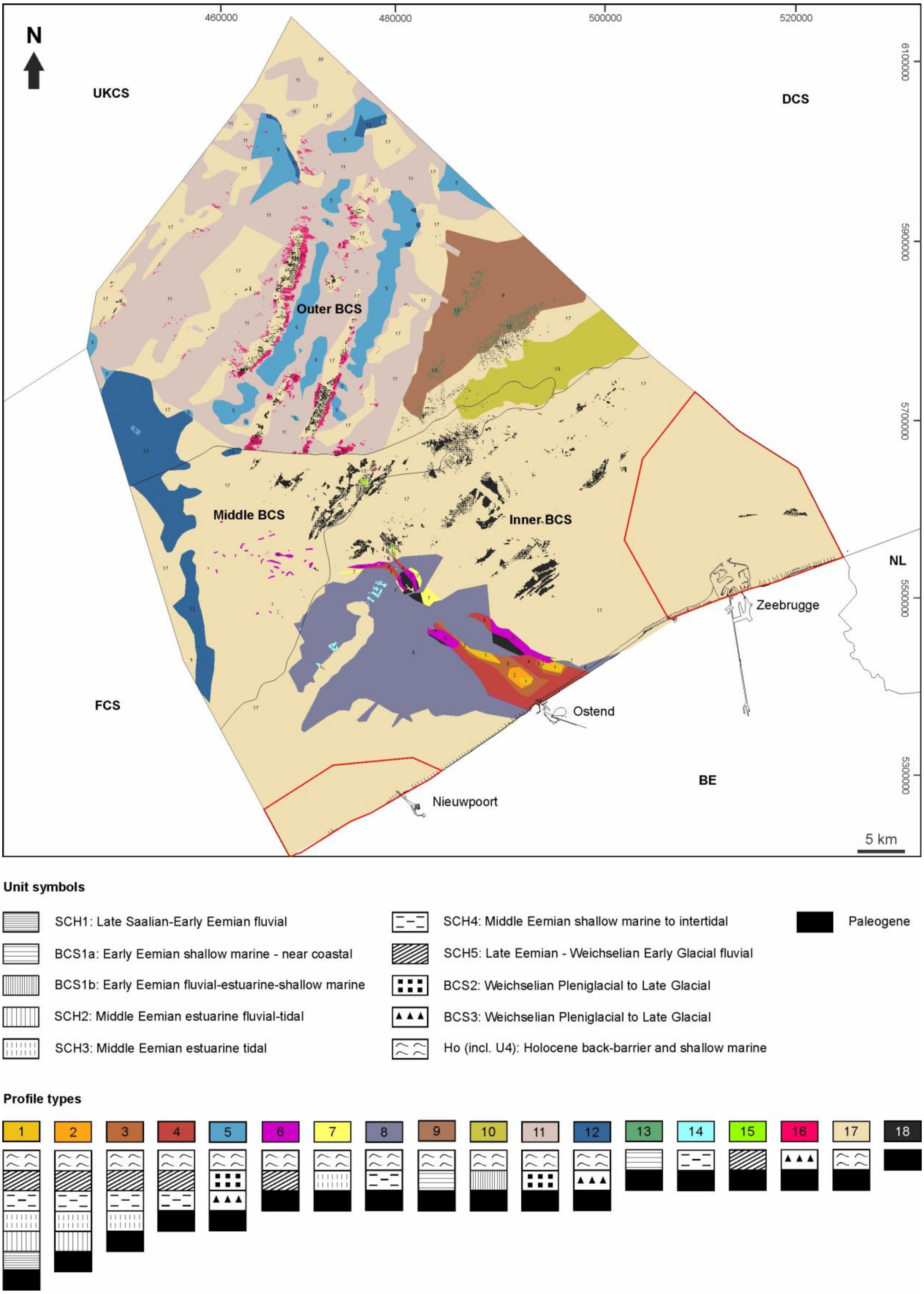


Figure 8.3 Profile type map of the BCS with indication of areas that have an unknown Pleistocene stratigraphy (red polygons).

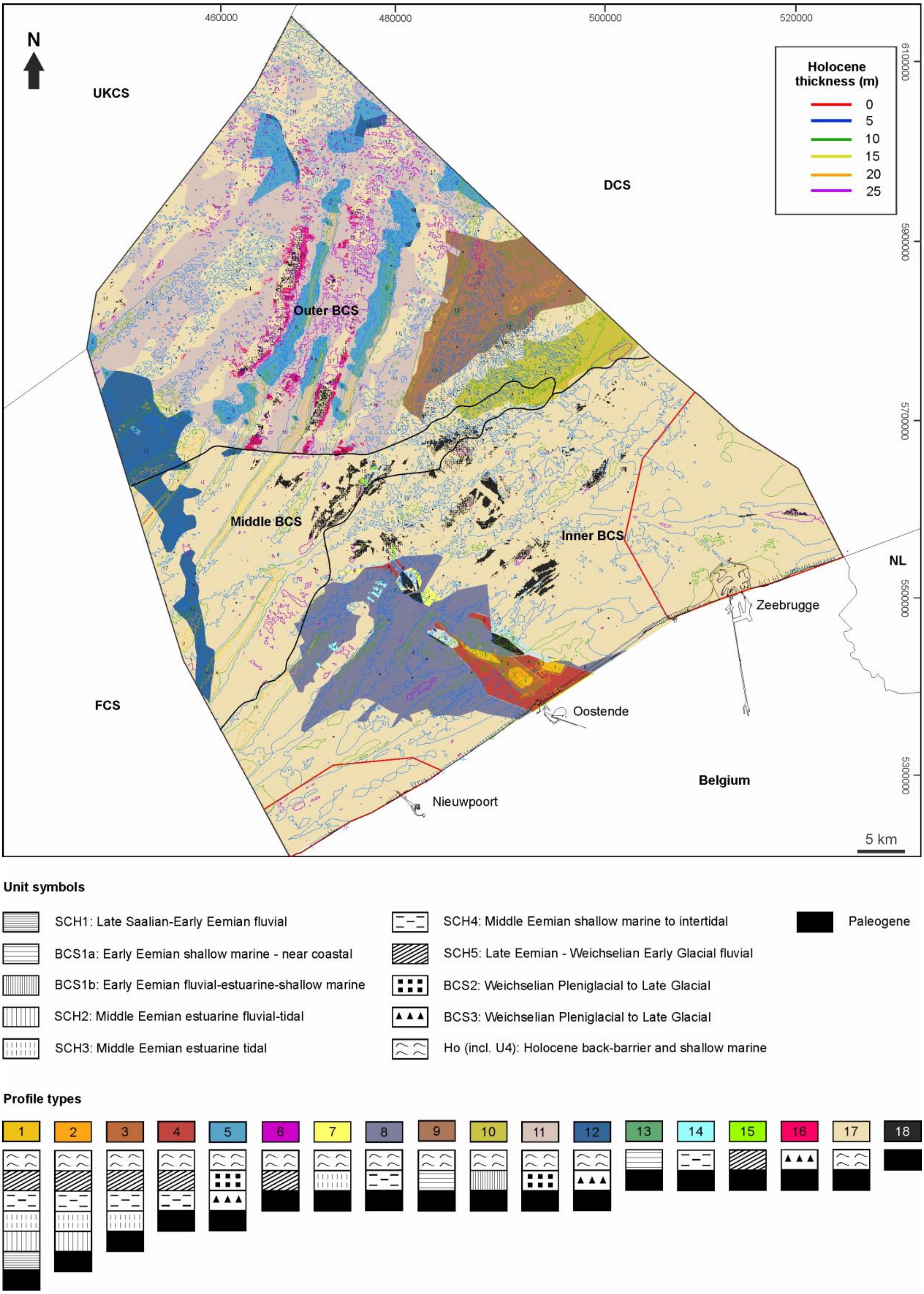


Figure 8.4 Same profile type map of the BCS as in Figure 8.3 with indication of the thickness of the Holocene sediments. The thicknesses of the Holocene sediments shows, which profile types, are vulnerable to natural or anthropogenic disturbance.

8.3 Results and Discussion

8.3.1 Profile Type Map

The final profile type map (Figure 8.3) consists of 18 different profile types that reveal several clear patterns in the Quaternary geology of the BCS. The red polygons in the western and eastern near-coastal areas represent zones where the Quaternary stratigraphy is not well constrained. This may seem contradictory to Figure 8.1 where the Pleistocene thicknesses of these areas have been calculated. However, this is the direct result of volume calculations by abstracting the Holocene thickness from the Quaternary thickness.

Profile type 17 is the by far the most widespread and covers most of the Inner and Middle BCS. It consists of Holocene sediments directly overlying the Paleogene substratum. This profile type demonstrates that extensive parts of the Quaternary geology have been removed from the geological record (or deposition did not take place on the Inner-Middle BCS). In some other locations Holocene sand is even absent (profile type 18) and in these locations various Paleogene strata surface at the seafloor. This profile type 18 is also mostly located on the Inner and Middle BCS. If we include the areas where unit Ho is less than 5 meter thick (Figure 8.4) it becomes obvious that in many areas the Pleistocene sediments in fact are buried quite shallow. This is particularly the case for the Outer BCS for units BCS2 and BCS3 (profile types 11 and 16) and BCS1a (profile types 9 and 13). Parts of these layers are actively being eroded or affected by anthropogenic activities.

