

EVOLUTION OF THE GOLLUM CHANNEL SYSTEM, OFFSHORE IRELAND

*Interplay between downslope and alongslope marine
processes and their effects on regional sedimentation*

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Academic year 2024-2025

A dissertation submitted to the Faculty of Sciences of Ghent University
in fulfilment of the requirements for the degree of Doctor of Science: Geology

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This research was carried out by Lotte Verweirder with financial support of Research Foundation – Flanders (FWO grant 1114521N) and the Special Research Fund (BOF) of Ghent University. The research project was conducted at the Renard Centre of Marine Geology, Ghent University (Belgium).

Please refer to this dissertation as: *Verweirder, L., 2025. Evolution of the Gollum Channel System, offshore Ireland. Interplay between downslope and alongslope marine processes and their effects on regional sedimentation. PhD thesis, Ghent University, Ghent, Belgium.*

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Cover images: Bathymetry of the Gollum Channel System and Kings Channel System in the Porcupine Seabight, offshore Ireland. The dataset is the GEBCO_2024 grid (GEBCO Compilation Group, 2024).

Chapter covers: artwork by Nelle Verweirder

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List of abbreviations

AMS	accelerator mass spectrometry
AUV	autonomous underwater vehicle
BA	Bølling-Allerød
BIIS	British-Irish Ice Sheet
BMP	Belgica Mound Province
BP	before Present
CSS	Celtic Sea Shelf
CT	computed tomography
CTD	conductivity, temperature, depth
CWC	cold-water coral
ENACW	Eastern North Atlantic Central Water
ESC	European Slope Current
GCS	Gollum Channel System
IODP Exp. 307	International Ocean Drilling Program Expedition 307
IRD	ice-rafted debris
ka	thousand years ago
KCS	Kings Channel System
kyr	thousand years
LGM	Last Glacial Maximum
LSW	Labrador Sea Water
Ma	million years ago
MBES	multibeam echosounder
MIS	Marine Isotope Stage
MOW	Mediterranean Outflow Water
MSCL	multi-sensor core logger
Myr	million years
NEADW	North East Atlantic Deep Water
PAP	Porcupine Abyssal Plain
PS	Porcupine Seabight
$\overline{SS}$	mean sortable silt
SS%	percentage sortable silt
SSS	side-scan sonar
TWT	two-way travel time
XRF	X-ray fluorescence
YD	Younger Dryas

Acknowledgements

One of the things that I enjoyed most about my PhD experience was having the opportunity to meet people from all over the world at the department, in other research institutes, at conferences, during courses, or during surveys. You all have given me new ideas from your different perspectives on research and on life, and courage to keep going.

First and foremost, I would like to thank the people without whom this PhD would literally and logistically not have been possible: David and Aggie, thank you so much for the opportunity to work in your beloved Gollum channels and for supporting me in pursuing my curiosity. Thank you for the time you put into this project, and ultimately, my development as a researcher.

Another big thank you goes to the crews of RV Celtic Explorer and RV Belgica for their help with data acquisition and sharing their wisdom about the different ships' systems; to the people at VLIZ, who taught me a large chunk of what I know about geophysical data processing and allowed me to create clean, good-looking datasets to be proud of; to the teams in Innsbruck, Southampton, and Coleraine, who helped with the sediment side of things; to Martin and Eoghan for your very useful input about the oceanography, which I didn't know anything about at the start of this PhD and now feel like I want to delve into a lot more; to the DEHEAT and Sanimed teams for allowing me to come with them to see how other people organize their surveys and to put some of the skills that I learned into practise; and to the INCISE community – a whole world opened up the first time I joined a conference – many thanks for the bursts of motivation and inspiration, and the very enjoyable atmosphere you managed to create.

I would also like to say thanks to the people at RCMG and the Geology department for the good times inside but also outside of the office, for help in the labs, a quick chat in the corridor or a drink after work, a collective rant about everything that was throwing off our planning and all the data that were impossible to figure out, or a sunny lunch on the benches at the back of the S8. Alice and Piet, I am very grateful to have had you as office mates – you made me feel at ease there. You were there when I was stuck in my head and wanted to discuss things, but also when I needed a laugh, or just to talk about the weekend. Both the tough times and the smaller or bigger wins – I feel like we experienced many of them together.

I can also not forget to thank my friends and family for reminding me there is a life outside of a PhD (especially during the last year when I felt like there was very little else going on). Thanks for meeting up for drinks, dinners, runs, movies, Gentse Feesten afternoons/nights/mornings, New Year's Eve parties, other parties, summer trips, board game nights, afterwork ping pong sessions, or for having me over at your place in Norway.

And lastly, mama, papa, Annouk and Nelle: thank you for all your support; you were there for me even when I was a bit grumpy when things were not going as smoothly as I had had in mind, or when I was not as present because I was going on cruises and to conferences. Mama en papa, bedankt om mij te laten opblijven om naar alle natuurdocumentaires over moessonregens en de Grote Slenk te laten kijken en mijn nieuwsgierigheid te laten groeien. Ongetwijfeld een goeie opvoedingskeuze; wie weet of ik hier geraakt was anders. Nellson, de vele wandelingetjes, het samen eten en de cosy zetelmomentjes voor toch wel een aantal seizoenen van potten en taarten bakken hebben mij enorm deugd gedaan.

Dankjulliewel.

Summary

The importance of the Earth's ocean for the global climate, evolution of life, and human society is increasingly recognised but large parts of it remain unexplored. Many of the ocean's functions are dependent on the input of particles such as nutrients, organic material, sediments, and pollutants, from the continents. For the transport of these particles from the continents and shallow seas into the deep ocean – typically through downslope gravity flows – submarine canyon-channel systems cutting through continental margins at the edges of oceanic basins provide the most efficient pathways. Besides the downslope activity that these systems host, their steep seabed topography can create locally enhanced bottom currents through interaction with oceanographic processes such as internal waves and alongslope currents. Their function as particle conduits along with the dynamic oceanographic environment they host makes of canyon-channel systems important biodiversity hotspots, a major part of the global carbon cycle, a necessary component in many geohazard risk assessments, and an archive of past marine and terrestrial processes that can provide valuable insight into, e.g., climatic variability.

Canyon-channel systems with a direct connection to the continent (e.g., at the mouth of a river) are generally considered more frequently and more intensely active with downslope flows than their land-detached counterparts. About 80 % of all canyon-channel systems are separated from direct terrestrial input by a continental shelf. They can still receive material through processes such as storms and earthquakes that have the capability of resuspending shelf sediments and transporting them further towards the shelf edge and into a land-detached system. Recent monitoring has showed that in some land-detached systems, these shelf processes enable a frequency and intensity of downslope activity similar to that in land-attached systems. Thus, the idea of inactivity in land-detached systems is being challenged more and more and the question is raised whether frequent downslope activity should be expected in land-detached systems globally or rather distinct environmental conditions allow or trigger regular activity in specific locations only.

A second major contributor to the dispersal of particles in ocean basins is the alongslope movement of water masses as part of the global system of thermohaline circulation. Alongslope currents are most intense at continental slopes, where they contribute to slope development through erosion and deposition, and commonly interact with downslope flows. Because of the long geological periods of time over which these processes occur, direct observation of the full processes and all their parameters is impossible. So, to understand them, their interaction, and their role in global systems such as climate, it is imperative to reconstruct their past variability. Since the effects of marine processes are often focused on the continental slope, their record can be studied in slope sediments. In this context, research on continental margins is of great worth for unravelling and understanding both terrestrial and marine processes, as well as the interplay between them.

The main aims of this work are to gain a better understanding of (1) the influence of the interplay between alongslope currents and downslope flows on the development of continental slopes, and specifically canyon-channel systems, (2) the downslope activity in land-detached canyon-channel systems, the processes that enable it, and the consequences for deep-sea habitats, and (3) the mechanisms that control the formation and evolution of canyon-channel systems on formerly glaciated margins. This investigation involves analysis of multi-channel seismic data, single-channel seismic data, well data, moored current meter data, hull-mounted multibeam

echosounder data, AUV-mounted side-scan sonar data, AUV-mounted photography data, and sediment cores.

This study focuses on the large-scale Gollum Channel System and the smaller Kings Channel System just north of it, two canyon-channel systems situated in the Porcupine Seabight basin offshore southwest Ireland (Northeast Atlantic continental margin). The Gollum Channel System connects the Celtic Sea Shelf to the Porcupine Abyssal Plain, while the Kings Channel System ends at the base of the slope and does not reach the abyssal plain. Both systems are presently disconnected from any terrestrial sediment sources. They are thought to have been active with downslope flows during Quaternary cold periods of sea-level lowstand, when an ice sheet covered part of the Celtic Sea Shelf and brought material from the continent close to the channel heads. It is, however, unknown when both systems were formed, which processes were involved in their development, and which processes are active in the systems today. Additionally, both sedimentary processes and associated benthic ecosystems on the continental slope of the Porcupine Seabight are known to be influenced by a northwards alongslope current that brings warm and saline Mediterranean Outflow Water into the basin during Quaternary warm periods. At the top of this water mass, the density difference with the overlying Eastern North Atlantic Central Water leads to the formation of internal waves and enhanced bottom currents over the upper slope. Thus, both along- and downslope impact on the slopes of this basin are thought to have varied through time and appear linked to climate variability. Here, this interaction will be explored both in the context of the development of the channel systems, as well as from a palaeoceanographic point of view.

In the first research chapter (Chapter 2), the sedimentary environment of the study area is characterised from the late Miocene to the present day through analysis of bathymetry data, sparker single-channel seismic data, and current meter data. The results indicate that a regional erosional event (represented by regional discontinuity RD1 in seismic data) forms the base of the Kings Channel System. This event was previously attributed to intense bottom current action from the introduction of Mediterranean Outflow Water into the basin in the Pliocene, and is here shown to be caused by the combined effect of intensified bottom currents and downslope flows that created the bases of the Kings Channel System. Although the RD1 horizon is here also proposed to form the base of the Gollum Channel System, the vertical extent of the seismic data is not sufficient to investigate the deepest incisions of this system. The erosive downslope flows are thought to originate from the Celtic Sea Shelf, where lowered sea levels in Quaternary cold periods generated proximal sediment sources. The Kings and Gollum channel systems exhibit morphological differences, which may be explained by different sediment sources (i.e., Ireland and the Celtic Sea Shelf, respectively) that may have brought gravity flows of different magnitudes to the two channel systems.

The second research chapter (Chapter 3) further investigates which processes were involved in the formation and evolution of the Gollum Channel System. Airgun multi-channel seismic data correlated to well and borehole data are analysed to identify the large-scale subsurface structures underneath the channel system and build a chronostratigraphic framework. A period of deposition in a low-energy environment from the Early Cretaceous to the early Eocene was terminated by large-scale erosion from downslope flows in the late Eocene. This was likely linked to differential subsidence and created a first system of channels draining the Celtic Sea Shelf. The late Eocene to Miocene succession contains evidence for bottom current influence on deposition. The bases of the oldest Gollum channels were incised sometime between the early middle Miocene and the late Pliocene, making them older than the early Quaternary age they were

thought to have. From the timing of incision, the formation of the channels is tentatively attributed to a phase of early Pliocene margin tilting. Quaternary gravity flows reused the channels and mostly bypassed the upper slope or contributed to the channel fill, though some flows were erosional and capable of incising several smaller channels. Symmetry between northern and southern levees in the Pliocene evolved into asymmetry in the early Quaternary that gradually disappears towards the present-day seafloor. This suggests interaction between downslope flows and alongslope currents in the Quaternary, diminishing as the channels become deeper and gravity flows completely contained within them.

Chapter 4 explores the downslope activity in the Gollum Channel System from the Last Glacial Maximum until today. Centimetre-resolution side-scan sonar, bathymetry, and photography data gathered with an autonomous underwater vehicle are analysed alongside two sediment gravity cores and the current meter data that were presented in Chapter 2. Turbidites in the sediment cores indicate that gravity flows were active in the channels during the last glacial period and until c. 9000 years ago. The present-day environment is dominated by bottom currents that are topographically steered along the channel axes and have a tidal frequency. They create megaripples and provide a favourable environment for the development of cold-water coral mini-mounds. At present, downslope activity is limited to a gully in the channel head of one of the channels, indicated by the presence of crescentic bedforms and scarps there. The limited spatial extent of these bedforms suggests present-day flows are minimal in volume and frequency. Hence, the Gollum Channel System can be considered inactive but hosting a dynamic sedimentary environment.

In the last research chapter (Chapter 5), a record of palaeoceanography since the last deglaciation is built from five sediment gravity cores on the open slopes next to the channels. The combination of four mid-slope cores (water depth 1200-1300 m) and one upper-slope core (water depth 700 m) allows bottom-current reconstruction at the core of the Mediterranean Outflow Water flow as well as its upper boundary. A low-energy environment with hemipelagic deposition was present across the slope at the end of the last deglaciation. It was followed by intensified bottom currents producing silty contourites at the middle slope in the Bølling-Allerød and Younger Dryas. Deposition on the upper slope remained hemipelagic for longer and gradually transitioned to muddy and then silty contourites by the Younger Dryas. This is interpreted as the introduction of the Mediterranean Outflow Water in the basin before the Bølling-Allerød, and a subsequent shoaling of the water mass until it reached its present-day water depth in the early Holocene. From then on, deposition is stable with muddy contourites on the middle slope and silty contourites on the upper slope, where the density difference between the Mediterranean Outflow Water and the overlying Eastern North Atlantic Central Water produces internal waves and enhanced bottom currents.