On the Inner BCS, offshore Ostend, a complex pattern of profile types is revealed that relate to the infilling of the Ostend Valley as part of the palaeo Scheldt Valley from late Middle to Late Pleistocene times. Most of the Ostend Valley infill is covered by unit Ho and is well protected (5–10 m in thickness), with exception of profile type 14 where Middle to Late Eemian marine sediments are located at the seafloor. On the Outer BCS several profile types with a large geographical distribution are present. In the southeast area of the Outer BCS these are represented by profile types 9, 10 and 13 and consist of early Eemian sediments. Unit Ho again largely covers these sediments. Pleniglacial sediments dominate the remaining part of the Outer BCS. The distribution of these profile types shows a strong correlation with the present-day sandbanks in that area. This is particularly the case in the central Outer BCS. Here profile type 5 demonstrates where units BCS2 and BCS3 are well protected. This is also the only zone where these two units occur simultaneously at one location.

8.3.2 Archaeological Potential Map

For matter of simplicity, the archaeological potential map in Figure 8.5 is exclusively based on the age of the stratigraphic units and the expected potential for preservation without considering any primary, secondary or tertiary context (see Chapter 7 for the geological context). Six distinct archaeological zones based on lithological units and one based on geomorphological landforms have been created. Archaeological potential zone (APZ) 1 contains the oldest possible archaeological material. However, a large part of is located at great depths, between 30 and 50 m deep, and therefore well out of range of most industrial activities. It is unlikely that discoveries will be made from this zone. Chances are more likely that this happens onshore, where these layers are located closer to the surface (Heyse, 1979; Heyse and Demoulin, 2018; De Moor and Heyse, 1974; Chapter 5). All other APZs are located at medium to shallow depths (5–20 m) or are sometimes even exposed at the seafloor. These zones are more vulnerable to damage or destruction from commercial and industrial activities. Large sections of APZs 2 and 5 are buried shallow or exposed at the seafloor (see Figure 8.4 and Figure 8.5). Large-scale infrastructural works to protect our coastline from future sea-level rise and millennial storms are planned within the coming years (Complex Project Kustvisie). Such infrastructural works require large amounts of sediments, mostly from offshore areas, where archaeological material may be close to the surface (APZs 2, 3 and 5). On the other hand, such works affect the subsurface up to great depths. Stability for a lot of construction works is best achieved on the BCS by reaching down to the Paleogene substratum. This means that that the entire Quaternary sedimentary sequence may be affected. This means that these construction works will almost always affect APZs 4 and 5.

Although commercial and industrial activities disturb the Quaternary sediments they are also one of the few, if not only, available source that provide information about buried submerged heritage. To this end the planned infrastructural works of the Complex Project Kustvisie at Zeebrugge provide an opportunity for new finds on the BCS in the near future. The success rate of such discoveries can be increased through extensive scientific research during the planning stage. A good example of this is the Rotterdam harbour extension. There a geogenetic approach was applied to detect drowned Mesolithic sites in the transgressive palaeoenvironment of the Holocene Rhine-Meuse delta (Vos et al., 2014).

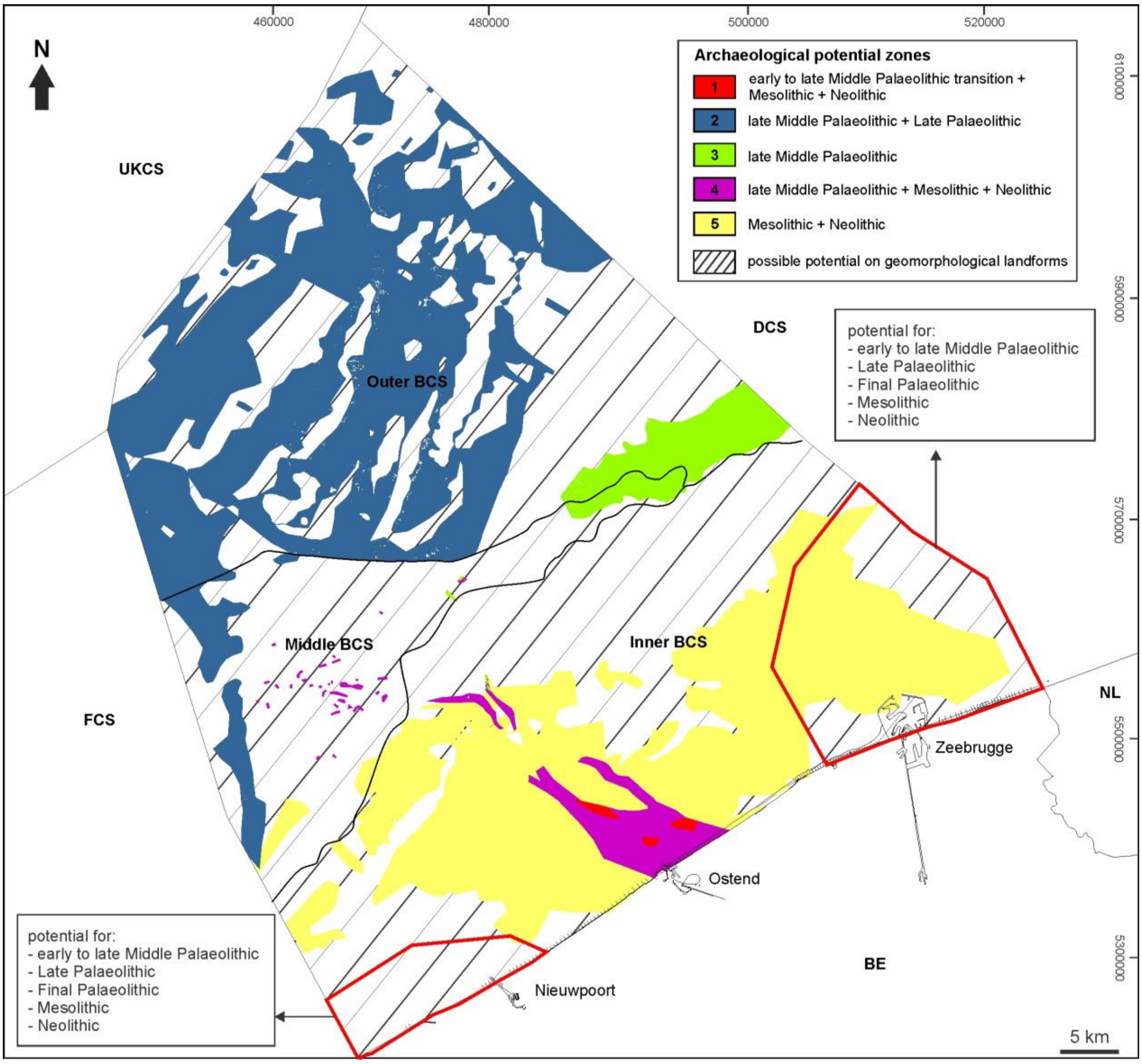


Figure 8.5 Archaeological potential map of the BCS with indication of the thickness of the Holocene sediments. Please note that in this map the Holocene unit U4 of Mathys (2009) as this unit was used to discuss the archaeological preservation rating in Chapter 7.

An example of an affected high-ranked potential zone is APZ 3, which is located underneath the Thornton sandbank. Here an Early Eemian palaeoshoreline with a wave-dominated estuary is preserved next to a Paleogene hill, a suitable landscape for occupation. On top of this area a wind farm was constructed in the period 2007–2013 (C-Power). The foundations of this wind farm penetrate deep in to the Holocene sediments and maybe even penetrate the upper part of the Pleistocene layers (Le Bot et al., 2005) thus possibly affecting the archaeological potential valuable layers. Future study of this area is now much more difficult as access to the area is restricted, and the area may therefore never reveal its full archaeological potential. A good collaboration between the scientific community and offshore industry is necessary. This way both parties can benefit from planned construction works at sea. The scientific community may benefit from new archaeological (and possibly palaeontological) information, while the industry saves time and money with improved knowledge of the landscape and its possible preservation potential. On top of this spectacular archaeological finds provide good public relations.

A good example is the so-called ‘Zeebrugge project’ within the framework of the Complex Project Kustvisie. A collaboration agreement between the Flemish Government and Ghent University was set up. A number of deep sediment cores and cone penetration tests have been planned for the summer of 2018. Scientists will use this information to create a three-dimensional model of the subsurface. At the same time the cores will be used to reconstruct the landscape and evaluate the archaeological potential of the area. Up to now this archaeological potential for the area was largely unknown but appears to be high in view of the Scheur findings (see Chapter 7).

8.3.3 Palaeontological Potential Map

The palaeontological potential map (Figure 8.6) is based on the age of the units and the possible occurrences of mammoth, elephant, Holocene, walrus and marine faunas as described in Chapter 7. Marine units have been given the potential to contain marine fauna material, while the Holocene units have been given the potential to contain a Holocene fauna. Eight different palaeontological potential zones (PPZ) with various faunas have been distinguished. These zones are relatively similar in extent to those for the archaeological potential but now also include layers that consist of marine sediments where a marine fauna is expected (PPZs 1, 3, 5 and 8).

Although the marine fauna up to now is restricted to ray-finned fish it cannot be excluded that other marine animals have been preserved within these layers. For this reason the marine fauna has been divided into the mammalian walrus and non-mammalian marine fauna. The mammoth and elephant faunas are restricted to fluvial environments that predate the Holocene. Finally, the Holocene fauna is expected in areas where Holocene non-marine sediments are preserved. The fauna is only expected on the Inner BCS.