The interaction between downslope and alongslope currents has been essential in the development of the eastern slope of the Porcupine Seabight, and more specifically the Gollum and Kings channel systems. Furthermore, the alongslope current enables the development of a dynamic environment that is crucial for deep-sea habitats within the channels. It is also the main source of sediment to the basin (and along the Irish margin), especially during periods of sea-level highstand, and a large contributor to the construction of the continental slope here.

In the Quaternary, the channels were most active during Late Quaternary cold periods of lowered sea levels when the ice margin on the shelf was close to the channel heads but activity persisted under rising sea levels when ice had retreated. The mechanism of sediment delivery to the channels during low sea levels is difficult to pinpoint due to limited knowledge of processes on

the outer shelf and, thus, a missing link between the ice margin and the channel heads. Other potential mechanisms put forward to be investigated further are slope failures and oceanographic processes such as near-inertial waves.

The Gollum Channel System was formed before the onset of northern hemisphere glaciations in the early Quaternary. Notwithstanding its long history, the deep-sea fan at the foot of the Gollum Channel System is very limited in extent, which raises questions about where all of the material brought in by the gravity flows has gone. Margin tilting is suggested as the cause of initial incision. The contrasting slope elements created along the Northeast Atlantic margin from this phase of tilting, however, indicates that other factors (slope angle, amount and type of sediment, process of delivery,...?) are in play to control where canyon-channel systems form. The formation of the Gollum Channel System before the Quaternary also suggests that glaciomarine processes are not the main control on the formation of large-scale canyon-channel systems on this margin. Rather, in the case of the Gollum Channel System, they reused existing pathways for downslope transport of the material they brought to the shelf edge.

Samenvatting

Het belang van de oceaan voor het wereldwijde klimaat, de evolutie van het leven en de maatschappij wordt steeds meer ingezien, maar grote delen ervan zijn nog niet verkend. Veel oceanische functies zijn afhankelijk van de aanvoer van deeltjes zoals voedingsstoffen, organisch materiaal, sedimenten en verontreinigende stoffen uit de continenten. De meest efficiënte route voor het transport van deze deeltjes van continenten en ondiepe zeeën naar de diepe oceaan – meestal hellingafwaarts via stromingen aangedreven door zwaartekracht – is doorheen onderzeese canyon-kanaalsystemen. Deze systemen doorsnijden de continentale randen langsheen oceanische bekkens. Naast de hellingafwaartse activiteit in canyon-kanaalsystemen, kunnen oceanografische processen zoals interne golven en hellingparallelle stromingen er lokaal de bodemstromingen versterken. Omdat ze zo'n dynamische oceanografische omgeving zijn en door hun functie in het transport van deeltjes in de diepzee, zijn canyon-kanaalsystemen hotspots voor biodiversiteit, een belangrijk onderdeel in de wereldwijde koolstofcyclus, een noodzakelijk element in risicobeoordelingen van geologische gevaren en een archief van mariene en terrestrische processen dat waardevolle inzichten kan bieden in bijvoorbeeld de evolutie van klimaatverandering.

Hellingafwaartse activiteit in canyon-kanaalsystemen die direct verbonden zijn met het continent (bijvoorbeeld aan een riviermonding) wordt over het algemeen beschouwd als frequenter en intenser dan die in systemen die gescheiden zijn van het continent. Ongeveer 80% van alle canyon-kanaalsystemen is gescheiden van directe input van het land door een continentaal plat. Ze kunnen nog steeds materiaal ontvangen via processen zoals stormen en aardbevingen die sedimenten op het continentaal plat in suspensie te brengen, die verder naar de continentale rand transporteren en uiteindelijk naar zo'n canyon-kanaalsysteem afgescheiden van land. Recente monitoring heeft aangetoond dat deze processen op het continentaal plat in sommige systemen een frequentie en intensiteit van activiteit mogelijk maken die vergelijkbaar is met die in systemen met een directe continentale verbinding. Het idee van inactiviteit in systemen afgescheiden van land wordt dus steeds meer in twijfel getrokken en de vraag kan gesteld worden of frequente hellingafwaartse activiteit te verwachten is in afgescheiden systemen wereldwijd, of dat hiervoor specifieke omgevingsfactoren nodig zijn.

Een tweede belangrijke factor voor de verspreiding van deeltjes in oceaanbekkens is de beweging van watermassa's langsheen de continentale helling. Dit is een deel van het wereldwijde systeem van thermohaliene circulatie, waarin hellingparallelle stromingen het intensst zijn langs continentale hellingen. Hier dragen ze bij aan het vormgeven van de helling door erosie en afzetting van deeltjes, en interageren ze met hellingafwaartse processen. Doordat deze mariene processen plaatsvinden over lange geologische periodes, is directe observatie van de volledige processen en al hun parameters onmogelijk. Om de processen, hun interactie en hun rol in wereldwijde systemen zoals het klimaat te begrijpen, is het daarom noodzakelijk om hun variabiliteit doorheen het verleden te reconstrueren. Omdat de effecten van mariene processen zich vaak concentreren op de continentale helling, kunnen ze worden bestudeerd in sedimenten op die helling. In deze context is onderzoek aan continentale randen van grote waarde voor het ontrafelen en begrijpen van zowel continentale als mariene processen, evenals de interactie ertussen.

De belangrijkste doelen van deze studie zijn om een beter inzicht te krijgen in (1) de invloed van de wisselwerking tussen hellingparallelle stromingen en hellingafwaartse stromingen op de opbouw van continentale hellingen, en specifiek de evolutie van canyon-kanaalsystemen, (2) de hellingafwaartse activiteit in canyon-kanaalsystemen zonder directe verbinding met een continent, de processen die dit mogelijk maken en de gevolgen voor diepzeehabitats, en (3) de

processen die betrokken zijn bij het ontstaan en de evolutie van canyon-kanaalsystemen langs continentale randen die beïnvloed zijn door Quartaire ijskappen. Dit onderzoek is gebaseerd op de analyse van meerkanaals-seismische data, enkelkanaals-seismische data, boorputdata, data van verankerde stroomsnelheidsmeters, bathymetriedata van *multibeam echosounders*, side scan sonardata en foto's opgenomen met een *autonomous underwater vehicle*, en sedimentkernen.

Deze studie focust op het grote Gollum-kanalensysteem en het kleinere Kings-kanalensysteem net ten noorden ervan, twee canyon-kanaalsystemen in het Porcupine Seabight-bekken ten zuidwesten van Ierland (continentale rand van de noordoostelijke Atlantische Oceaan). Het Gollum-kanalensysteem verbindt de Celtic Sea Shelf met de Porcupine Abyssal Plain, terwijl het Kings-kanalensysteem eindigt aan de voet van de continentale helling en de abyssale vlakte niet bereikt. Beide systemen zijn momenteel niet verbonden met continentale sedimentinput. Er wordt vermoed dat ze actief waren tijdens koude periodes van lage zeespiegelstand in het Quartair, toen een ijskap een deel van de Celtic Sea Shelf bedekte en materiaal van het continent dichterbij de kanalen bracht. Het is echter onbekend wanneer beide systemen zijn ontstaan, welke processen betrokken waren bij hun ontwikkeling en welke processen er vandaag de dag actief zijn. Bovendien worden zowel sedimentaire processen als benthische ecosystemen op de continentale helling van de Porcupine Seabight beïnvloed door een hellingparallelle noordwaartse stroming die warm en zout Mediterranean Outflow Water binnenbrengt in het bekken tijdens Quartaire warme perioden. Aan de bovenkant van deze watermassa leidt het dichtheidsverschil met het Eastern North Atlantic Central Water tot de vorming van interne golven en versterkte bodemstromingen over de continentale helling. Het lijkt er dus op dat zowel de hellingparallelle als de hellingafwaartse processen langs de randen van dit bekken variëren doorheen de tijd en verband houden met variabiliteit in het klimaat. Deze interactie zal hier worden onderzocht, zowel in de context van de ontwikkeling van de kanaalsystemen als vanuit een paleoceanografisch perspectief.

In het eerste onderzoekshoofdstuk (Hoofdstuk 2) worden de sedimentaire processen in het studiegebied beschreven van het Laat-Mioceen tot nu via de analyse van bathymetriedata, *sparker* seismische data (enkelkanaals) en stroomsnelheidsdata. De resultaten tonen dat een erosief oppervlak (aanwezig in seismische data als regionale discontinuïteit RD1) aan de basis ligt van het Kings-kanalensysteem. Deze erosieve fase werd eerder toegeschreven aan intense hellingparallelle bodemstromingen die aanwezig waren wanneer het Mediterranean Outflow Water voor het eerst in het bekken stroomde in het Pliocene. Hier wordt aangetoond dat de RD1 erosie veroorzaakt werd door een combinatie van deze intense bodemstromingen en de erosieve hellingafwaartse stromingen die de basis van het Kings-kanalensysteem gevormd hebben. Hoewel hier wordt geopperd dat de RD1 horizon ook de basis van het Gollum-kanalensysteem is, is het verticale bereik waarover de seismische data reikt niet voldoende om de diepste insnijdingen van dit systeem te onderzoeken. De erosieve hellingafwaartse stromingen worden verondersteld afkomstig te zijn van het Ierse continentale plat, waar lage zeespiegels in koude perioden van het Quartair zorgden voor een kortere afstand tussen het continent en de kanalen. De Kings- en Gollum-kanalensystemen zijn morfologisch verschillend, wat mogelijk verklaard kan worden door ongelijke sedimentbronnen (respectievelijk Ierland en de Celtic Sea Shelf) die mogelijk hellingafwaartse stromingen van verschillende omvang naar de twee systemen hebben gebracht.

Het tweede onderzoekshoofdstuk (Hoofdstuk 3) bestudeert verder welke processen betrokken waren bij de vorming en evolutie van het Gollum-kanalensysteem. *Airgun* seismische data (meerkanaals), gecorreleerd met put- en boorgatgegevens, worden geanalyseerd om de grootschalige structuren onder het systeem te identificeren en een chronostratigrafisch kader te construeren. Een periode van afzetting in een laagenergetische omgeving van het Vroeg-Krijt tot het Vroeg-Eoceen werd beëindigd door grootschalige erosie door hellingafwaartse stromingen in

het laat-Eoceen. Dit kwam waarschijnlijk door differentiële subsidentie en creëerde een eerste systeem van kanalen dat de Celtic Sea Shelf draineerde. Van het laat-Eoceen tot het Mioceen vond afzetting plaats beïnvloed door bodemstromingen. De basissen van de oudste Gollum kanalen werden ergens tussen het Vroeg-Mioceen en het Laat-Pliocene ingesneden, wat aantoont dat ze ouder zijn dan hun vooraf veronderstelde vroeg-Quartaire ouderdom. Op basis van het tijdstip van insnijding wordt de vorming van de kanalen voorlopig toegeschreven aan een fase van *tilt* van de continentale rand (opheffing van het continentaal plat en het continent en simultane subsidentie in het bekken) in het Vroeg-Pliocene. De hellingafwaartse stromingen uit het Quartair hergebruikten de kanalen, waarbij sommige stromingen bijdroegen aan de opvulling van de kanalen, terwijl andere in staat waren tot erosie en het insnijden van nieuwe kanalen. De noordelijke en zuidelijke kanaalflanken (*levees*) waren symmetrisch in het Pliocene en werden asymmetrisch in het vroege Quartair. Deze asymmetrie verdween geleidelijk richting het heden. Dit wijst op interactie tussen hellingafwaartse en hellingparallelle stromingen in het Quartair, die afnam naarmate de kanalen dieper werden en de hellingafwaartse stromingen er volledig in pasten.

Hoofdstuk 4 onderzoekt de hellingafwaartse activiteit in het Gollum-kanalensysteem vanaf het Laatste Glaciale Maximum tot nu. Side scan sonar data, bathymetriedata en foto's verzameld met een *autonomous underwater vehicle* worden geanalyseerd, samen met twee sedimentkernen en de stroomsnelheidsdata die in hoofdstuk 2 werden besproken. Turbidieten in de sedimentkernen tonen aan dat hellingafwaartse stromingen actief waren in de kanalen tijdens de laatste glaciële periode en tot ongeveer 9000 jaar geleden. Momenteel zijn bodemstromingen het belangrijkste proces op de kanaalbodems. Ze stromen afwisselend hellingopwaarts en hellingafwaarts met een frequentie gestuurd door getijden. Ze creëren megaribbels en bieden een gunstige omgeving voor de ontwikkeling van *mini-mounds* van koudwaterkoralen. Momenteel is de hellingafwaartse activiteit beperkt tot een geul in het ondiepste deel van een van de kanalen, wat blijkt uit de aanwezigheid van sikkelvormige structuren daar. Het voorkomen van deze structuren is heel lokaal, wat suggereert dat de hedendaagse stromingen een minimaal volume en frequentie hebben. Het Gollum-kanalensysteem wordt daarom als inactief beschouwd, maar herbergt een dynamische sedimentaire omgeving.