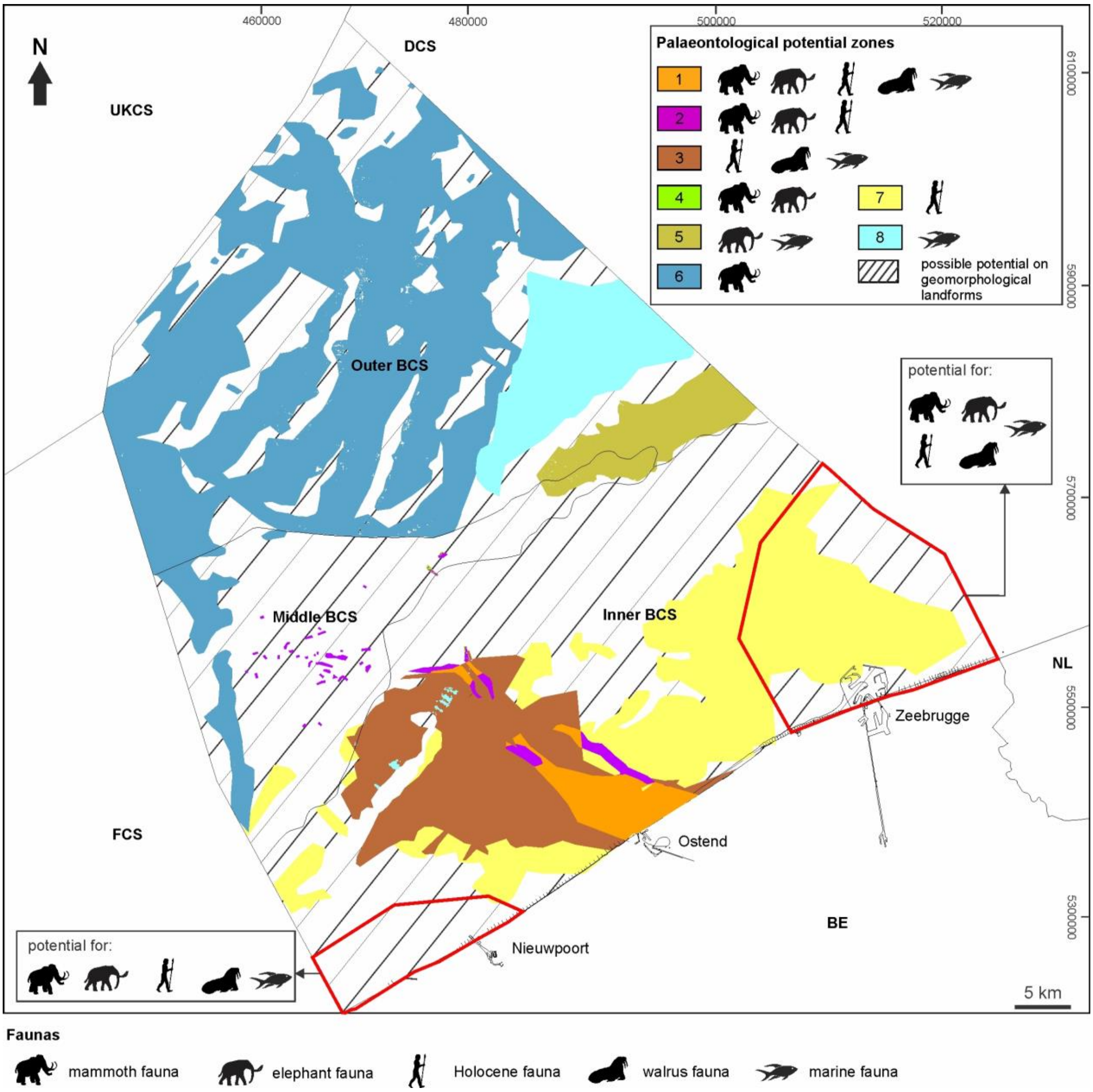


Figure 8.6 Palaeontological potential map of the BCS that indicates where faunas can be expected based on the preserved stratigraphic units. Please note that in this map the Holocene unit U4 of Mathys (2009) was applied as this unit was used to discuss the palaeontological occurrence of fossil material in Chapter 7.

8.4 How reliable are potential maps?

The archaeological and palaeontological potential maps provide an indication what can be preserved within the Quaternary sediments of the BCS as a result of the spatio-temporal impact of the syn- and post-depositional processes caused by landscape evolution. The final result is based on a stratigraphic framework constructed in Chapters 5 and 6 and the landscape analysis vs. location of the finds in Chapter 7.

The archaeological potential map cannot be ground-truthed because until now no archaeological material from the Palaeolithic-Neolithic period has been found offshore. The presented map is of a primary nature and is based on a stratigraphic framework, landscape analysis and extrapolations from taphonomic observations from onshore Flanders. It is possible that within each APZ specific zones exist that deviate from the assigned potential (f.e. caused by local post-depositional processes of which we are yet unaware of). Future archaeological finds will therefore help refine the primary map that has been presented here.

The palaeontological potential map on the other hand is the result of a landscape analysis and a listed archive of palaeontological finds (see Chapter 7). Most of the fossil finds appear to relate to the buried palaeo-Scheldt Valley and its downstream branches and to a lesser extent also the buried Yser Valley to the southwest. This demonstrates that, based on the present state of knowledge, a relationship exists between palaeontological bone material and buried valley structures (e.g. Vermeersch et al., 2015). This is likely a taphonomic effect as described by Gautier (1974, 1985) and Germonpré (1993b) for the upstream Flemish Valley. However, the Scheur find location is not directly related to a buried palaeovalley. It is more located in between the Zeebrugge and Waardamme-Lys Valleys and is the only location that shows indication of preservation in a primary context, opposed to the secondary context within the Flemish-palaeo-Scheldt Valley system). Though the stratigraphy at this location is not yet understood it provides proof that bone material can also be expected in large quantities outside buried valley systems when syn- and post-depositional processes are favourable. This warrants caution when interpreting the palaeontological potential map that even outside valley systems lots of material may be preserved and that a clear study of the palaeoenvironment must be performed before taking any further actions.

8.5 Conclusions

This chapter presents two models that indicate the archaeological and palaeontological potential of the Quaternary sediments of the BCS, and are based on a basic profile type map building on the results of Chapters 5, 6 and 7. Both maps can be used as a guiding tool during prospective investigations for areas at sea pending future exploitation. As a tool, these maps can be used as a guideline as to how extensive these prospective investigations have to be, saving time and money during this stage of planning. In case of high potential areas it would be beneficial to collaborate with the scientific community, as is the case for the Zeebrugge project, for an optimal chance of success. On the other hand, a certain degree of scepticism is needed as these maps do not provide the full story and are based on the present state of knowledge. Future geophysical investigations provide additional information that may change these maps. This is especially the case for the western and eastern areas of the Inner BCS.

No standardised process exists to determine the archaeological and palaeontological potential of the continental shelves. Recently, an indicative model for the archaeological potential was made for the DCS based on a complicated point system of vertically stacked landscapes (van Heteren et al., 2016). However, that system could not be applied to the BCS because of the different stratigraphic situation. In that sense every potential map, be it for archaeology or palaeontology, is unique for each continental shelf.

9 DISCUSSION, CONCLUSIONS AND OUTLOOK

9.1 Discussion

The main aim of this dissertation was to create a first order assessment of the submerged heritage of the Belgian Continental Shelf and to visualise this into a map. In order to do this several intermediary issues needed to be addressed. Because of the structure of this dissertation, the data gathered throughout this period and the geological architecture of the Belgian shelf (Inner, Middle and Outer) these intermediary issues are discussed throughout Chapters 4 to 8. This concluding chapter will therefore provide an overview of all the research aims as postulated in Chapter 1. Next, as an epilogue to this dissertation, an overall conclusion is presented alongside the wider implications of this research, and what these results imply for the (geo)archaeological scientific field. Finally, a scientific reflection on the approach of this dissertation is given.

Preservation of prehistoric landscapes from two end-members: fluvial and marine

Since the 1960s it was widely accepted that the Belgian shelf had not preserved many prehistoric landscapes, and that those that had been preserved were largely reworked. Indeed, it was acknowledged that past terrestrial sedimentary environments, that had developed during glacial low-stands, were almost always completely removed and reworked from the sedimentary record during the subsequent interglacial transgression. This dissertation has demonstrated the opposite. The sediment-starved Belgian shelf has preserved a wide variety of prehistoric landscapes that were formed across full glacial-interglacial cycles, within a tectonic 'stable' setting (i.e. the North Sea Basin shoulder), only affected by glacio-isostatic adjustment from ice masses located to the north of the study area. These landscapes are the combined result of climate change, climate-induced sea-level changes, and glacio-isostasy and to a lesser extent also tectonics and include erosional geomorphological landforms carved in the base Quaternary substratum (e.g. top of the Paleogene) and date from the Late Saalian to the Holocene (MIS 6–1). They lie in between the end-members of terrestrial and marine environments and comprise braided to meandering fluvial deposits, GLOF-deposited gravel fields and near-coastal to shallow marine environments.