In het laatste onderzoekshoofdstuk (hoofdstuk 5) wordt een archief van de paleoceanografie sinds de laatste deglaciatie opgebouwd aan de hand van vijf sedimentkernen op de continentale helling naast de kanalen. De combinatie van vier kernen in middelmatige dieptes op de helling (waterdiepte 1200-1300 m) en een kern in het ondiepe deel van de helling (waterdiepte 700 m) maakt een reconstructie van de bodemstromingen in de kern van het Mediterranean Outflow Water en aan de bovengrens ervan mogelijk. Aan het einde van de laatste deglaciatie was er een laagenergetisch milieu met hemipelagische afzetting over de volledige helling. Dit werd gevolgd door intensere bodemstromingen die siltige contourieten produceerden op de middenhelling tijdens het Bølling-Allerød en het Jonge Dryas. De afzetting op de bovenhelling bleef langer hemipelagisch en ging geleidelijk over in kleiige en vervolgens siltige contourieten omstreeks het Jonge Dryas. Dit wordt geïnterpreteerd als de introductie van het Mediterranean Outflow Water in het bekken voor het Bølling-Allerød, die daarna verticaal verplaatste naar kleinere waterdieptes totdat ze in het vroege Holoceen haar huidige waterdiepte bereikte. Vanaf dat moment was sedimentatie stabiel met kleiige contourieten op de middenhelling en siltige contourieten op de bovenhelling, waar het dichtheidsverschil tussen het Mediterranean Outflow Water en het bovenliggende Eastern North Atlantic Central Water interne golven en versterkte bodemstromingen veroorzaakt.

De interactie tussen hellingafwaartse en hellingparallelle stromingen is essentieel geweest voor de opbouw van de oostelijke helling van het Porcupine Seabight-bekken en de Gollum- en Kingskanalensystemen in het bijzonder. Bovendien creëert de hellingparallelle stroming een dynamische omgeving die cruciaal is voor de ontwikkeling van diepzeehabitats in de kanalen.

Deze stroming is ook de voornaamste bron van sediment in het bekken (en langs de Ierse continentale rand) – vooral tijdens perioden van hoge zeespiegel – en levert zo een belangrijke bijdrage aan de evolutie van de continentale rand.

In het Quartair waren de kanalen het meest actief tijdens koude perioden in het laat-Quartair wanneer het zeeniveau laag was en de rand van de ijskap dichtbij de kanalen, maar activiteit hield aan tijdens perioden van stijgende zeespiegel wanneer de ijskap verdwenen was van het continentaal plat. Het mechanisme van sedimentinput naar de kanalen bij lage zeespiegels is moeilijk te bepalen door een beperkte kennis van de processen op het buitenste gedeelte van het continentaal plat, waardoor er een *missing link* is tussen de rand van de ijskap en de kanalen. Andere potentiële mechanismen die hier naar voren geschoven worden voor verder onderzoek zijn massabewegingen door afglijdingen aan de flanken van de kanalen en oceanografische processen zoals *near-inertial waves*.

Het Gollum-kanalensysteem werd gevormd voor het aanbreken van de ijstijden op het noordelijk halfrond in het vroege Quartair. Ondanks zijn lange geschiedenis is de *deep-sea fan* aan de voet van het systeem zeer beperkt in omvang, wat vragen oproept over waar al het materiaal dat aangevoerd werd door hellingafwaartse stromingen, heen is. *Tilt* van de continentale rand wordt geopperd als de oorzaak voor de initiële insnijding van de kanalen. De verschillende soorten afzettingen die door deze fase van *tilt* gecreëerd werden langsheen de noordoostelijke Atlantische rand suggereren dat ook andere factoren (hellingshoek, hoeveelheid en soort sediment, proces dat zorgt voor de input,...?) dan het onderliggende mechanisme belangrijk zijn in het bepalen waar canyon-kanaalsystemen vormen. De vorming van het Gollum-kanalensysteem voor het begin van het Quartair duidt er ook op dat glaciomariene processen niet de belangrijkste component zijn in het ontstaan van canyon-kanaalsystemen langs deze continentale rand. In het geval van het Gollum-kanalensysteem hebben deze processen bestaande paden hergebruikt voor het hellingafwaartse transport van het materiaal dat ze naar de continentale rand brachten.

*“There’s truth to the old cliché
that a picture is worth a thousand words
and that seeing is believing.”*

Marie Tharp

CHAPTER 1 Introduction

1.1 Canyon-channel systems as connections between land and ocean

Since the ocean was formed on Earth 4.4-3.8 Ga (Cavosie et al., 2005; Nutman, 2006; Valley, 2006), it has shaped its surface, regulated its atmosphere, and allowed the evolution of life (NOAA, 2020). Today, human society is dependent on the marine environment for food, trade, and resources. The importance of the ocean is increasingly recognised but large parts remain unexplored (Becker et al., 2009; Harris et al., 2014; Rebesco et al., 2014), which leaves much of the complexity of oceanic processes, their interactions, and their influence on global systems such as climate as only partly understood (Toucanne et al., 2021). Efforts are made to address these knowledge gaps, which is highlighted by international initiatives such as UNESCO's Ocean Decade (UNESCO, 2021).

Many of the ocean's functions are dependent on the input of particles from the continent. For example, a limiting factor for primary productivity in several marine environments is the availability of phosphorus, which is part of the structural backbone for DNA and RNA molecules and therefore essential for all of Earth's life forms (Paytan and McLaughlin, 2007; Syverson et al., 2021). The main source of phosphorus on Earth is the chemical weathering of continents (Syverson et al., 2021), so processes that link the terrestrial to the marine realm are required to deliver this nutrient to the oceanic primary productivity system. This connection is provided at the margins of oceanic basins through a series of processes (Figure 1.1). Nutrients, organic matter, sediments, and pollutants from the hinterland are brought onto the continental shelf through rivers, glacial processes, wind, or coastal erosion (Syvitski et al., 2003). These particles can be resuspended and transported further towards the shelf edge by storms and tsunamis (Puig et al., 2004a; Ikehara et al., 2021), long-shore drift cells (Oliveira et al., 2007), dense shelf-water cascading (Canals et al., 2006; Puig et al., 2013), anthropogenic processes (e.g., trawling; Daly et al., 2018; Martín et al., 2014), or earthquakes (Mountjoy et al., 2014) (Figure 1.1). From the shelf edge, downslope gravity flows carry particles over the continental slope to the deep sea. Here, wind-blown particles (Harrison et al., 2001) or particles resulting from primary productivity at the sea surface add to the deep-sea particle flux by vertical settling through the water column, which may include temporary concentration of particles in nepheloid layers (McCave, 1972; Amin and Huthnance, 1999). Additionally, the alongslope movement of water masses as part of the global system of thermohaline circulation is most intense at continental slopes and provides a mechanism for further dispersal of particles through oceanic basins (Figure 1.1) (Rebesco et al., 2014; Toucanne et al., 2021).

Some large-scale marine processes (e.g., thermohaline circulation, continental slope construction) occur over long geological periods of time, which makes direct observation of the full process and all its parameters impossible. To understand these processes, their interaction, and their role in global systems such as climate, it is imperative to reconstruct their past variability. Since the combination of marine processes results in the deposition or erosion of particles on the continental slopes, their record can be studied in slope sediments (Nittrouer, 1999). Many expeditions have tried to uncover and interpret this archive based on the understanding of the present-day environment (i.e., 'the present is the key to the past') (e.g., Mestdagh et al., 2019; Sierro et al., 2020; de Castro et al., 2021; Toucanne et al., 2021; Normandeau et al., 2024). In this context, research on continental margins is of great worth for unravelling and understanding both terrestrial and marine processes, as well as the interplay between them.

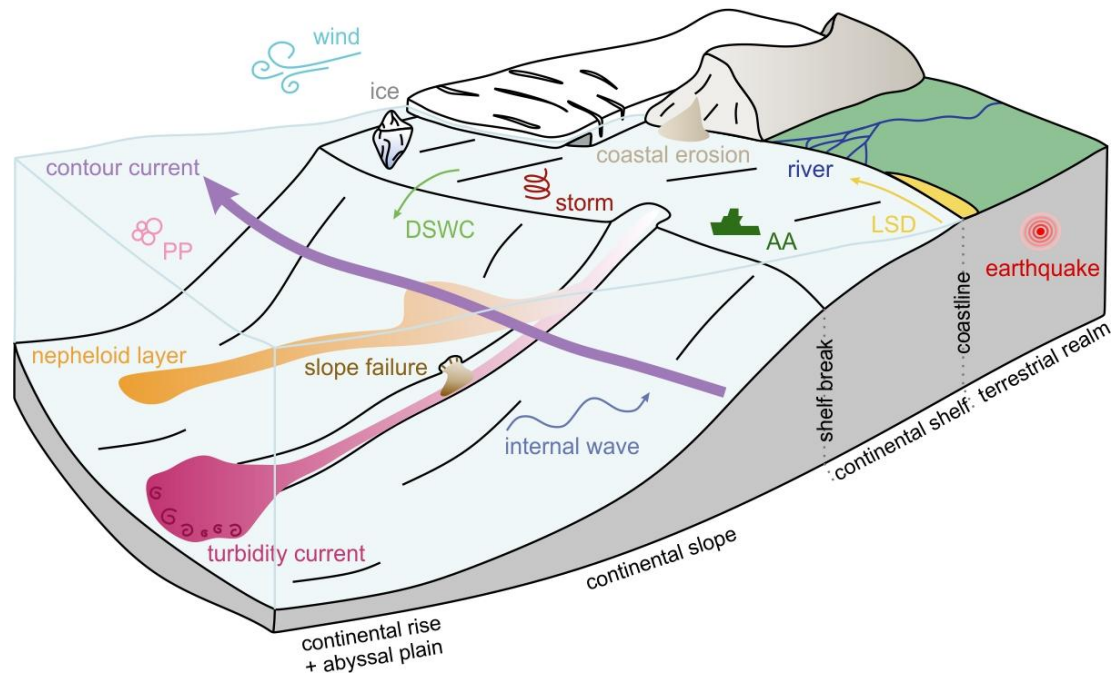


Figure 1.1. Sketch of a hypothetical continental margin showing processes providing (sediment) particle input to the continental slope and transport through submarine canyon-channel systems. AA = anthropogenic activity; DSWC = dense shelf-water cascading; LSD = long-shore drift; PP = primary productivity.

All continental slopes are cross-cut by submarine canyon-channel systems (Figure 1.2) (Harris and Whiteway, 2011). Submarine canyons are steep-walled features governed by erosional processes (Shepard, 1965) that may transform basinwards into submarine channels (e.g., Babonneau et al., 2002; Deptuck et al., 2003) that are characterized by depositional components such as overbank levees, and have lower thalweg slope angles, a gentler topography and a generally more U-shaped cross-section compared to the upstream V-shaped canyons (Shepard, 1965; Ambblas et al., 2018). These canyon-channel systems act as preferential pathways for the downslope gravity flows that carry nutrients, organic matter, sediments, and pollutants from the terrestrial or shallow marine domain downslope (Shepard, 1965a; Canals et al., 2006; Puig et al., 2014; Pierdomenico et al., 2023; Baker et al., 2024). In this function, they provide the most efficient connection between continents and the ocean. Additionally, the interaction of oceanographic processes such as internal tides (Hall et al., 2017) and alongslope currents (Marchès et al., 2007; Voigt et al., 2013) with their steep seabed morphology may lead to locally enhanced bottom currents and/or create nepheloid layers where particulate matter is concentrated (Amin and Huthnance, 1999; Frignani et al., 2002; Wilson et al., 2015; Mencaroni et al., 2021). This combination of particle input and the dynamic oceanographic environment they host can create biodiversity hotspots (De Leo et al., 2010; Amaro et al., 2016; Fernandez-Arcaya et al., 2017), further highlighting their importance. Additionally, submarine canyons play a major role in the global carbon cycle by transferring organic carbon from the terrestrial and shallow marine environment to the deep sea, where it can be buried (Maier et al., 2019; Baker et al., 2024). Furthermore, since they are the location of mass-transport processes and often linked to slope failures from high sedimentation rates and slope undercutting, the study of submarine canyons is a necessary component in many geohazard risk assessments, especially for submarine infrastructure and tsunami risk (Lo Iacono et al., 2011; Carter et al., 2014; Mueller et al., 2016; Power et al., 2016). In 2020-2022, large floods of the Congo River produced strong downslope flows through the submarine Congo Canyon that connects directly to the river estuary (Talling et al., 2022). The flows caused submarine cable breaks, disabling internet connections in parts of West Africa amid the global COVID-19 pandemic (Talling et al., 2022; Bricheno et al., 2024). Lastly, the sedimentary record of downslope flows that they contain is an archive of past marine

and terrestrial processes and can provide valuable insight into, e.g., climatic variability (Toucanne et al., 2008)

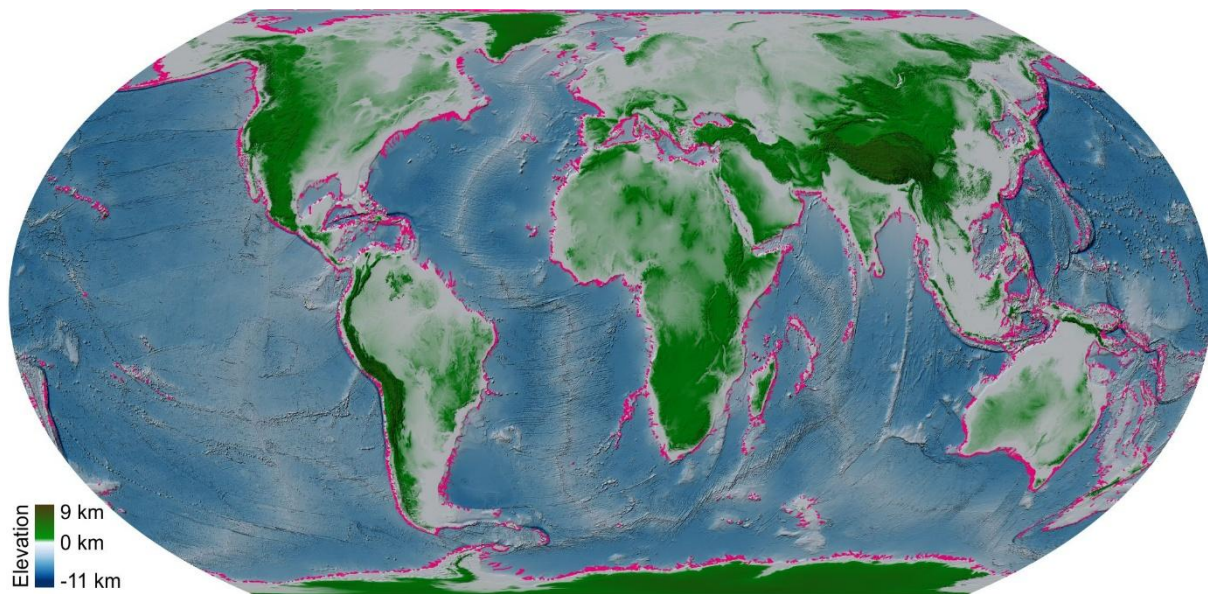


Figure 1.2. Global distribution of large submarine canyons (indicated in pink) from Harris et al. (2014). The background map is the hill-shaded GEBCO_2024 bathymetry grid (GEBCO Compilation Group, 2024) in WGS84/Equal Earth Greenwich projection (epsg:8857).

1.2 Slope sediments as a record of past marine processes

1.2.1 Study of slope sediments

Continental slope processes have traditionally been subdivided into three groups: gravity-driven processes, current-driven processes, and vertical settling. Deposits from these processes appear in the sediment record as mass-transport deposits and turbidites, contourites, and (hemi)pelagites, respectively (Figure 1.3) (Stow and Piper, 1984; Rebesco et al., 2014; Stow and Smillie, 2020). Analysing the characteristics of slope deposits and distinguishing between these three main groups should in theory make it possible to distinguish between dominant processes and interpret the sediment record in terms of process variability. However, each of the three main groups of processes can produce a plethora of sedimentary features as a result of the variability in the process' scale, current speeds, and sediment source (Stow and Smillie, 2020). This makes it hard to distinguish one from the other. Moreover, it is now more and more recognised that these processes interact with each other and create deposits with characteristics that are intermediate between the three main facies (turbidites, contourites, hemipelagites). For example, an alongslope current can deviate the path of a downslope-running turbidity current and thief fine-grained sediment from its suspension cloud (Rodrigues et al., 2022). Additionally, the three facies and their intermediates are often found intercalated (e.g., de Castro et al., 2021; Mencaroni et al., 2021; Miramontes et al., 2021). Hence, the three main groups of processes are understood to represent end-members of a continuum (Figure 1.3) (Stow and Smillie, 2020). To ensure robust interpretation of slope sediments and their associated processes, it is common to combine several methods that allow establishing the large-scale context of the deposits as well as investigating small-scale sediment features.

A common approach in continental slope research (and marine research in general) is to survey a large area at low resolution first, after which smaller areas of interest can be defined for study at higher resolution, from which findings are then extrapolated to the larger area. This helps overcome the temporal and spatial variability with which marine processes occur, e.g., from instantaneous cable breaks caused by turbidity flows through a single canyon (Carter et al., 2014), to the seasonality of plankton blooms in regions of upwelling (Joint et al., 2001), to the

hundreds to thousands of years it may take for a parcel of water to move through the global cycle of thermohaline circulation (Ballarotta et al., 2014). Additionally, this method allows investigation of both the general context and the small-scale properties of a certain process, which enables a more robust interpretation and argumentation (Stow and Smillie, 2020). Traditionally, the analysis of oceanic and sedimentary processes is based on geophysical methods and sediment coring (Clare et al., 2017). These methods are often combined to take advantage of the variety in temporal scales and resolutions they allow to study. Continuous advances in technology and especially in marine robotics allow increasingly efficient (unmanned) investigation, enabling more detailed analysis across spatial scales (Kirkwood, 2007; Xu, 2011; Fabri et al., 2022). In submarine canyon-channel systems in particular, robotics have helped overcome difficulties arising from the rugged terrain that characterises them (Huvenne et al., 2011; Fernandez-Arcaya et al., 2017; Hall et al., 2017; Robert et al., 2017; Normandeau et al., 2024).

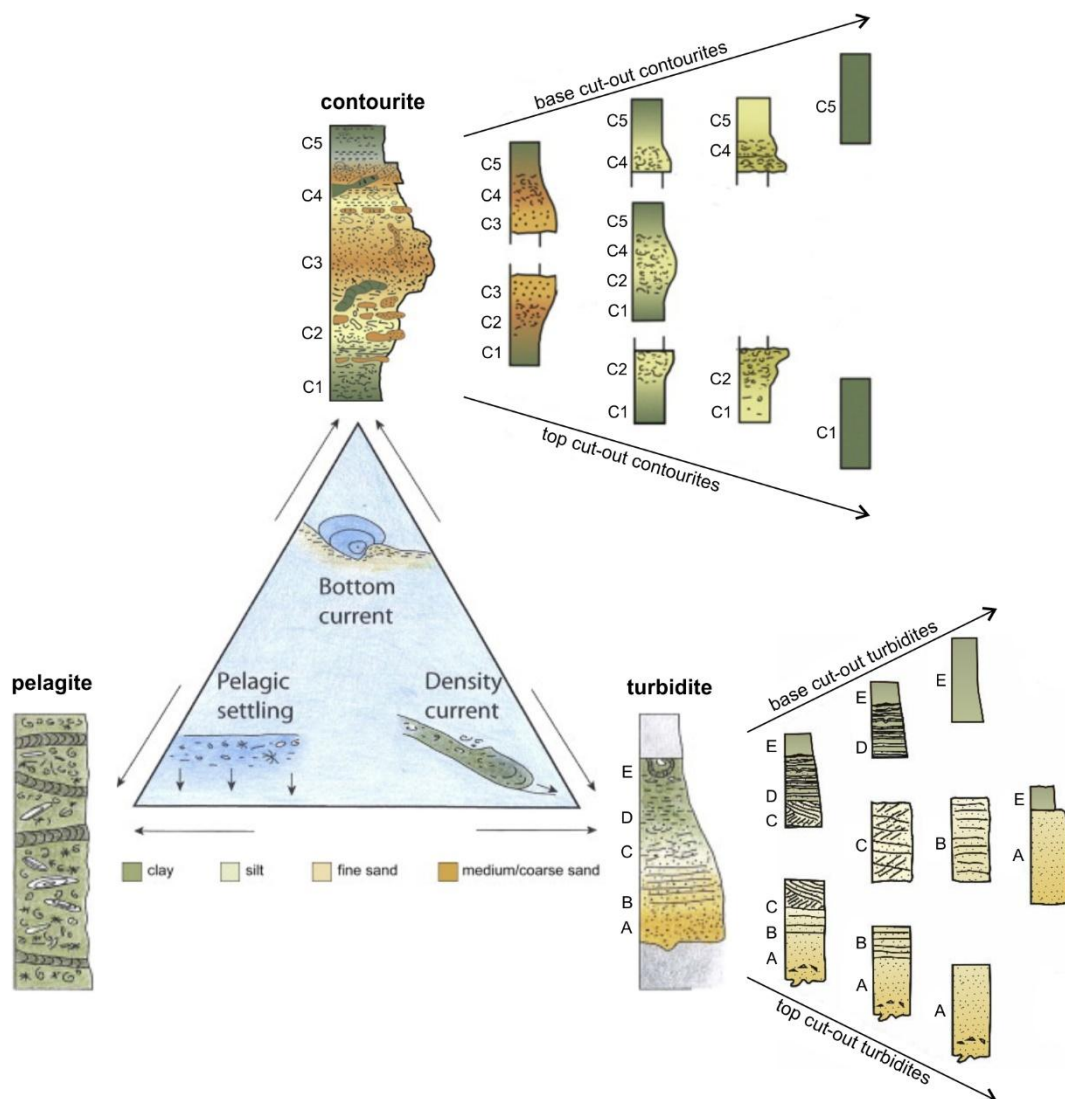


Figure 1.3. Sketch showing the three main types of sedimentary processes active on the continental slope. The complete (ideal) facies models of their deposits (i.e., bi-gradiational contourite sequence (Stow et al., 2002) and Bouma turbidite sequence (Bouma, 1962)) are shown, along with variants commonly found in the sediment record. Adapted from Rebesco et al. (2014) and Stow and Smillie (2020).

Traditional geophysical methods can be used to establish the general characteristics of an area. On a large spatial scale ($10^3 - 10^5$ m), hull-mounted bathymetry and side-scan sonar

measurements visualise the present-day seafloor topography at a horizontal resolution of $10^0 - 10^2$ m, while single- and multi-channel seismic measurements reveal the subsurface structure ($10^0 - 10^1$ m vertical resolution). This baseline information can first be interpreted and then used as a background against which to investigate areas and processes of interest in more detail (Nittrouer and Kravitz, 1996). Underwater robotics (autonomous underwater vehicles, AUVs; remotely operated vehicles, ROVs) can be used to obtain bathymetry and seismic data as well as video or photography data to reveal recent processes at a much higher resolution ($10^{-2} - 10^{-1}$ m) while covering a shorter transect ($10^2 - 10^3$ m). Additionally, very high-resolution point measurements serve as local records that can ground-truth the large-scale data (Nittrouer and Kravitz, 1996). Present-day oceanographic processes can be measured instantaneously and in time series of weeks to months with CTD casts, gliders, or moorings carrying current meters, turbidity sensors, etc., while sediment cores record past processes as variations in lithology (geochemistry, microfossils, grain size, sedimentation rate) (Nittrouer and Kravitz, 1996; Clare et al., 2017). Sediment core scanning and imaging techniques such as computed tomography now enable collection of lithological, geophysical and geochemical data at $10^{-6} - 10^{-2}$ m downcore resolution (Croudace et al., 2006; Cnudde et al., 2011). This has increased the overall potential temporal resolution of sediment analysis in recent decades, but the type of deposits (e.g., seasonal to yearly resolution in marine varves; Schimmelmann et al., 2016) and the sedimentation rate (e.g., coarser-than-millennial-scale resolution in Porcupine Abyssal Plain sediments where sedimentation rates are 0.14 cm/kyr; Carvalho et al., 2011) remain the main controls on actual temporal resolution.

1.2.2 Continental slope sediment characteristics

Even though there are no unique sets of sediment characteristics that define each of the three end-members of deep-sea sediments (i.e., turbidites, contourites, and pelagites), attempts have been made to group sediment properties for each from examples across the globe to provide a framework for their identification (Bouma, 1962; Stow and Tabrez, 1998a; Rebesco et al., 2014; Stow and Smillie, 2020). Some of these characteristics are summarised below.

Turbidites are the resulting deposits of turbidity currents, which consist of a mixture of sand and mud suspended in water (Mulder, 2011). A dense basal layer that transports gravel and coarse sands as bedload may be present underneath the suspension cloud (Paull et al., 2018; Talling et al., 2023). Because of the sediment they contain, the density of turbidity currents is larger than that of the surrounding water mass and they move downslope under their own gravity. Turbidity currents can be the last phase in flow transformation from mass-transport processes such as slides, slumps, and debris flows (Piper et al., 1999; Mulder and Alexander, 2001; Talling et al., 2007; Mulder, 2011). Deep-sea turbidite systems generally consist of an upslope submarine canyon that may transition downslope into a submarine channel with overspill levees and finally a submarine fan at the distal end (Mulder, 2011). These architectural elements can be recognised in bathymetry and seismic data as V-shaped incisions (canyon), U-shaped depressions (channel) flanked by wedge-shaped sediment stacks (levees) and lobes with convex-up topography (fans) (Flood et al., 1991; Mulder, 2011). Turbidites infilling channels can be recognised as high-amplitude reflectors in seismic data (Flood et al., 1991; Deptuck et al., 2003). They are event beds that are deposited on time scales of hours to days and have variable thicknesses of $< 10^{-2}$ m – > 10 m depending on the scale of the source flow and the location of deposition along the flow path (Azpiroz-Zabala et al., 2017; Paull et al., 2018). The turbidite sedimentary facies is the product of a flow that is decreasing in velocity as particles fall out (Mulder, 2011). As such, it can consist of a large variety of lithologies from clay to gravel that often contain primary sedimentary structures such as parallel or cross-lamination. An erosive base and normal grading towards the top of the deposit are features commonly used for turbidite identification (Stow and Smillie, 2020). Bioturbation is common, especially at the top of the deposit, but depends on the frequency of the turbidity currents (i.e., whether sufficient time elapses between events to allow bioturbation to

develop). Typical turbidite facies models for varying grain sizes are presented by Bouma (1962), Stow and Shanmugam (1980), and Lowe (1982), but only < 10 % of turbidites identified in cores, wells, and outcrops present as a complete sequence as presented by those authors (Figure 1.3) (Stow and Smillie, 2020).