The depositional units with the greatest preserved volume derive from the marine end-member (subunits BCS1a, BCS1b facies 2 and 3 and SCH4a, unit SCH2 and SCH3). These units are spatially spread across the base Quaternary substratum where they form an assortment of near-coastal to shallow marine geomorphological landforms, and where they infill fluvial incised-valley systems carved into the substratum during previous glacial periods as a result of falling sea levels. The origin of these near-coastal to shallow marine depositional environments was derived from seismic (acoustic facies) and sedimentary data (primarily shell material). An assessment of the conditions (e.g. high-energy or low-energy conditions) during deposition is derived from the degree of abrasion of the shell material. During transgression incised valleys increase their accommodation space. Within the study area, accommodation space was likely even larger as a result of glacio-isostatic relaxation, resulting in a rapid infill of these valley systems.

Sediments of the fluvial end-member are mostly correlated with erosional landforms (e.g. river terraces and elongated palaeodepressions) alongside a complex network of incised-valley systems. Preservation of fluvial sediments appears to be ubiquitous for a sediment-deprived shelf environment positioned on a tectonic basin shoulder (units SCH1, SCH2, BCS2, BCS3 and subunit BCS1b facies 1). They are dominantly built of fine- to medium-grained sands with local levels of organics, clay, silt and gravel derived from areas to the north of the study area. Coarse-grained sands have only been observed on the Outer BCS. On the Belgian shelf this type of sediments has two geographic origins: the Belgian interior (Flemish Valley) and the northern section of the southern North Sea (Axial Channel). The latter river system finds origin in the Alps (Rhine), Eifel (Meuse) and large sections of the North European Plain (the so-called Urstromtal). Finally, finer-grained sediments are more confined to topographically protected settings such as estuaries (units SCH4, BCS1b), abandoned channels, pools and floodplains (units BCS2–3, SCH5–6).

A wide range of processes determines how the Belgian antecedent palaeogeography developed and how it evolved. On a regional scale, processes like tectonics and glacio-isostasy exert a major influence on the landscape development. The thin and fragmented nature of the Quaternary cover on the Belgian shelf is the result of its position on the North Sea Basin shoulder, a tectonically stable area marked by near-zero net subsidence/uplift (i.e. the hinge zone of the basin; Cohen et al., 2014). The hinge zone itself received large amounts of sediment, delivered by rivers originating in upstream areas of uplift (Cohen et al., 2014), though the BCS is not connected to the uplifted areas through major river systems such as the Rhine and Meuse in the Netherlands and therefore received less sediment. Furthermore, the area between Belgium and East Anglia has been marked by mild uplift depriving it of sufficient accommodation space to preserve such a layer-cake stacked sequence of sediments. Now, this is only possible in deeply incised-valley systems such as the palaeo-Scheldt Valley and its offshore extension.

Glacio-isostatic adjustment forms a second controlling factor of differential vertical crustal movement, next to tectonics, operating at the same time scales as the glacial-interglacial cycles, driven by the growth and decay of ice masses to the north and the associated changes in mass distribution over the Earth’s surface (cf. Cohen et al., 2014). During each glaciation, the crust below the ice-sheet is depressed by the weight of the ice. In compensation, the peripheral zone around the area of crustal depression experiences forebulge upwarping. This was for example the case during the Elsterian and the Late Saalian when the area between respectively northwest France-Britain (Moreau et al., 2015) and Belgium-East Anglia was uplifted. During deglaciation, and for some continued relaxation time thereafter, the suppressed area rebounds while in the periphery the forebulge relaxes. The loading and unloading of shelf areas with the weight of seawater causes further perturbing warping effects (hydro-isostasy; cf. Cohen et al., 2014), though this subject has not been addressed in the presented research. Together, glacio- and hydro-isostasy controlled the timing and speed of transgression, the positioning of continuously moving palaeoshorelines at specific times and the depths at which sedimentary indicators of these former sea levels are preserved today (e.g. Lambeck, 1995; Sturt et al., 2013). Glacio-isostatic adjustment also explains the complex pattern and differential timing of the Eemian transgression along the southern North Sea palaeoshorelines as seen on the Belgian and Dutch shelves (cf. Kiden et al., 2002). The magnitude of warping varies spatially and causes palaeoshorelines of the same Eemian age to differ many metres in elevation. These processes and conditions will have differed between glacial cycles. The relatively close location of the study area to the successive ice-sheets makes glacio-isostatic adjustment patterns differ more strongly from one glaciation to the next. In our study area this is for example strongly expressed between the Saalian and Weichselian glaciations.

These patterns of rather short-term differential crustal movement have strongly influenced the preservation of the Quaternary stratigraphic record. The long-term mild tectonic uplift/stability creates unfavourable conditions to the durability of interglacial deposits over multiple glacial-interglacial cycles. However, sediment traps, such as the deeply incised-valleys of the palaeo-Scheldt and Thornton Valleys that are located relatively far offshore, are able to preserve sediments of the early part of the interglacial. This was, however, not the case for the Holocene because no incised-valley system was present on the Belgian shelf.

During glacial inception, when relative sea levels fall, valley incision is promoted across the entire shelf. When the glacial period nears its termination the ice masses to the north start to disintegrate, promoting ice-front oscillatory behaviour, resulting in episodes of increased fluvial discharge and sediment transportation from the North European Plain (source) down to the European continental shelf edge (sink), e.g. the Bay of Biscay (e.g. Toucanne et al., 2009a, 2009b, 2010, 2015). Next to episodes of valley incision also other processes influence the northwest European drainage system during glacial periods: 1) the configuration, volume and location of the ice masses; 2) glacio-isostatic adjustment of the landscape; 3) antecedent palaeotopography, such as a land bridge positioned between East Anglia-Belgium and southwest Netherlands. These principal temporary landscape elements control the drainage configuration and its discharge from source to sink, e.g. from northeast to western Europe.

Sediment storage capacity during a sea level lowering is low within the incising valley systems. Sediments are preferentially removed and transported across the shelf environment from source to sink. During the glacial-to-interglacial transition, when climate ameliorates, and sea levels start to rise, sediment accumulation takes place along palaeoshorelines in shallow marine, near-coastal and estuarine depositional environments. Here, the balance between sedimentation and erosion is driven by the ratio between sediment supply and available accommodation space (e.g. Mathys, 2009 for the Holocene transgression of the study area), which in turn is influenced by vertical land movements such as tectonics and glacio-isostasy. During the full interglacial period, between sea-level rise and high-stand, incised-valley systems will become inundated and submerged/buried increasing their capacity to store large volumes of sediments. When the cycle is complete and glacial sea level lowering occurs again these sediments will be removed and transported again if the fluvial environments retains its former position.

In the study area the configuration and depths of incised-valley systems does not appear to be strongly governed by the consolidated pre-Quaternary strata of Paleogene age, though several examples of substratum control do exist. For example, the upstream Coastal Valley section of the palaeo-Scheldt Valley is oriented nearly perpendicular to the Paleogene strata, while the downstream Ostend Valley section is oriented oblique to the strata at an angle of ca. 45°. It remains unclear what the controlling factor of this change in orientation is though it can safely be assumed that the substrate exerts some control as the homogeneous compact Paleogene clay units Y1 and Y2 are much more erosive-resistant compared to the younger layers to the east, that contain larger concentrations of sand, sandstone and silt.

Local to regional palaeogeographic correlations

The newly constructed Quaternary stratigraphy of the Belgian shelf provides new and additional information as to how the northwest European landscape evolved and tests current hypotheses that have been developed over the past decades.

Saalian

The relatively rapid Eemian infill history of the palaeo-Scheldt Valley provides indirect evidence for Late Saalian glacio-isostatic forebulge upwarping between Belgium and East Anglia. The presence of a forebulge crest in this area resulted in a temporary land bridge that forms a sequel to the Elsterian breaching hypothesis of the Dover Strait (Cohen et al., 2014; Collier et al., 2015; García-Moreno, 2017; Gibbard, 1995; Gupta et al., 2017; Smith, 1985). The retrogressive erosion of the Paleogene remains of the Elsterian land bridge reached up to the area between Belgium and East Anglia, where a denudated topographic barrier between the English Channel and the southern North Sea remained in place. During the Late Saalian, the height of this land bridge was influenced by glacio-isostatic flexural adjustment of expanding ice masses to the north. Forebulge upwarping affected the palaeo-Scheldt Valley system in Flanders-the Netherlands causing the river system to deflect westward in the direction of the Dover Strait. The timing of deflection and the underlying glacio-isostatic process is different than the current hypotheses of the palaeo-Scheldt Valley formation (Gibbard, 1988; Mathys, 2009). Up to now it was hypothesised that the change in orientation of the palaeo-Scheldt River was caused by the breach of the land bridge by a proglacial lake and that the river system adjusted its course in the direction of the Dover Strait because this was the direction in which the dominant flow was oriented.

Eemian

The Eemian sediments that infill the Thornton Valley and the Ostend Valley section of the palaeo-Scheldt Valley demonstrate that two types of estuaries were present in the Eemian landscape during the transgression of the shelf environment: 1) Early Eemian (lagoon-shaped) wave-dominated estuaries that were likely protected by a barrier complex (e.g. the Thornton Valley); 2) Middle-Late Eemian tide-dominated estuaries (e.g. the Ostend Valley section of the palaeo-Scheldt Valley).

Lagoon-shaped wave-dominated estuaries form in a microtidal environment while tide-dominated estuaries form in a macrotidal environment (Boyd et al., 2006; Dalrymple and Choi, 2007). The current macrotidal environment was established ca. 8–8.5 ka ago when the southern and northern North Seas connected (Stride, 1989). Before both seas connected the tidal range in both seas was likely much lower (Denys and Baeteman, 1995; Shennan et al., 2000). The regional palaeogeographic scenarios of the Eemian interglacial suggest that a similar situation occurred. This is corroborated by the microtidal Thornton Valley estuary and the macrotidal palaeo-Scheldt estuary that suggest that the southern and northern North Seas connected somewhere during the Early-to-Middle Eemian transition.

Weichselian

The far-travelled erratic clasts ranging up to 3 m in diameter at the foot and on top of the Offshore Scarp terrace demonstrate that a high-energy transportation mechanism took place somewhere during the Weichselian. This transportation mechanism was likely fluvial in origin and requires the presence of large volumes of water possibly in the form of a proglacial lake in front of a merged FIS and BIIS. The present hypothesis states that the FIS and BIIS may have merged during the Early Pleniglacial (ca. 70 ka; Carr et al., 2006; Mangerud et al., 2011; Svendsen et al., 2004) and the LGM (30–18.7 ka; Clark et al., 2012; Sejrup et al., 2016) and that large volumes of water, channeled through the Axial Channel trunk valley, were transported to the Dover Strait. The edge of this trunk valley is situated 13 to 19 meters below the foot and top of the Offshore Scarp terrace and suggests that a high-energy transportation mechanism with large volumes of water was needed to exceed the confinements of the Axial Channel trunk valley. The braidbelt units BCS2 and BCS3 in combination with the palaeodepressions scoured in the pre-Quaternary substratum suggest a northern origin of this flood, almost parallel to the Axial Channel trunk valley. No geochronological control is available for these units making it hard to date these sediments and their formation process(es). However, the presence of the braidbelt deposits in combination with merged ice-sheets during the Early Pleniglacial and LGM suggest that a proglacial lake may have been present at least once in the southern North Sea. This implies that more than one event may have taken place that resulted in the formation of unit BCS2, BCS3 and the far-travelled erratic clast field.

No antecedent high topography was present in the southern North Sea like during the Elsterian and Saalian glaciations. However, in both cases the position and confinement of a proglacial lake was controlled by glacio-isostatic forebulge upwarping (Moreau et al., 2015). No evidence of moraines is preserved in the landscape that would indicate the position of a lake boundary. It can therefore not be excluded that evidence of these were removed from during one or multiple GLOFs. The presence of a moraine in the Early Pleniglacial southern North Sea is hypothesised by the southern position of the BIIS south of the Dogger Bank by Cotterill et al. (2017). This in combination with the unknown position of the forebulge crest may support the presence of a proglacial lake.

Several authors (García-Moreno, 2017; Patton et al., 2017; Sejrup et al., 2016) hypothesise that a proglacial lake was also present during the LGM, however no supporting information is provided why this lake may have been present. In Chapter 6 the presence of this lake and its possible conditional parameters such as forebulge position was discussed. However, this time no clear indication of a moraine is present because no evidence of a more southern position of the ice-sheet exists.

In Chapter 6, a transportation mechanism for the far-travelled clasts is discussed. This mechanism explains the position of the clasts by one or more GLOFs. At its head a GLOF is composed of a dense mixture of sediments and gravels that have a high transportation capacity possibly for boulders up to 3 m in diameter. However, for such large clasts to be transported on a relatively flat floodplain like the Outer BCS an extremely high velocity is required to maintain this transportation capacity. Alternatively, icebergs transported by the GLOF can be invoked as a secondary transportation mechanism. During each deglaciation the ice front calved off and icebergs, carrying sediments and gravels, were likely to be present in these proglacial lakes. When the lake breached the icebergs were transported south and deposited at the foot and top of the Offshore Scarp, where the latter was an ideal location to trap erratics and icebergs due to a bend that is present in the Axial Channel trunk valley system.

Prehistoric landscapes

For decades published research has established that preservation of prehistoric landscapes and their depositional environments was very low for the Belgian shelf environment. This principle was based on the notion that, on a shallow sediment-deprived continental shelf with a relatively low accommodation space, lots of sediment recycling occurred. As a result, researchers interpreted that mostly geomorphological landforms (mostly scour hollows) filled with marine sediments were preserved. This bias towards low preservation potential, where depositional environments are constantly recycled over multiple glacial-interglacial cycles can now be corrected. A similar observation was made on the erosional eastern English Channel shelf by Mellet (2012).

To the area north of the basin shoulder, where subsidence dominates, sediments actively accumulate to form thick sedimentary sequences that have a high preservation potential (Cohen et al., 2014; Hijma et al., 2012). These depositional environments contain high-quality palaeonvironmental archives that are useful for palaeogeographic reconstructions (Busschers et al., 2007, 2008; Hijma et al., 2012; Meijer and Cleveringa, 2009; Peeters et al., 2015, 2016). On the other hand, the subsidence of the area makes it more difficult to access progressively older sediments as they become buried deeper.

The area positioned on and south of the basin shoulder, experiences stability to mild uplift, promoting fragmented sedimentary archives within fluvial terraced landscapes (cf. Bridgland, 1994, 2000). On the Middle-Outer BCS, an incoherent set of terraces and escarpments with various orientations are preserved that allow the construction of a relative chronological framework. Where sediments overlie stepped surfaces, it should be possible to establish a chronological stratigraphy using OSL, and possibly also pollen, as a framework to constrain these enigmatic erosive episodes. If this were to be done for terraces T1 and T2 and palaeodepression OHD a depositional sequence may be established for almost the entire Outer BCS. It has to be noted that not every erosive event in geological history is characterised by a net vertical incision. The Offshore Scarp for example was incised during the Late Saalian, but its western approaches were likely reused as a geographical boundary for the enigmatic GLOFs that occurred during the Early and Late Pleniglacial. Here, it is unclear if the Early-Late Pleniglacial GLOFs further eroded the base and top of the Late Saalian fluvial terrace.

Unravelling the fragmented sedimentary record of the Belgian shelf into a coherent chronological palaeogeographic map series is a precarious endeavour. At only a few locations, where sediments are preserved above a distinct geomorphological landform a connection between preservation potential (i.e. erosion vs. deposition) can be made. However, as mentioned above, sediments preserved above terraces are vulnerable to recycling events over multiple glacial-interglacial cycles creating a more complex geological architecture, including hiatuses. This geological setting for the study area and the area of the Dover Strait and English Channel (e.g. Mellett, 2012; Mellett et al., 2013) is different from classical studies performed in a subsiding basin where fluvial response is strongly linked to climate change and sea-level variations (e.g. Busschers et al., 2007; Peeters et al., 2016).

Transgressive and high-stand sediments

In Chapters 5 and 6 it is demonstrated that predicting the preservation potential of the highly diverse buried Quaternary prehistoric landscapes preserved on the BCS is an arduous undertaking. Several studies have highlighted that near-coastal depositional environments, such as the Brown Bank Formation (e.g. Hijma et al., 2012; Laban, 1995; Peeters et al., 2016) and the Dutch Channels unit MIS5 (e.g. Hijma et al., 2012) are the result of prograding palaeoshorelines during regression. However, this is not the case here due to a lack of sediment supply and the relatively limited accommodation space outside incised-valley systems. Furthermore, this dissertation demonstrates that near-sediments have a high preservation potential during a transgression as demonstrated in Chapters 5 and 6 (e.g. Hijma et al., 2010; Mathys, 2009; Mellett, 2012; Mellet et al., 2013; Roy et al., 1995; Storms et al., 2008).

In the study area three types of transgressive sediments are partially or well-preserved. The Thornton Valley infill (unit BCS1b) describes a microtidal wave-dominated estuarine environment that preserves a complete drowning sequence during the early part of the Eemian transgression. Part of the upper estuarine sequence was removed by the flooding and wave-ravinement surfaces. As this sequence likely dates to the Early Eemian transgression in combination with post-Saalian forebulge collapse this area may have been transgressed relatively fast. The Ostend Valley is a second example of a partially preserved macrotidal estuarine sediments during the Middle-Late Eemian transgression (units SCH2, SCH3, and subunits SCH4b and SCH4c). Such environments experience strong reworking of older sediments. However, pollen analysis has demonstrated that a good chronology is preserved in this area, although sterile zones do occur. OSL dating on the other hand provided an underestimated age without a clear chronology but provided a clue to the sedimentation rate of the transgressive sequence. The last transgressive environment concerns sediments from the marine environment. These have been preserved as subunits BCS1a and SCH4a. Considering the regional palaeogeography subunit BCS1a has been deposited in a microtidal shallow inland sea, while subunit SCH4a was deposited in a macrotidal regional sea. Both units were deposited relatively close to the coastal area at the time of transgression so no deeper marine sediments are preserved.

Regressive sediments

On the BCS no forced marine regressive sediments have been identified that date from the gradual sea-level fall of the Weichselian Early Glacial. Sediment supply was likely insufficient and most available accommodation space was already filled by the previous Eemian transgression and high-stand. On the other hand the slow regression allows lateral migration of fluvial channels as has been observed in the palaeo-Scheldt Valley fluvial unit SCH5. A similar observation is made to the north of the BCS where the Rhine-Meuse River system developed a delta system in the southern North Sea Basin as a result of lack in accommodation space upstream in the river system (Peeters et al., 2016). On the other hand, when sea level drops relatively fast during the Early Glacial Rederstall stadial, fluvial incision occurs that is expressed as shallow channels incised in the Paleogene substratum of the Middle BCS. The base of these channels is filled with reworked medium-grained sands of older deposits. During the following transgression of the Odderade interstadial a fluvial wetland develops in which fine-grained sediments and peat are able to accumulate. These regressive sediments have been preserved in only a few sections of the Middle BCS valley system. Post-deposition of unit SCH5, terrestrial and marine processes have eroded large sections of the palaeovalley system infill as the palaeovalley system became exposed. However, the presence of roots in the peat layers shows that the soil is quite mature and provides additional resistance to erosion.