Contourites are deposited by bottom currents that may result from a number of processes (Stow et al., 2002; Rebesco et al., 2014). Thermohaline currents flow parallel to the depth contours of oceanic basins, with the Coriolis effect forcing them to the right in the northern hemisphere and to the left in the southern hemisphere. Theoretically, their velocity is dependent on the difference in density with the ambient water as well as the angle of the slope (Rebesco et al., 2014). In reality, obstacles in the bathymetry (e.g., seamounts, submarine canyons, topographic constrictions) can cause local enhancement of bottom current velocity and change of the flow orientation (e.g., Hanebuth et al., 2015; Matossian and Van Rooij, 2024). Bottom currents impacting slope sedimentation are also commonly affected by internal waves and tides oscillating at the interface between water masses of different densities. In the specific setting of a submarine canyon-channel system, internal waves interact with the steep topography and they are often found to enhance bottom currents locally and force bottom currents in an along-axis orientation (i.e., up- and downslope) (Wain et al., 2013; Aslam et al., 2018). In seismic data, contourite bodies are often recognised as mounded or sheeted deposits of parallel to wavy, continuous strata that are often elongated along the depth contours and can be confined by upper and lower unconformities (Nielsen et al., 2008; Rebesco et al., 2014). Bottom currents are capable of producing a range of bedforms from ripples and dunes to furrows and scour depending on the characteristics of the seafloor sediment and the velocity of the flow (Stow et al., 2009). Contourites can range in lithology from clay and silt (muddy and silty contourites) over sand (sandy contourites) to gravel lags and are often recognised by the cyclicity that results from alternating strengthening and weakening of the generating currents, or differing sediment input (Stow et al., 2002). This results in the development of bi-gradational sequences, which is recognised as the features commonly described as contourite facies (Rebesco et al., 2014). As in turbidites, however, observed contourites rarely appear as a complete sequence (*sensu* Stow et al., 2002). The sedimentary record often preserves partial sequences that contain hiatuses due to erosion or non-deposition during peak flow strength (Figure 1.3) (Stow and Smillie, 2020). Primary sedimentary structures are absent in muddy contourites but horizontal lamination and cross-lamination may develop in silty to sandy sections (Rebesco et al., 2014). Many contourites are highly bioturbated throughout, with bioturbation likely to destroy any primary structures and create a homogenous-looking facies.

In (hemi)pelagic deposition, particles are sourced from primary production and/or terrigenous input (from aeolian input, river plumes, icebergs, ...) at the water surface and settle vertically through the water column under the force of gravity (Henrich and Hüneke, 2011). While pelagic deposits are the results of vertical settling only, hemipelagic deposition includes a component of slow lateral advection. Nevertheless, hemipelagites appear as draping strata of continuous thickness in seismic data (Imbert, 2011). Hemipelagites are fine-grained (mean grain size 5-35 μm) deposits characterised by a lack of primary sedimentary structures (Stow and Tabrez, 1998a; Stow and Smillie, 2020). Bioturbation intensity depends on the oxygenation of the bottom waters. A hemipelagic facies model has been proposed by Stow and Tabrez (1998), where any changes in deposition are gradual and linked to variations in terrigenous input into the basin, or productivity at the top of the water column.

1.3 Objectives and research questions

This study focuses on the Northeast Atlantic continental margin offshore southwest Ireland, and more specifically on the Gollum and Kings channel systems in the Porcupine Seabight basin (Figure 1.4). During periods of sea-level highstand, both systems are disconnected from direct terrestrial input. During Quaternary glacial periods of sea-level lowstand, the Celtic Sea Shelf was

exposed and partly covered by the marine-terminating British-Irish Ice Sheet (Clark et al., 2022). The channel systems are thought to have transported material brought in by the ice sheets from the shelf to the deep-sea through turbidity currents (Wheeler et al., 2003). Additionally, both sedimentary processes and associated benthic ecosystems in the Porcupine Seabight are known to be influenced by a northwards-flowing alongslope current. During Quaternary warm periods, this contour current brings warm and saline Mediterranean Outflow Water into the basin (De Mol et al., 2002; Van Rooij et al., 2007; Huvenne et al., 2009; Matossian and Van Rooij, 2024). At the top of the Mediterranean Outflow Water, a permanent pycnocline is present at the interface with the Eastern North Atlantic Central Water, where the density difference leads to the formation of internal waves and enhanced bottom currents over the upper slope (White, 2007; White and Dorschel, 2010). Thus, both alongslope and downslope processes in the Porcupine Seabight are thought to have varied through time and appear linked to climate variability (Wheeler et al., 2003; Van Rooij et al., 2007; Huvenne et al., 2009). This makes this site an ideal study area to reach the first aim of this study: to contribute to the understanding of the interaction between alongslope and downslope processes through the variable Quaternary climate, how it affects deposition and erosion on continental slopes, and how it controls the distribution of particles throughout oceanic basins.

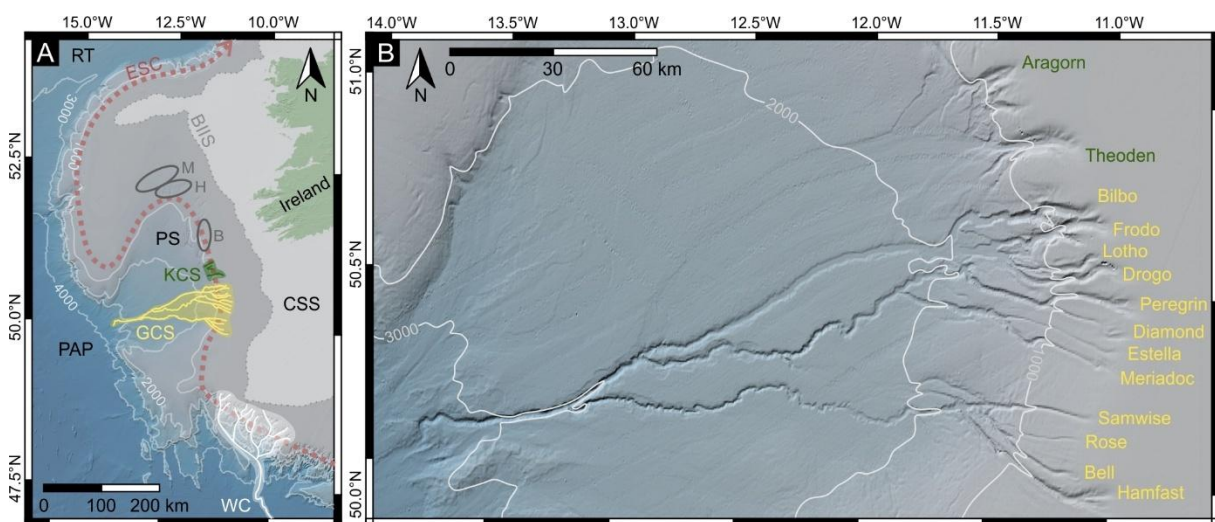


Figure 1.4. (A) Location of the Porcupine Seabight offshore southwest Ireland. BIIS = British-Irish Ice Sheet at its maximum extent during the last glacial period c. 26 ka (from Clark et al. (2022)); main cold-water coral mound provinces are circled in grey: B = Belgica Mound Province, H = Hovland Mound Province, M = Magellan Mound Province; CSS = Celtic Sea Shelf; ESC = European Slope Current; GCS = Gollum Channel System; KCS = Kings Channel System; PAP = Porcupine Abyssal Plain; PS = Porcupine Seabight; RT = Rockall Trough; WC = Whittard Canyon. (B) Bathymetry map of the southern Porcupine Seabight, showing the names of the individual branches of the Kings (green) and Gollum (yellow) channel systems.

Global studies on submarine canyon-channel systems (Matos et al., 2018) are biased towards active shelf-incising systems with efficient connections to continental input through fluvial systems (e.g., Yu et al., 2009; Khripounoff et al., 2012; Babonneau et al., 2013; Azpiroz-Zabala et al., 2017), long-shore drift cells (e.g., Paull et al., 2005; Covault et al., 2007), or sediment resuspension processes on a narrow shelf (e.g., Palanques et al., 2006; Oliveira et al., 2007). Nevertheless, land-detached systems on broad continental shelves have been proven efficient pathways for downslope transport of particles even in the present-day sea-level highstand conditions (e.g., Heijnen et al., 2022), as well as valuable archives for changes in sediment input related to climatic variability (Toucanne et al., 2008). However, the question remains whether these characteristics are applicable to land-detached systems globally, or if specific environmental conditions are necessary to allow activity of similar frequency and intensity to land-attached systems. The Gollum Channel System – presumed inactive at present (Tudhope and Scoffin, 1995; Wheeler et al., 2003) – has been draining the same continental shelf as the

Whittard Canyon, where frequent present-day activity has been recorded (Heijnen et al., 2022). The second aim of this work is to test the (in)activity hypothesis in the Gollum Channel System and improve the understanding of controls on activity in land-detached submarine canyons.

Lastly, although the assumption is that the Kings and Gollum channels received sediment input from the British-Irish Ice Sheet when it extended closest to the shelf edge (Wheeler et al., 2003), the dynamics of this ice sheet at its southwestern edge, along with the position of the ice margin relative to the shelf edge and the efficiency of the connection to the channel heads, are uncertain (Giglio et al., 2021; 2022). Since there are no major fluvial systems on this margin either (Marine Institute, 2016), the mechanisms that supplied sediment to create and maintain such a large-scale canyon-channel system are unclear. Thus, the final aim of this project is to investigate which processes were involved in the formation and evolution of the Kings and Gollum channel systems.

These objectives will be addressed through four research questions:

- 1) When were the Kings and Gollum channel systems formed and which processes were involved in their development?
- 2) Do the downslope processes in the Gollum Channel System interact with the alongslope currents over the eastern slope of the Porcupine Seabight?
- 3) Is downslope activity in the Gollum Channel System linked to Quaternary glacial-interglacial cycles?
- 4) Which processes are active in the Gollum Channel System presently?

Answers to these questions will be sought through analysis of multi-channel seismic data, single-channel seismic data, well data, moored current meter data, hull-mounted multibeam echosounder data, AUV-mounted side-scan sonar data, AUV-mounted photography data, and sediment cores (Figure 1.5). The combination of these geophysical, oceanographic, and sedimentary datasets allows identification of slope processes at basin scale as well as their local variability.

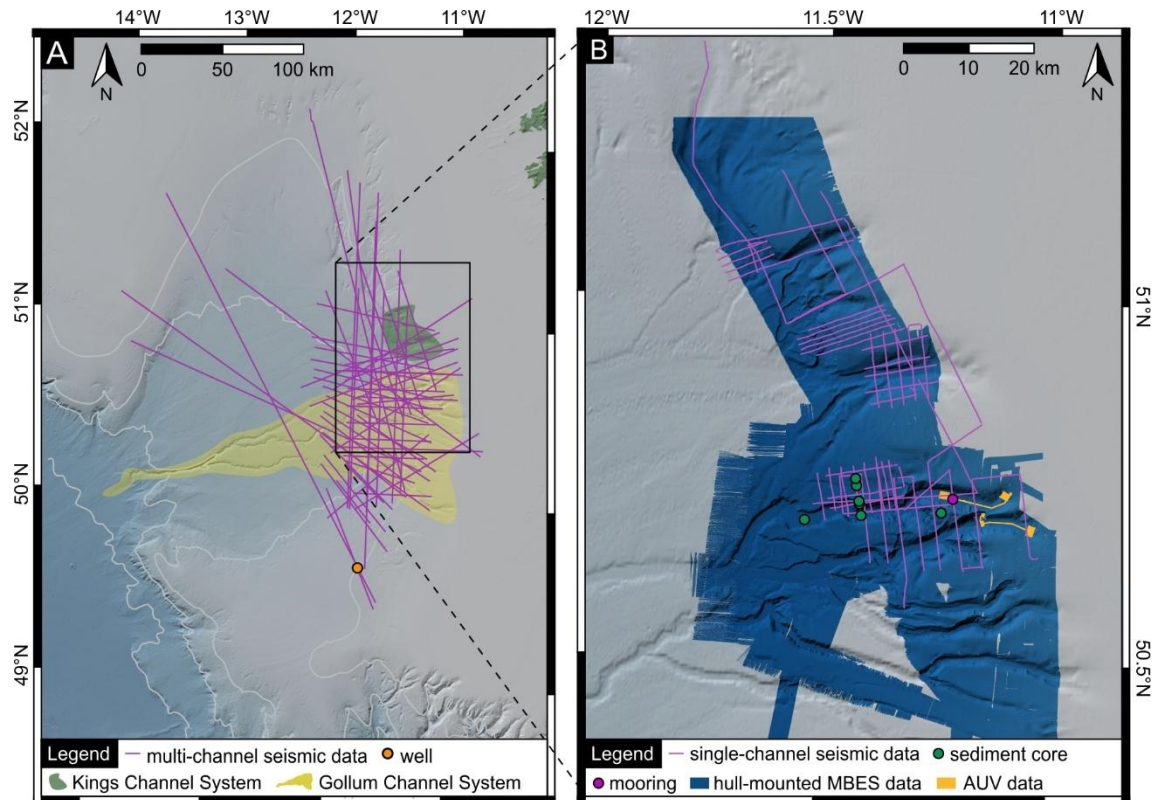


Figure 1.5. Bathymetry maps of (A) the Porcupine Seabight offshore southwest Ireland, (B) the upper eastern slope of the Porcupine Seabight over the Kings Channel System and northern branches of the Gollum Channel System showing the locations of the data analysed in this thesis.