Low-stand sediments

The most common low-stand deposits preserved on the BCS belong to a braided river system (unit BCS2 and BCS3) called the Axial Channel trunk valley. Considering the cross-sectional area of this river system the BCS belongs to the relatively flat and high area that can only be flooded during peak discharge, i.e. the floodplain. Several cycles of peak and normal discharge have been registered in the sedimentary sequences of unit BCS2 and BCS3. The peak discharges were highly erosive, possibly linked to GLOFs, and have removed all older Eemian marine sediments in their path by incorporating them in their own sediment body or transporting them downstream to the FCS. These sediments are redeposited as the discharge subsides and when lateral reworking occurs within the remaining channels of the floodplain. Furthermore, sediments of normal and low discharge of the Axial Channel trunk valley are not preserved on the BCS as the river system is concentrated in the channel itself and the floodplain is abandoned. During this floodplain abandonment pools may survive on the floodplain for a prolonged time that are preserved as isolated sediment bodies with high concentrations of organic material and laminated sands, clays and silts.

Erosional landforms

The position of the BCS on the southern North Sea Basin shoulder resulted in many erosional landforms cut in the Paleogene substratum. Sediment bodies do not accompany most of these as they have been removed during later glacial-interglacial cycles. Examples of such barren erosional landforms are the Middle and Offshore Scarps. The Offshore Scarp is for example an erosive landform that started out as a fluvial terrace during the Late Saalian that has been re-used as a coastline during the Eemian interglacial (and possibly also during the Early Glacial and the Holocene) and as a fluvial system boundary during the Early Pleniglacial and the Late Glacial Maximum. From these depositional environments only sediments of this last stage have been preserved. The Middle Scarp may even be older than the Offshore Scarp (Hijma et al., 2012). Similar to the Offshore Scarp it may have started out as a fluvial terrace that has subsequently been reworked by marine processes thereby removing all original sediments. These erosional landforms are formed out of the consolidated Paleogene strata that are much more erosive-resistant compared to the unconsolidated Quaternary sandy sediments.

To conclude, the prehistoric landscapes of the Belgian shelf, alongside their geomorphological landforms and depositional environments, have a relatively high preservation potential over a timescale of two glacial-interglacial cycles. The results of this dissertation demonstrate that present ideologies, where the impact of sea-level rise and its erosional effect on previous low-stand deposits have been overestimated, and that, though the Belgian offshore stratigraphic record is generally highly fragmented, a relatively high preservation potential exists to preserve sediments of preceding glacial-interglacial cycles. Incised-valley systems offer the best preservation conditions due to the creation of accommodation space. The Ostend Valley section of the palaeo-Scheldt Valley system is a good example of this, and particularly its offshore position offers great potential to have preserved transgressive marine and near-coastal sediments. Based on the palaeogeographic map series it is possible to predict where preservation potential would be optimal or absent, though this greatly depends on other facts such as inherited palaeotopography, the transition of wave- to tide-dominated systems, deflection of river systems, etc. This dissertation has demonstrated that near-coastal and fluvial prehistoric landscapes can be preserved over multiple glacial-interglacial cycles and that, even though spatially fragmented, important sections of the stratigraphic record can be preserved. These results now shed a new light for future research within this field.

Submerged heritage

Determining the preservation potential of submerged heritage requires a multidisciplinary approach of the sedimentary environment and a thorough understanding of the structures and functions of these in the overall landscape development. Here, the Quaternary stratigraphic record ranges from the Late Saalian to the Holocene, though the depositional landforms of the latter have not been the focus of this dissertation, we did incorporate them in Chapters 7 and 8. The sedimentary information contained within the stratigraphic record now provides a geological context. Now, for the first time ever, the preservation potential for submerged heritage based on inferences made by Ward and Larcombe (2008) for the wider southern North Sea could be made for the study area. This is a prerequisite for solid inferences on and predictions of the submerged heritage potential. Some degree of scepticism is needed though as many uncertainties and ambiguities such as preferential land use by hominids have to be inferred.

The Belgian offshore stratigraphic record shows an irregular, patchy distribution pattern and strongly varying depths of layer bases. In other words, the use of simple layer-cake-style (cf. Berné et al., 1994; Trentesaux et al., 1999) or depth-relates-to-age expectation models (cf. Mathys, 2009) do not suffice anymore. Such depositional records are hard to unravel and both the lithology and its submerged heritage content often show many elements reworked from earlier phases and require palaeogeographical reconstruction to stratify and order them, as attempted in this research.

The research on the submerged heritage potential has determined that the Middle Pleistocene record is largely absent on the Belgian shelf. This is due to the tectonic basin shoulder setting in combination with a sediment-depleted shallow shelf environment and episodes of enigmatic drainage network alterations during the Saalian and Weichselian glaciations. This resulted in progressive reworking and net erosion of sediments in the study area thus lowering its preservation potential for buried heritage. Additionally, these processes likely displaced a lot of submerged heritage from its primary context. The Late Pleistocene record, although fragmented, is much better preserved. Within this landscape sediments with a high preservation potential for submerged heritage are more common. This is for example the case at the Scheur, which appears to be an area where material is present in primary context. Other locations from the Eemian interglacial with a high preservation potential may be located in the Thornton Valley and along the estuarine margins of the palaeo-Scheldt Valley system and the drowned Marginal Platform that transformed into islands.

The Early Glacial landscape is still not well understood. Sedimentary evidence shows that fluvial sediments are preserved upstream the palaeo-Scheldt Valley and that many fossil bones originate from this time period though they cannot be directly linked to each other. This is supported by the sedimentary and seismic characteristics of this environment. The clay and peat layers downstream likely represent pristine preservation conditions to preserve submerged heritage. This paradox is yet to be resolved.

The Pleniglacial landscape was characterised by high-energy fluvial environments. The drop in sea level resulted in reworking of older deposits in river systems. This process removed much of the former Eemian sandy infill in the river valleys thereby destroying much of its preservation potential. However, as the upper Eemian sandy infill consists of dominantly marine or estuarine sediments this is not problematic. The process of incision and reworking resulted in a specific taphonomic process through which bones are accumulated within the river system (primary to secondary context). This process was first observed and described in the Flemish Valley by Gautier (1974) and Germonpré (1993b). The palaeo-Scheldt Valley forms the downstream section of the Flemish Valley. This suggests that the same process likely occurred in this downstream section. However, as the activity of the palaeo-Scheldt River was displaced from the Ostend Valley to the Zeebrugge and possibly also the Waardamme and Lys Valleys it is suspected that similar bone accumulations are preserved at these locations. The bone material found in the Zeebrugge navigation channel supports this. On the other hand, the abandonment of the Ostend Valley may have resulted in shallow pools and abandoned river branches in which fine-grained sediments were able to settle. Such environments have a similar high preservation potential.

The GLOF-induced sedimentary units on the Outer BCS were formed in a very high-energy environment. During these GLOFs older sediments were reworked and/or removed from the stratigraphic record. These sediments were likely marine in origin (as testified by the mollusc fauna) and did not have a high potential for submerged heritage. However, the possibility of multiple episodes of GLOFs may have removed the low-energy sediments that form each time when the high-energy environment dissipates. In other words, the braidbelt deposits on the Outer BCS have a highly uncertain preservation potential. It is likely that the upper section of these sediments may preserve submerged heritage.

Similar to the Eemian transgression the Holocene transgression removed parts of the upper sections of the exposed deposits. This process likely reworked and/or removed part of the submerged heritage. The Holocene landscape is represented in the stratigraphic record by marine and back-barrier deposits. The marine deposits have negligible preservation potential, similar to the Eemian marine deposits. The back-barrier environment consists of a wide variety of sub-environments each having its own preservation potential for submerged heritage. On top of that a back-barrier environment is highly dynamic with areas of erosion and deposition at very short distances. Even with a very dense seismic and sediment core information grid it remains virtually impossible to infer areas with low and high preservation potential.

9.2 Conclusion and wider implications

This dissertation has contributed considerably to the research fields of palaeogeography and continental shelf prehistoric research and provided new daring insights into the Quaternary evolution of the southern North Sea Basin, specifically over the last glacial-interglacial-glacial cycle (e.g. Saalian-Eemian-Weichselian). Our results demonstrate that a wide series of geological processes are preserved in the stratigraphic record as prehistoric landscapes, erosional landforms and sedimentary environments. The presented research still lacks the support of a robust absolute geochronology, however the use of pollen dating, cross-cutting sequences and reworked shell material, in combination with regional-scale geological processes (e.g. palaeogeographic feedback to climate change, climate-induced sea-level changes and glacio-isostasy), has helped to better constrain the relative chronology of the Belgian offshore stratigraphy. This new relative stratigraphic framework and the effect of regional- to local-scale geological processes were analysed to determine the impact on the preservation of prehistoric landscapes and the submerged heritage within.

The late Middle-Late Pleistocene regional and local palaeogeographic map series demonstrate the complex nature of geological events (erosive and depositional) that may take place during one single glacial-interglacial cycle within the southern North Sea Basin. Because of the location of the study area on the basin shoulder only fragmented pieces of these processes can be preserved thereby under-representing them in the stratigraphic record, yet still providing important clues to their nature and origin. Alhough our stratigraphic model is not representative for other areas in the basin shoulder it may yet provide clues to determine what sedimentary processes are operating on it and what impact they have on the preserved sedimentary environments.

In this dissertation, the stratigraphic model, though specific for the Belgian shelf, was a specific guiding tool for the development of a palaeogeographic map series on a local to regional scale over the past ca. 160 ka. This new map series now provides a new solid ground to further improve our understanding of the late Middle-Late Pleistocene geological history, not only of the Belgian shelf, but also of the southern North Sea Basin as a whole, down to the Dover Strait and English Channel region (e.g. Mellett, 2012).

Finally, this dissertation has demonstrated the use of palaeogeographic map series for the study of submerged heritage and its preservation potential within prehistoric landscapes. The applied technique aids in our endeavour to determine where, and more specifically why, submerge heritage can be expected on the continental shelf environment and what its context to the past landscape and its geology is. The involved processes provide clues to behaviour of the biosphere to geological processes, human migration patterns across dynamic landscapes over glacial-glacial-interglacial cycles and taphonomic processes.

9.3 Archaeology

The palaeogeographic map series and the palaeoenvironmental analysis presented in this dissertation demonstrate that ancient coastal environments similar to those in East Anglia (0.5–1 Ma) have been preserved on the Belgian shelf (f.e. the Early Eemian Thornton Valley). The discovery of numerous artefacts as well as footprints preserved in estuarine clays testifies that these ancient coastal environments of East Anglia were suitable occupation sites provided that the post-depositional preservation of these environments was suitable. Due to the parallelism between the Eemian palaeoshoreline development and the good post-depositional preservation conditions in the Thornton Valley, but also in areas as the Scheur, it can be expected that archaeological material is preserved at these locations.

The Zeeland Ridges Neanderthal originates from Middle Pleniglacial Rhine-Meuse braidplain deposits. The positive effect of the Rhine-Meuse braidplain contraction during the following Late Pleniglacial and only modest reworking of the top of the braidplain sediments during the Holocene transgression resulted in relatively good post-depositional preservation of the site. The floodplain deposits observed on the Outer BCS experienced a similar development under the influence of the Axial Channel trunk valley. In other words there is no reason to assume that no archaeological artefact or human bone fragments are preserved within these sediments.

If similar positive post-depositional preservation conditions occurred across the Belgian shelf than why have we not yet discovered any artefacts or bones? The archaeological discoveries in East Anglia were made on a receding shoreline. Each time the erosion uncovers new sediment layers that preserved archaeological evidence that can easily be reached. However, some of these finds are also found several kilometres inland. Similar archaeological discoveries from the much younger and technologically more developed Neolithic period have been made near the Belgian shoreline/dune belt, but are, based on the landscape analysis, likely related to a different palaeoenvironment. The Zeeland Ridge Neanderthal discovery, alongside other artefacts, is the result of aggregate extraction and the keen eye of amateur archaeologists and palaeontologists who are present during the sieving process onshore. Aggregate extraction on the Belgian shelf is mainly focussed on the crests and flanks of several Holocene tidal sandbanks (Hinderbanks, Thornton and Kwinte sandbanks, Buitenratel and Oostdyck), and is therefore considered not to contain any archaeological material. Occasionally, palaeontological material is found, such as the mammoth skull and the red deer antler, along the exposed Pleistocene cores of these sandbanks. This may be one of the main reasons why no prehistoric archaeological material has been discovered yet on the Belgian shelf.

9.4 Scientific reflection

9.4.1 Approach of the dissertation

In this dissertation a bottom-to-top structure was followed. In a first step the pre-Quaternary surface was constructed (Chapter 4) and updated from previous versions, while in a next step a palaeogeographic map series was constructed based on the pre-Quaternary surface and was combined with a sedimentary and seismic facies analysis of the overlying Quaternary deposits (Chapters 5 and 6). This updating process of the pre-Quaternary surface resulted in new discoveries concerning the development of this palimpsest surface. In the course of this dissertation the same pre-Quaternary surface was updated two more times (Chapters 5 and 6). This updating process was necessary to get a better grip on the geomorphology of the pre-Quaternary surface and later provided additional information about the processes that are responsible for their formation. In other words, new and more geophysical data from specific targeted locations resulted in new insights or altered observations from previous authors. More so, the updating process resulted in a new and improved understanding of the Quaternary evolution of the BCS. However, the continuous updating of such a surface cannot be pursued forever. At a certain point in time new data will not alter or modify the surface significantly and no new insights will come forth of this process.

The palaeogeographic map series on the other hand, a first in its kind for the Belgian shelf, can still be improved. New insights from sediment cores such as palaeoenvironmental information (pollen, shells, diatoms) and absolute dating (OSL, ^{14}C) will result in alterations of this map series. This is for example the case for the Eemian interglacial where the results are based on several simplifications such as the 4 m tidal range and the absence of tidal friction in the palaeo-Scheldt estuary during transgression. Additional sea-level index points from tidal environments and calculation of the compression factor for these points in the palaeo-Scheldt Valley will improve the presented first order Eemian sea-level curve. This, in combination with geophysical modelling of the glacio-isostatic impact of the Saalian ice-sheet will improve our understanding of the Late Saalian-Eemian (MIS 6–5e) landscape evolution of northwest Europe.

The palaeogeographic map series was an indispensable tool for this dissertation to better understand the syn- and post-depositional impact of geological processes during landscape evolution (Chapter 7). This was a necessary product to provide a geological context and understand why and where submerged heritage is encountered on the Belgian shelf (i.e. actual preservation of the deposits). The manner how incision, transgression and infilling have been described in the palaeogeographic map series does not alter the insights of the preservation potential of the record nor its age if only a sedimentary study or a seismic facies analyses were to be performed, however it does improve our understanding. For example, the new insights that were developed for the Scheur find location testifies that a palaeogeographic map series are a necessary constituent to understand preservation and why bone material is found at this location, even though we lack sedimentary and seismic information.

9.4.2 Current understanding landscape evolution

In Chapter 1 it was stated that the Quaternary stratigraphy of the Belgian shelf is very fragmented and is very difficult to reconstruct, whereas on the neighbouring French and Dutch shelves the stratigraphy is more continuous. It was my personal understanding that when reading the literature the Belgian shelf stratigraphy deviated strongly from those of our neighbouring shelves. This struck me as odd and the presented information or clarification for the decisions made on the Belgian shelf appeared sometimes insufficient (e.g. complete marine reworking of terrestrial deposits as a result of low accommodation space or tidal scouring for valley formation). To understand the landscape evolution of northwest Europe one needs as many jigsaw pieces as possible to understand the complete picture even if the studied area does not appear important, as was the view in the past for the Belgian shelf. The results presented in this study demonstrate that the fragmentary record of the Belgian shelf yielded new crucial jigsaw pieces on both local and regional landscape evolution: 1) a better understanding of the approximate location of the Late Saalian forebulge position; 2) the timing of the deflection of the palaeo-Scheldt River; 3) how glacio-isostatic relaxation and inherited Saalian palaeotopography influenced the Eemian sea-level rise both on the BCS and in the southern North Sea; 4) that we lack a thorough understanding of the Weichselian Early Glacial landscape both on a local and a regional scale; 5) that also intermediate glaciations and not only glacial maxima seriously influenced the landscape (Early Pleniglacial vs. LGM: MIS 4 vs. MIS 2); and 6) that glacio-isostasy and not only aeolian processes may have influenced the most recent deflection of the Scheldt River to its current position.

In Chapter 1 it was also mentioned that in recent years, several erosion hypotheses have been developed on the various pathways the major palaeodrainage systems may have taken during the Saalian-LGM period. Each study tried to determine this through different proxies (bathymetry, sediment cores, seismic reflection data, mechanistic modelling, isotope ratios, XRF, etc.) from various locations ranging from source (upstream of river systems) to sink (Bay of Biscay). This study also contributes to these series of erosion hypotheses by combining a swath of data consisting of bathymetry, depth-converted structure maps, palaeoenvironmental analysis, OSL-dating, archived sedimentary cores/vibrocores and landscape modelling at a location between the depositional southern North Sea and the erosive area of the Dover Strait. The presented study underpins the presence of a land bridge consisting of Paleogene clays between East Anglia and Belgium and that it was the combined result of retrogressive erosion of a chalk bridge at the Strait of Dover and glacio-isostatic updoming. The presented study however, does not inform on the time of breaching but demonstrates that the palaeo-Scheldt River deflected to the Axial Channel-Dover Strait before the breaching of the land bridge happened (ca. 155–140 ka) as a result of glacio-isostatic updoming (possibly during the Drenthe glaciation: ca. 160 ka). Similarly the timespan in which these events occurred could not be presented. However, evidence from the Bay of Biscay sink area by Toucanne et al. (2009b) suggests that, assuming there is a positive correlation between XRF and MAR measurements, this may have happened relatively fast. One may infer that, as a result of the Elsterian Dover Strait *breach event*, a primitive Axial Channel trunk valley may have already been incised to considerable depths between East Anglia and Belgium and that the glacio-isostatic upwarped section of this channel on the Paleogene land bridge acted as a low on this structure and was the point of initiation for the Saalian *breach event*. The impact of glacio-isostasy, its equilibrium with the oscillating ice masses and its subsequent relaxation combined with post-glacial inherited palaeotopography (points of first inundation) determines how fast the landscape inundates during the following interglacial sea-level rise (demonstrated in Chapter 5). On the other hand, at the interglacial-glacial transition it is important to understand the connection of tributary valley systems to major trunk valleys (e.g. palaeo-Scheldt-Axial Channel). Understanding this relationship determines incision depths along a valley system, something not yet well understood for the palaeo-Scheldt Valley. For example why is the upstream Flemish Valley deeper incised (up to 20 m) than the downstream palaeo-Scheldt Valley (only 15 m) as a result of sea level lowering during the Early Glacial and the following Pleniglacial? The concept of retrogressive erosion dictates that the downstream section of the palaeo-Scheldt Valley should be incised deeper than the upstream Flemish Valley due to its proximal connection with the Axial Channel trunk valley. Why do we also not find a connection with the Axial Channel trunk valley? What is the impact of the elevation difference and the substrate? Are we missing something? It would be interesting to study the fluvial processes forming the Axial Channel and palaeo-Scheldt-Flemish Valley independently and perhaps compare the results in a later stage in order to get to a more profound and fundamental understanding of this concept in this area of the southern North Sea.