1.4 Study area

The Porcupine Seabight is a pear-shaped basin that forms an embayment on the Northeast Atlantic margin (Figure 1.4). It overlies the Porcupine Basin, which is a failed rift basin that formed from the Permian onwards and experienced a major extensional period related to the opening of the North Atlantic in the Late Jurassic (Shannon, 1991; Naylor and Shannon, 2005; Shannon et al., 2007). Since the Early Cretaceous, the basin has been undergoing thermal subsidence and open marine conditions were first established in the Late Cretaceous (McCann et al., 1995; McDonnell and Shannon, 2001). A period of regression occurred in the late Paleocene–late Eocene (Moore and Shannon, 1992), after which the open marine environment returned in the latest Eocene–earliest Oligocene and persists until today (McDonnell and Shannon, 2001; Stoker et al., 2005b).

Several unconformities have been recognized in the Cenozoic sequence across the Porcupine Seabight and the Rockall Trough to the north (for location, see Figure 1.4A). An uppermost Eocene erosional unconformity (reflector C30 in seismic data) marks the onset of deep-water current circulation and formation of sediment drifts in both basins (Stoker et al., 2001; Stoker et al., 2005b; Shannon et al., 2005). An uppermost lower Miocene regional unconformity (reflector C20 in seismic data) is related to a change in the circulation pattern from Tethyan-influenced to Nordic seas-influenced bottom current activity (Stoker et al., 2005b). Regional submarine erosion (reflector C10 in seismic data) occurred in the early Pliocene, likely due to increased North Atlantic Deep Water input from the Nordic seas along with basin margin tilting (Stoker et al., 2001; 2005a; 2005b). Over the eastern slope of the Porcupine Seabight, studies have recognised two discontinuities: one of early middle Miocene age (seismic reflector RD2 of Van Rooij et al. (2003)) (Louwye et al., 2008), and one representing a hiatus between upper Miocene and upper Pliocene sediments (seismic reflector RD1 of Van Rooij et al. (2003)) (Kano et al., 2007; Louwye et al., 2008). There is strong evidence that the RD1 erosional event is linked to the introduction of

Mediterranean Outflow Water into the Porcupine area (Van Rooij et al., 2003; Khélifi et al., 2009; Delivet et al., 2016). The RD2 and RD1 reflectors have been proposed as equivalents to the regional latest early Miocene (C20) and early Pliocene (C10) unconformities, respectively (Van Rooij et al., 2003), but have never been tied on any seismic line.

Two important water masses in the Porcupine Seabight are the Eastern North Atlantic Central Water (ENACW), present below the seasonal thermocline at c. 200-700 m water depth, and the Mediterranean Outflow Water (MOW) with a high-salinity core at 800-1000 m water depth (Rice et al., 1991; White, 2007). During Quaternary warm periods, they are carried northward by a contour current that is concentrated in the European Slope Current at the continental slope (Rice et al., 1991; White, 2007; Toucanne et al., 2021). At the interface between the ENACW and MOW, a permanent pycnocline is present at c. 700 m water depth (White, 2007; White & Dorschel, 2010). At this water depth, it is possible for freely-propagating internal waves of semi-diurnal frequency or bottom-trapped oscillations of diurnal frequency to be formed that enhance bottom current intensity (White, 2007; White & Dorschel, 2010).

Past scientific interest in the Porcupine Seabight stems largely from the occurrence of cold-water coral mounds that are seated on the RD1 erosional surface in three main provinces: the Hovland, Magellan and Belgica Mound Provinces (Figure 1.4), and several smaller provinces: the Moira, Macnas, Viking, and Enya Mound Provinces (Hovland et al., 1994; Henriët et al., 1998; De Mol et al., 2002; De Cock, 2005; Wilson et al., 2007; Van Rooij et al., 2009; Foubert et al., 2011). Since the discovery of the main provinces in 1994-1997, they have been studied to understand how they are formed and able to survive without any photosynthesis-dependent symbionts as shallow-water corals in the tropics do (Hovland et al., 1994; Henriët et al., 1998; De Mol et al., 2002). This led to drilling to the base of Challenger Mound during International Ocean Drilling Program (IODP) Expedition 307. The data gathered there unveiled a strong relationship between mound growth and climate (Huvenne et al., 2009; Thierens et al., 2010, 2013; Pirlet et al., 2011; Raddatz et al., 2011, 2014). The enhanced bottom currents formed at the ENACW-MOW interface during Quaternary warm periods are thought critical for the corals' nutrient supply as well as the prevention of coral burial by sediments (Raddatz et al., 2014). The interaction between these enhanced bottom currents and the topographic confinement that is created by two chains of coral mounds has allowed a contourite drift to develop between the mounds (Van Rooij et al., 2003; Matossian and Van Rooij, 2024). Thus, studies from the Belgica Mound Province have elucidated on the evolution of alongslope processes in the Porcupine Seabight throughout the Quaternary glacial-interglacial cycles (Van Rooij et al., 2003; Matossian and Van Rooij, 2024).

Downslope processes in the Porcupine Seabight, on the other hand, are less well-studied. The 200 km-long Gollum Channel System (GCS) was first recognised in bathymetry data recorded in 1962 (Brenot and Berthois, 1962) and named by Kenyon et al. (1978). An additional smaller system consisting of four channels was recognised just north of the GCS and named the Kings Channel System (Kenyon, 1987; Beyer et al., 2003, 2007; Van Rooij, 2004). Both channel systems, as well as their individual branches, were named by later researchers after characters in the "The Lord of the Rings" books of Tolkien (1954) in keeping with the tradition started by Roberts (1975) for seafloor features along the Rockall Plateau and Trough (Figure 1.4B) (Kenyon et al., 1978; Van Rooij, 2004).

The Kings and Gollum channel systems are detached from any terrestrial sources at present (Figure 1.4). Sediment cores from the thalweg and levees in the lower reaches of the GCS (> 2800 m water depth) suggest turbidity current activity during the last glacial period and the absence thereof in the Holocene (Kenyon et al., 1998). Observations during submersible dives indicate that reversing bottom currents are the main process presently shaping the channel floors (Tudhope and Scoffin, 1995). Due to its assumed present-day downslope inactivity, the GCS has received little further attention.

1.5 Outline

The outline of the research chapters in this thesis is schematically visualised in Figure 1.6. The aim of each of the chapters is outlined below.

Chapter 2 provides a general characterisation of the study area from the moment of channel formation until the present day. This chapter aims to answer research questions 1 and 3 through analysis of single-channel seismic data, present-day seabed features in bathymetry data, and present-day bottom current measurements (Figure 1.5B). A modified version of this chapter is published as Verweirder et al. (2021).

In chapter 3, the large-scale subsurface structure of the GCS is established, which allows verification of the timing of channel formation as well as the processes responsible to answer research questions 1 and 2. This is achieved through analysis of multi-channel seismic data that are correlated to well data in the southern part of the basin (Figure 1.5A). A modified version of this chapter is published as Verweirder et al. (2023).

Chapter 4 investigates which processes are active in the GCS at present and whether it hosts present-day downslope activity through examination of bedforms from bathymetry, side-scan sonar, and photography data. Additionally, the record of downslope activity after the Last Glacial Maximum is explored in two channel floor sediment gravity cores (Figure 1.5B). Thus, the goal of this chapter is answering research questions 1, 3 and 4. This chapter is in review to be published (Verweirder et al., in review).

In chapter 5, a post-deglaciation record of palaeoceanography over the slope is built from sediment cores just outside of the channels to provide context to the deposition inside the channels (Figure 1.5B). This chapter addresses research questions 1, 2 and 4.

Finally, the results of these research chapters are connected and briefly discussed with respect to the four research questions as well as the margin-wide implications in Chapter 6, which also provides some potential avenues for further research.

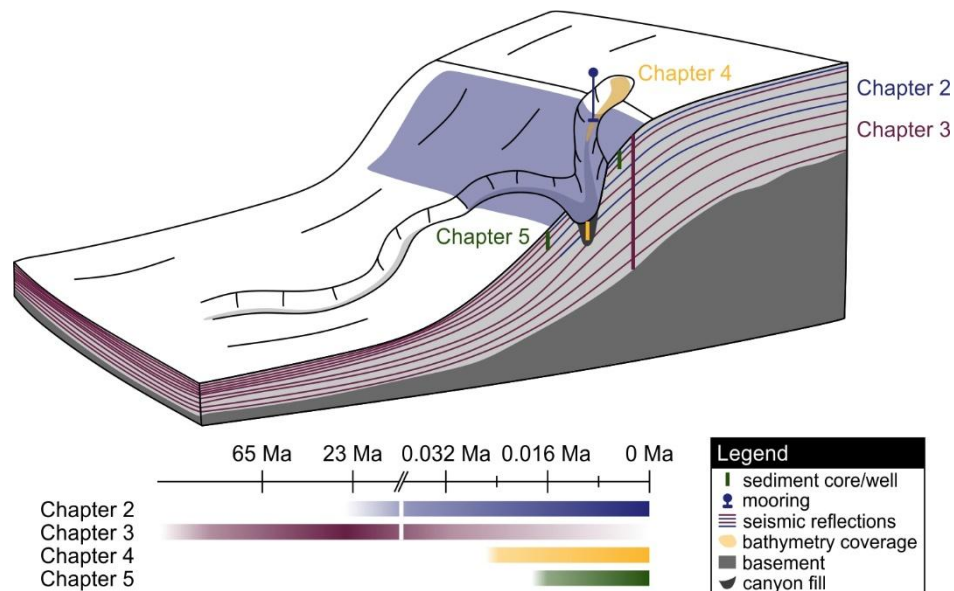


Figure 1.6. Schematic sketch (not to scale) of a section of continental shelf and slope, showing the spatial, temporal, and methodological focus of the four research chapters in this thesis.

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CHAPTER 2 Combined control of bottom and turbidity currents on the origin and evolution of channel systems, examples from the Porcupine Seabight

This chapter has been slightly modified from the published version:

Verweirder, L., Van Rooij, D., White, M., Van Landeghem, K., Bossée, K., and Georgiopoulou, A. (2021). Combined control of bottom and turbidity currents on the origin and evolution of channel systems, examples from the Porcupine Seabight. *Mar Geol* 442, 106639. doi: 10.1016/j.margeo.2021.106639

Abstract

The Gollum Channel System (GCS) and Kings Channel System (KCS) are situated at a key location on the eastern side of the Porcupine Seabight to provide valuable insight into British-Irish Ice Sheet dynamics and sediment supply to the Belgica cold-water coral mound province. These channel systems are the most efficient pathways for particles from the Irish Shelf edge to the Porcupine basin. The spatial and temporal variability of their activity are, therefore, likely to have significant regional consequences. However, the sedimentary processes involved in the evolution of both systems have not yet been comprehensively studied. Here, bathymetric, 2D seismic reflection and oceanographic data are used to reconstruct and compare the interplay between along-, across- and downslope processes through geomorphologic and seismic stratigraphic analyses. The initial seafloor topography of the systems was shaped in the late Miocene-late Pliocene by intense northward-flowing bottom currents during the first phases of the composite RD1 erosion event. The bases of the KCS and GCS were eroded by downslope-flowing turbidity currents during the last phase of the RD1 event. Sediment transport within the channels was probably most active during Quaternary glacial periods of lowered sea levels, and sediment carried downslope by turbidity currents was likely pirated and transported northwards by contour currents. Therefore, the channels are proposed to have been of major importance as a source of sediment and nutrients for the Belgica cold-water coral mounds and associated sediment drifts to the north. The GCS and KCS represent an area where bottom currents, turbidity currents, slope failures and hemipelagic processes have interacted throughout the Neogene and Quaternary.

Keywords Porcupine Seabight, submarine channel system, continental margin processes, seismic stratigraphy, sediment gravity flow, bottom current

Author contributions The manuscript was written by L. Verweirder and improved by discussions with D. Van Rooij, A. Georgiopoulou, M. White, and K. Van Landeghem. Figures were made by L. Verweirder. The seismic data were analysed and interpreted by L. Verweirder and K. Bossée. The current meter data were analysed and interpreted by L. Verweirder and M. White.

2.1 Introduction

Submarine canyons are steep-walled features governed by erosional processes that are found incising shelves and slopes on continental margins across the world (Shepard, 1965b; Harris and Whiteway, 2011; Harris et al., 2014). Canyons may transform basinwards into submarine channels (Babonneau et al., 2002; Deptuck et al., 2003; Cunningham et al., 2005) that are characterised by depositional components such as overbank levees, and have lower thalweg slope angles, a gentler topography and a generally more U-shaped cross-section compared to the upstream V-shaped canyons (Shepard, 1965b; Weaver et al., 2000; Wynn et al., 2007; Amblas et al., 2018). These canyon-channel systems are major pathways for the transport of sediments and organic matter from the shelf and upper slope down to the abyssal plains (Shepard, 1965b; Mulder, 2011; Puig et al., 2014). Given how important they have proven to be in, for example, regional sediment distribution (Amaro et al., 2016; Mountjoy et al., 2018; Maier et al., 2019) and hydrography (Xu and Noble, 2009; Hall et al., 2017; Waterhouse et al., 2017), it is important to understand how such large systems have behaved through time, especially long-lived canyon-channel systems.

The Gollum Channel System (GCS) on the Irish continental margin was first recognised on the bathymetric maps made by Brenot and Berthois (1962) and further described by Kenyon et al. (1978). In total, sixteen slope-confined to shelf-indenting tributary channels have now been distinguished on the upper eastern slope of the Porcupine Seabight (Figure 2.1) (Kenyon et al., 1978; Beyer et al., 2007). The four northernmost channels were found to differ morphologically from the southern ones and were on that basis distinguished as a separate system known as the Kings Channel System (KCS; Figure 2.1A) (Van Rooij, 2004; Beyer et al., 2007). Whereas the channels in the GCS join basinwards into one main channel leading to the Porcupine Abyssal Plain, the Kings channels terminate in an area on the slope that has a complex geomorphology and shows evidence of mass-wasting (Figure 2.1C). At the distal end of the GCS, a deep-sea fan is present but only poorly developed (Rice et al., 1991; Akhmetzhanov et al., 2003), which is thought to be due to reworking of sediments by bottom currents (Rice et al., 1991).