Finally, the presented research demonstrates that inherited palaeotopography in combination with glacio-isostasy exerts a major control on landscape evolution both during a glaciation and the following interglacial. In other words glaciations and interglacials should be studied as interconnected cycles and not as independent entities of a glacial-interglacial cycle.

9.4.3 A multi-disciplinary study

Within this dissertation different scientific fields (sedimentology, seismic facies analysis, palaeogeography, archaeology, palaeontology, taphonomy, etc.) were combined resulting in a multi-disciplinary study trying to explain local/regional landscape evolution and the spatio-temporal impact of syn- and post-depositional processes on actual preservation of submerged heritage. The presented study was not able to explain everything and not every section received as much attention or detail and includes generalisations that can be improved. For instance, the palaeogeographic map series has improved our understanding of the local landscape evolution and its impact on a more regional scale, however, as mentioned in Section 9.4.2, it would be better to divide the study of the Belgian shelf and its link to regional landscape elements into separate valley systems/areas from now on. This way process interactions can be studied more extensively and at a later stage these processes can maybe be better understood or connected between these different valley systems/areas resulting in new and improved insights.

9.5 Outlook and where next?

In order to determine a more precise stratigraphy of the Belgian shelf, geochronological control and additional palaeoenvironmental information should be obtained from the sedimentary record. Combining the sedimentary information with seismic reflection analysis and age information will prove invaluable as hiatuses within the sediment record might be resolved and a better-fixed chronological framework can be developed. The sediment cores needed to get this information are already partially available and are stored at a repository of RBINS in Brussels. However, additional sediment cores from areas with low data density have to be obtained to be able to reconstruct a reliable stratigraphy from incised-valley systems or underneath sandbanks. A good example is the Scheur find location where sediment cores/vibrocores need to be taken to get a better understanding of the palaeoenvironment and its stratigraphy.

A next step would be to better connect the stratigraphic record with that of the Netherlands, France and the United Kingdom. This dissertation demonstrated that significant inconsistencies remain present within the stratigraphic record between the neighbouring territories. For example, the Rhine-Meuse River units B2–B6 by Hijma et al. (2012: their Figure 17) are possibly incorrect and should be reconsidered. If we want to understand the landscape evolution of the North Sea as a whole, more and better international research needs to take place. A first step in the right direction was recently taken through a collaboration agreement between Ghent University (Belgium), Bradford University (United Kingdom) and TNO and Utrecht University (the Netherlands). Within this collaboration, researchers want to better understand the landscape evolution of the southern North Sea from the Late Pleistocene to the Holocene period and what the impact of the changing landscape is on human migration patterns such as the occupation of Doggerland. This is a first step towards a long-term multidisciplinary research in the southern North Sea.

It would also be interesting to study the onshore connection from the Waardamme and Lys Rivers with the palaeo-Scheldt River. In this study we assumed, partly based on literature, the latter captured these rivers during the Late Saalian as a result of glacio-isostatic upwarping, but no true geophysical evidence was presented. Additional information such as geomorphology and sedimentology may provide important clues to this phenomenon and how glacio-eustatic processes are stored within their sedimentary archives.

Finally, a better stratigraphic connection is needed between the Belgian shelf and Flanders. Many local geological investigations were performed over the past decades in the coastal area. However, most of these local investigations are hard to correlate between one another. This was demonstrated in Chapter 5 with the study of the palaeo-Scheldt Valley system. Offshore the stratigraphic record is difficult to untangle with seismic information alone due to presence of shallow gas in the subsurface. It is therefore important to increase the amount of sediment cores/vibrocores in this region to correlate the palaeoenvironments based on their sedimentary characteristics. At the same time a better connection can be provided between the onshore and offshore Paleogene units. The Complex Project Kustvisie (www.kustvisie.be) is a first step to overcome this problem. Within this project, cone penetration tests, discontinuous and continuous cores that penetrate the complete Quaternary sequence and to a certain extent also the Paleogene, will be performed. The information from this project will provide additional stratigraphic information for the area between Zeebrugge and the Belgian-Dutch border and how the Zeebrugge and Waardamme Valleys of the palaeo-Scheldt Valley system developed. Because the Scheur and the Zeebrugge navigation channels are located within this study area the information from these cone penetration test and sediment cores will provide a better and more detailed assessment of the preservation potential of the stratigraphic record in this area.

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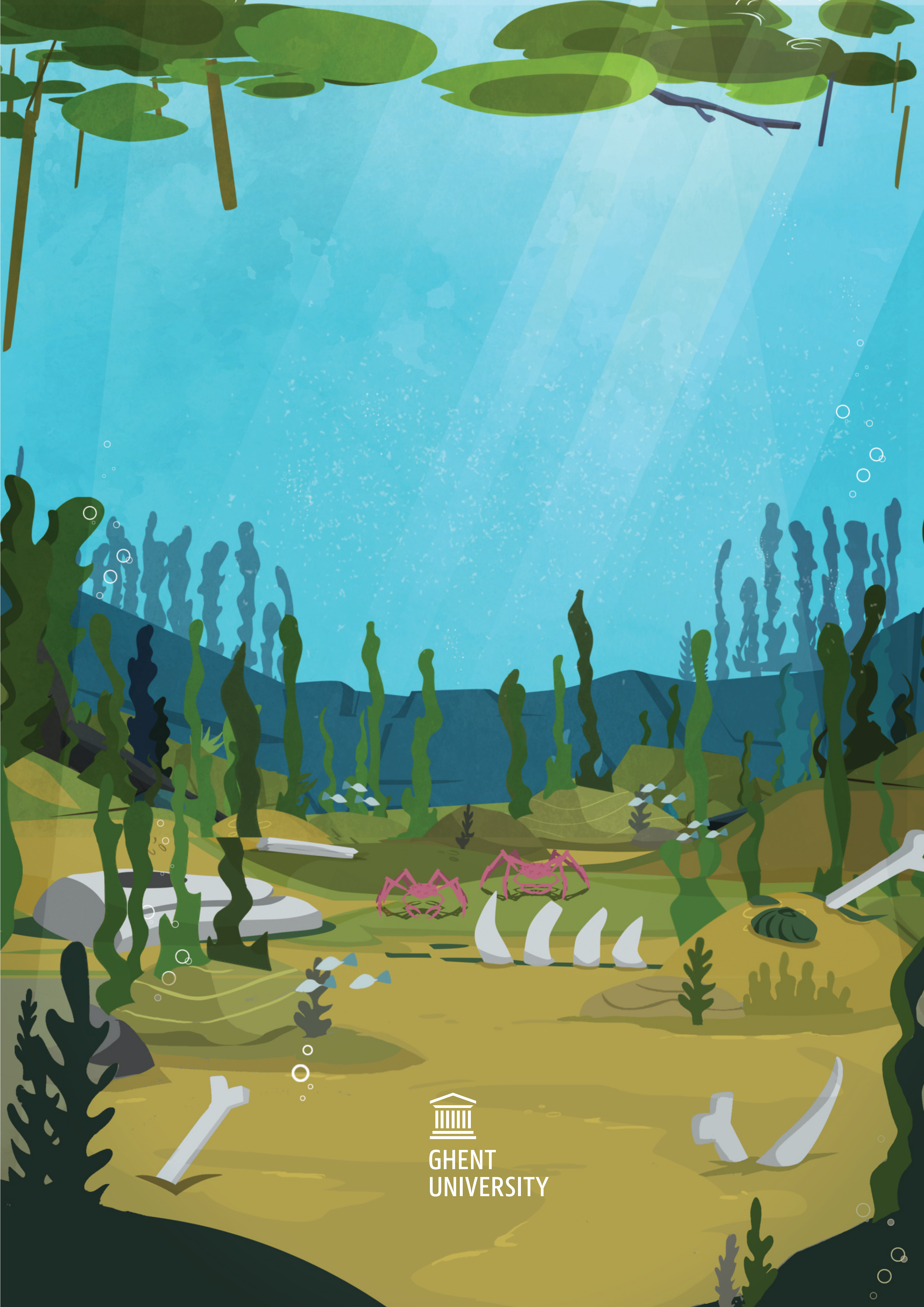
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