Several sedimentary processes have been suggested as important components in shaping the KCS and GCS. The interplay between these processes, their variability through time, and their influence on channel evolution have not yet been comprehensively studied. They might, however, have significant regional consequences, as the GCS is the only major channel system in the Porcupine Seabight and therefore forms the most efficient pathway along which sediment and organic matter can be transported from the shelf downslope into the basin. This knowledge gap becomes especially obvious when comparing the GCS and KCS to canyon(-channel) systems on the Celtic and Armorican margins. Whittard Canyon, for example, has been investigated extensively with regard to the deglaciation history of the European Ice Sheet (Zaragosi et al., 2000; Bourillet et al., 2003; Toucanne et al., 2008), present-day canyon processes (Cunningham et al., 2005; Amaro et al., 2015; Carter et al., 2018; Daly et al., 2018), habitat suitability (Huvenne et al., 2011; Amaro et al., 2016), and hydrography (Wilson et al., 2015; Hall et al., 2017; Aslam et al., 2018). It has been proposed that during Quaternary glacial lowstands, the GCS formed the extension of a river system on the Irish Shelf (Kenyon et al., 1978; Weaver et al., 2000), although there is a lack of evidence to support this statement. Nevertheless, the proximity of sediment sources during glacial periods probably led to active downslope sediment transport within the system, i.e., mainly through turbiditic or other mass-wasting processes (Tudhope and Scoffin, 1995; Wheeler et al., 2003). Therefore, studying the variability in channel activity could provide insight into ice sheet dynamics during (the most recent) glacial period(s) (Toucanne et al., 2008). Furthermore, northerly contouritic currents infilling the channels have been inferred from geomorphology (Wheeler et al., 2003). Thus, the temporal variability in activity through the GCS and KCS is likely of high importance for sediment supply to the cold-water coral (CWC) mound provinces and contourite drifts that are found just north of the study area (Figure 2.1A) (De Mol et

al., 2002; Dorschel et al., 2007; Van Rooij et al., 2007a, 2009; Lim et al., 2018). Lastly, on the flanks of the channels in the GCS, broad terraces and numerous mass-wasting features such as slide and slump scars were recognised (Kenyon et al., 1978; Wheeler et al., 2003).

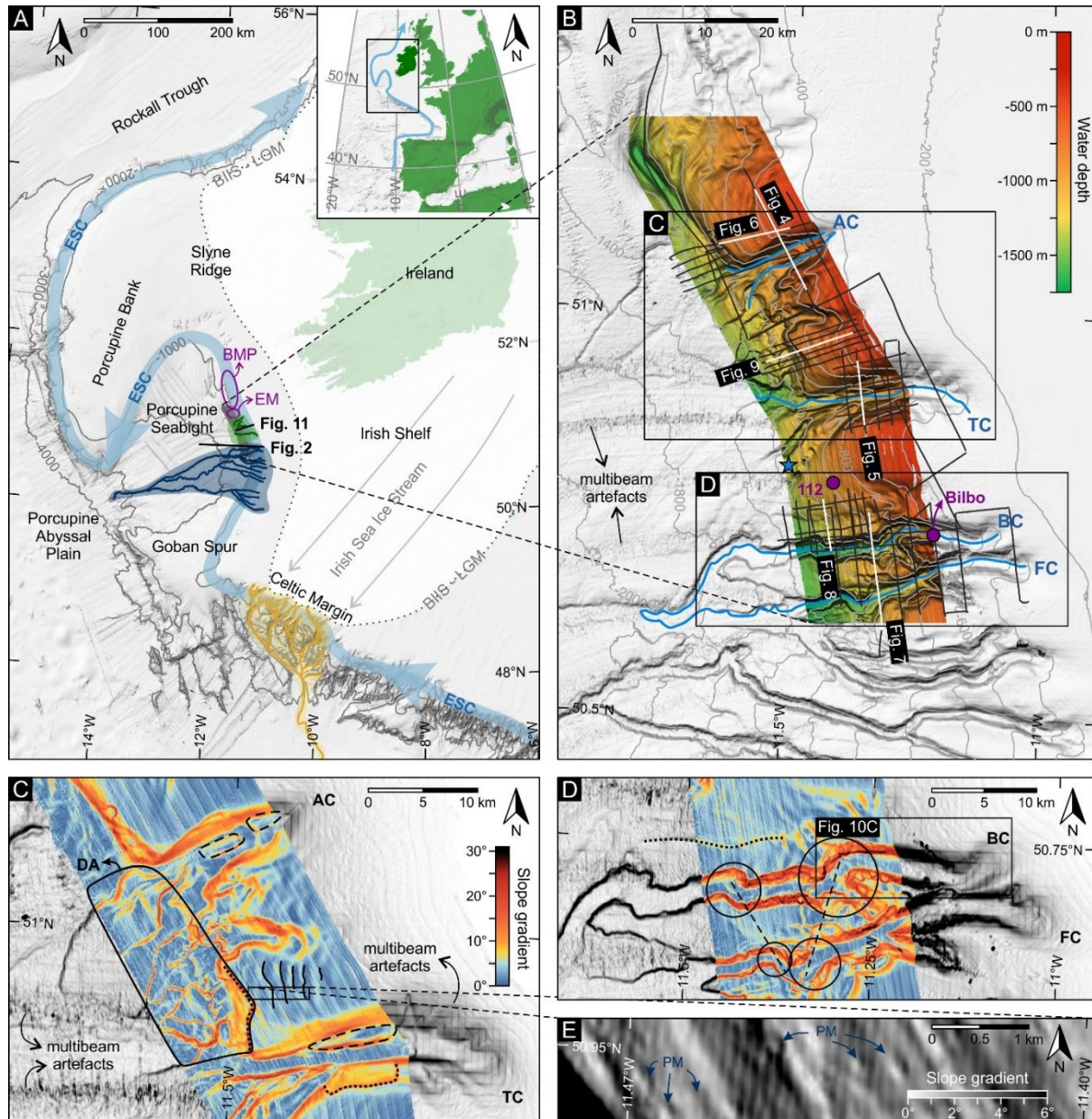


Figure 2.1. (A) Shaded relief bathymetry map showing the location of the Gollum Channel System (GCS; blue shaded area) and Kings Channel System (KCS; green shaded area) within the Porcupine Seabight, southwest of Ireland. The study area is indicated by the dark grey shaded area. The Belgica Mound Province (BMP) and Enya mounds (EM) north of the study area are circled in purple. The Whittard Canyon System on the Celtic Margin is shown in yellow. The European Slope Current (ESC) is indicated with light blue arrows (based on Toucanne et al. (2021)). The maximum extent of the British-Irish Ice Sheet (BIIS) at the Last Glacial Maximum (LGM) is indicated as a white shaded area contoured by a grey dotted line (based on Scourse et al. (2021)). Contour lines are drawn every 1000 m. The inset map shows the location of the Porcupine Seabight on the NW European continental margin. The regional extent of the ESC is shown in light blue. The bathymetry map is adapted from the GEBCO_2014 grid from General Bathymetric Chart of the Oceans (GEBCO; Weatherall et al., 2015). (B) Coloured bathymetry map of the study area (Beyer et al., 2003), with the thalwegs of the four channels under investigation here indicated in blue and with contour lines drawn every 200 m. Also indicated are the seismic profiles available for this study. The purple dots indicate the locations of two moorings, one of which is mooring 112 from Pingree and Le Cann (1990). The blue star indicates the location of the CTD cast that was used for Figure 2.11. (C) Close-up slope angle map of the KCS with indication of the Denethor area (DA; encircled in black), the crests of sediment waves (solid black lines), three internal levees (dashed black lines), and the head-and/or sidewall scarps of two slope failures (dotted black lines): one at the edge of the Denethor area and one on the

southern flank of Theoden Channel. (D) Close-up slope angle map (same legend as in (C)) of the two northernmost channels of the GCS with indication of a scarp north of Bilbo Channel (dotted black line) and sudden changes in orientation in both channels (black circles) seemingly lining up (black dashed lines). (E) Close-up slope angle map of a section of the pockmark (PM) field north of Theoden Channel. The channels in (B)-(D) are labelled as follows: AC = Aragorn Channel, TC = Theoden Channel, BC = Bilbo Channel, FC = Frodo Channel. The grey background maps in (B)-(D) are composed from data from the Irish National Seabed Survey.

The aim of this study is, therefore, to characterise the sedimentary processes that affected the evolution of the KCS and GCS and more specifically the interplay between bottom and turbidity currents. The seabed morphology of four main channels (two in the KCS and two in the GCS; Figure 2.1) and their surrounding slopes is first described. Subsequently, the subsurface sedimentary structures are linked to the existing chronostratigraphic framework and interpreted with regard to the processes that formed them, focussing on their role in regional sediment dynamics. Particular attention is paid to differences between the two channel systems, as well as between the channels of the eastern Porcupine Seabight and the Whittard Canyon on the Celtic Margin.

2.2 Regional setting

2.2.1 Geological setting

The two channels in the KCS and two channel sections in the GCS studied here are situated on the eastern slope of the Porcupine Seabight, a pear-shaped oceanic basin located to the southwest of Ireland (Naylor and Shannon, 1982; Beyer et al., 2003). The basin is more than 3000 m deep in its deepest parts and is bounded by four shallower areas: Slyne Ridge to the north, the Irish Shelf to the east, Goban Spur to the south and the Porcupine Bank to the west. It has only a narrow opening towards the Atlantic Ocean in the southwest, therefore forming an embayment on the eastern North Atlantic continental margin (Figure 2.1A). Underlying the Porcupine Seabight is the Porcupine Basin, which is a failed rift basin that formed from the Permian onwards (Shannon, 1991; Shannon et al., 2007). It experienced a major extensional period related to the opening of the North Atlantic in the Late Jurassic, followed by thermal subsidence from the Early Cretaceous to the Holocene (Shannon, 1991; McCann et al., 1995; Naylor and Shannon, 2005). The basin is now filled with up to 6 km of Mesozoic and up to 4 km of Cenozoic sediments (Moore and Shannon, 1992; Shannon et al., 2007).

Over the eastern side of the Porcupine Seabight, at least two major unconformities in the Cenozoic succession were recognised (Van Rooij et al., 2003) and sampled during Integrated Ocean Drilling Program Expedition 307 (IODP Exp. 307) (Expedition Scientists, 2005). The oldest unconformity, represented by the RD2 seismic reflector (Van Rooij et al., 2003), was attributed an early middle Miocene age (Louwye et al., 2008). It is roughly coeval to the erosional event represented by the C20 seismic reflector in the Rockall Trough and the Porcupine Basin (late early Miocene age) (McDonnell and Shannon, 2001; Stoker et al., 2001), and the two reflectors have been proposed to correspond, but have never been tied on any seismic line (McDonnell and Shannon, 2001; Van Rooij et al., 2003). Seismic data suggest that the C20 reflector is related to bottom current intensification, which was likely due to the introduction of Norwegian Sea Water into the North Atlantic (de Graciansky et al., 1984; McDonnell and Shannon, 2001; Stoker et al., 2005a). The youngest unconformity, identified on seismic data as the RD1 reflector (Van Rooij et al., 2003), represents a large hiatus between upper Miocene and upper Pliocene sediments (Kano et al., 2007; Louwye et al., 2008). There is strong evidence that it is linked to the introduction of Mediterranean Outflow Water into the Porcupine area (Van Rooij et al., 2003; Khélifi et al., 2009; Delivet et al., 2016). It also marks the onset of Quaternary glacial-interglacial cycles and their influence on the oceanographic regime in the North Atlantic (McDonnell and Shannon, 2001; Van Rooij et al., 2003). Van Rooij et al. (2003) proposed that the RD1 discontinuity is equivalent to the intra-Pliocene C10 unconformity that is found across the NE Atlantic margin and thought to originate from tectonic uplift and deep-water erosion (Stoker et al., 2001, 2005a). Based on the

length of the RD1 hiatus at IODP Exp. 307 (late Miocene - late Pliocene), the C10 unconformity (intra-Pliocene) is enclosed within the RD1 unconformity (Stoker et al., 2001, 2005a).

The main sediment sources for the eastern slope of the Porcupine Seabight are the broad Irish and Celtic shelves, and northwards flowing contour currents. The input from the Porcupine Bank is (and has been) likely much lower, which explains why only the eastern edge of the Seabight is characterised by channels (Rice et al., 1991). Numerous fine-grained turbidites have been found in cores in channel thalwegs and flanks downslope of the study area (Kenyon et al., 1998) and on the continental rise at the western end of the system (Akhmetzhanov et al., 2003). However, there are no indications of recent turbiditic transport through the channels, and both visual and core data suggest that downslope processes in the GCS have been inactive during the Holocene (Tudhope and Scoffin, 1995; Kenyon et al., 1998; Van Rooij et al., 2002; Akhmetzhanov et al., 2003; Wheeler et al., 2003). As the eastern slope of the Porcupine Seabight was situated 30-60 km from the ice limit of the British-Irish Ice Sheet (BIIS) during the Last Glacial Maximum (LGM) (Figure 2.1A; Scourse et al., 2021), late Quaternary sedimentation is noticeably influenced by glacial activity. Ice-rafted debris, including dropstones, and increased sedimentation rates are related to the deglaciation of the area and retreat of the BIIS from the Irish Shelf after the LGM (Tudhope and Scoffin, 1995; Van Rooij et al., 2007a, 2007b). Reconstruction of the offshore extent of the BIIS before the LGM is problematic. Onshore, however, its lateral extent has been larger prior to the LGM, in particular during marine isotope stage 12 (MIS12) (Sejrup et al., 2005). Therefore, it could be expected that the offshore BIIS extent southwest of Ireland has been similar to the extent at the LGM, at least during MIS12. Finally, the importance of bottom currents for the depositional regime is demonstrated by the presence of contourite drifts associated with the CWC mounds of the Belgica Mound Province (BMP) (Van Rooij et al., 2003, 2007b, 2009), asymmetrical megaripples in the BMP (Lim et al., 2018), and asymmetrical cross-channel profiles in the GCS (Wheeler et al., 2003). In the upper section of one of the channels of the GCS, Tudhope and Scoffin (1995) reported ripples consisting of moderately well-sorted silty sands, and suggested they result from reversing currents within the channel.

2.2.2 Oceanographic setting

Within the Porcupine Seabight, bottom currents show a cyclonic flow pattern, flowing northwards along the eastern edge (White, 2007). Bottom residual currents flow on average towards 309°N, in a poleward and slightly off-slope direction, with a velocity of 2.5 cm/s (mooring 112 in Pingree and Le Cann (1990); location see Figure 2.1B). The present-day water masses in the Seabight have been characterised as warm and saline Eastern North Atlantic Central Water (ENACW) in the upper 600 m, carried northward as part of the European Slope Current (ESC), and overlying Mediterranean Outflow Water (MOW) with a high-salinity core between 800 and 1000 m depth (Figure 2.2) (Pingree and Le Cann, 1989; White, 2007; Toucanne et al., 2021). Additionally, low-salinity Labrador Sea Water (LSW) is present at about 1500-1900 m depth (Rice et al., 1991; Wienberg et al., 2020) and North East Atlantic Deep Water (NEADW) with influences of overflow waters from NE Atlantic basins below that (Rice et al., 1991; Van Aken, 2000). In particular, the long-term variation of the MOW has been a major influence on sedimentation in the Porcupine Seabight (Van Rooij et al., 2007a; White, 2007). At the permanent thermocline, which is situated at the depth of the boundary between the MOW and the overlying ENACW, freely propagating semi-diurnal or bottom-trapped diurnal internal tides (internal waves at tidal frequency) can be generated between 600 and 900 m water depth (Figure 2.2) (Pingree and Le Cann, 1990; White, 2007; White and Dorschel, 2010). Both processes can enhance bottom current velocities, which may lead to resuspension and transport of sediments, thus affecting the slope morphology (White and Dorschel, 2010; Lamb, 2014).

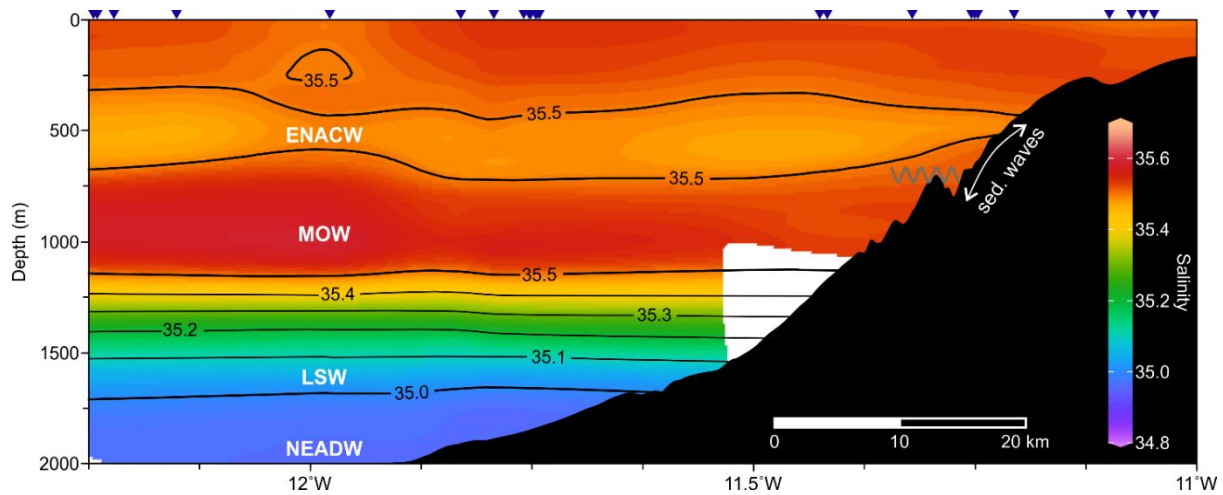


Figure 2.2. Salinity versus depth profile of a section within the Porcupine Seabight crossing the eastern slope in between the KCS and GCS (location: see Figure 2.1A), based on CTD data (blue triangles) of the World Ocean Database (Boyer et al., 2018) and produced using Ocean Data View (Schlitzer, 2020). The white zone near the slope between 1000 and 1500 m water depth is due to a lack of data in that area. Salinity contour lines are drawn every 0.1. The depth range of the sediment waves occurring north of Theoden Channel is indicated by the white arrow. Also added (in white text) are the different water masses present: ENACW = Eastern North Atlantic Central Water, MOW = Mediterranean Outflow Water, LSW = Labrador Sea Water, NEADW = North East Atlantic Deep Water, NSW = Norwegian Sea Water. Internal waves, indicated in grey, can occur at the interface between ENACW and MOW (White, 2007).

2.3 Material and methods

The study area comprises the upper slope section of the two main channels in the KCS and of the two northernmost channels of the GCS, which are the four channels that are covered by the multibeam bathymetry of Beyer et al. (2003) and by a grid of single-channel high-resolution seismic profiles (Figure 2.1B). To better understand the local hydrography within a channel, data from a mooring within Bilbo Channel is added to the oceanographic information that is available from literature (Pingree and Le Cann, 1989; White, 2007).

2.3.1 Multibeam bathymetry

The bathymetry data used here are part of a larger dataset acquired during RV Polarstern cruise ANT XVII/4 in 2000 using a Hydrosweep DS-2 multibeam echosounder. The acquisition details are discussed in Beyer et al. (2003). The relevant section of the resulting DTM (with 50 m grid spacing) is here used without any additional processing (coloured area in Figure 2.1B).

In the bathymetry analysis, the channel lengths are measured along the channel thalwegs defined in Figure 2.1B. Channel width, channel depth and channel floor width are measured perpendicular to the thalweg and are the distances between, respectively, the two channel shoulders, the shallowest of the two shoulders and the thalweg, and the two points at the base of the channel where the slope angle changes abruptly (Figure 2.3).

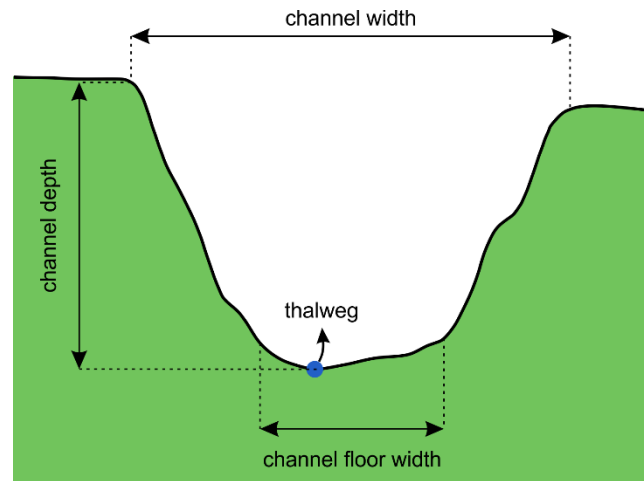


Figure 2.3. Sketch showing the definitions of channel dimensions used in the bathymetry analysis.

2.3.2 Single-channel seismic profiling

The single-channel seismic reflection profiles (Figure 2.1B) were collected during five campaigns in 1999 (Belgica 99/13), 2001 (Belgica 01/12), 2003 (Belgica 03/13), 2006 (Belgica 2006/13) and 2018 (Belgica 2018/14). They were all obtained with RV Belgica, using an SIG sparker source and a single-channel SIG streamer. The main characteristics of each survey are summarised in Table 2.1. The profiles have a vertical resolution of about 1.5 m and the penetration depth is about 400 ms two-way travel time (TWT). The raw data were processed in RadExPro 2016.3 using an Ormsby bandpass filter, amplitude correction and a swell filter, where the settings for each filter were tailored for each profile. For some of the 1999 profiles, an F-K filter was applied as well. After processing, IHS Kingdom Suite software was used to plot and visualise the seismic profiles and trace the reflectors.

Table 2.1. Main technical specifications of the seismic reflection acquisition during each of the RV Belgica campaigns.

	1999	2001	2003	2006	2018
Source	sparker	sparker	sparker	sparker	sparker
Energy	500 J	500 J	500 J	500 J	600 J
Sampling frequency	4 kHz	8 kHz	5 kHz	8 kHz	8 kHz
Shot interval	2.5-4 s	3 s	3 s	3 s	3 s
Vessel velocity	3-4.5 kn	2-4.1 kn	0.3-3.3 kn	3 kn	3 kn

2.3.3 Oceanographic data

The mooring in Bilbo Channel sat underwater at a depth of 900 m for a period of 325 days, between June 2005 (Julian day 166) and May 2006 (Julian Day 126). It was located on the floor of the channel (Figure 2.1B) and had one current meter at 8 m above the seafloor providing a measurement every 30 minutes. Due to biofouling, the current data is accurate only until December 2005. Therefore, current data from mid-December (Julian day 351) until the end of the measuring period is omitted from the analyses.

To estimate how internal waves interact with the seabed in the study area, the ratio of the slope angle γ to the angle of the internal wave energy propagation c is calculated over one slope transect (see Figure 2.1A for location). Here, γ is calculated from bathymetry data, taking the slope correction into account (after Ribó et al. (2016)), while c is determined from the following formula (Cacchione et al., 2002):

$$c = \sqrt{\frac{\sigma^2 - f^2}{N^2 - \sigma^2}}$$

Herein, the internal wave frequency σ (expressed in cycles per hour, cph) is dependent on the dominant tidal frequency, which is here determined from the Bilbo mooring current meter data. The Coriolis frequency f [cph] at latitude φ is given by $f = (\sin \varphi)/12$. The Brunt-Väisälä frequency N [cph] varies with water depth and is calculated from a World Ocean Database 18 (WOD18) CTD station in the study area (Boyer et al., 2018; see Figure 2.1B for location) using Ocean Data View (Schlitzer, 2020). The obtained values for N are averaged into depth intervals of 10 m to reduce noise.

2.4 Results

2.4.1 Seabed morphology

2.4.1.1 Kings Channel System

The KCS is, except for two channel heads, completely covered by the multibeam bathymetry of Beyer et al. (2003) (Figure 2.1B). The two largest channels of the system are Aragorn Channel in the north and Theoden Channel in the south (Figure 2.1C). Both channels are U-shaped in cross-section. Aragorn Channel is oriented roughly ENE-WSW, is about 13 km long, 4.7 to 6.3 km wide and on average 300 m deep, with the channel depth increasing from 220 m in the east to almost 400 m in the west. The channel floor has an average slope gradient of 2.6° and is about 2.3 km wide in the eastern part, widening to 4.5 km towards the west. In the centre of the channel, an elongated mounded feature of 3.8 km long and 1.2 km wide is elevated maximum 100 m above the adjacent channel floor and separates the channel into two flow paths (Figure 2.1C). A second mounded feature of about 45 m high lying against the northern channel flank measures 3.2 km in length and 1 km in width. The northern channel wall, with slope values of about 10-16° and up to 22°, is a bit steeper than the southern flank, where the slope measures 7-14° and up to 22°.

Theoden Channel is also oriented ENE-WSW, is 23 km long, and widens first from 4 to almost 6 km, after which it narrows again to about 3 km. The channel floor has a relatively constant width of about 1.3 km and an average basinward-dipping slope of about 2.2°. The depth of Theoden Channel first increases westwards, from 120 to 520 m, and then decreases again to 200 m. The northern flank of the channel has typical slope angles of 7-13° and up to 17°. The southern flank is a bit steeper with angles of 10-17°. Inside the channel against the northern channel flank lies an elongated mound of sediment that is about 7.5 km long, maximum 1.5 km in width, and at its highest point a little under 200 m above the adjacent channel floor. The southern flank of Theoden Channel is characterised by a clearly visible slide scar of 11 km² (Figure 2.1C). On the smooth slopes north of Aragorn Channel and north of Theoden Channel, respectively 17 and 28 pockmarks with a diameter of about 150 m can be identified (Figure 2.1E). It is likely that these pockmark fields contain even more individual pockmarks, as the numbers reported here are limited by the artefacts and the resolution of the bathymetry map (50 m; Beyer et al., 2003).

The Denethor area (Figure 2.1C) covers at least 150 km², although its western extension is difficult to distinguish as the area is not fully enclosed in the multibeam data of Beyer et al. (2003). It has a very chaotic bathymetry where many scarps of 40-100 m high divide the seafloor into several blocks, ridges, and terrace-like features. Many of these scarps seem to be continuations of the flanks of Aragorn and Theoden channels, as well as the smaller channels between them. The largest of the scarps, however, is oriented NW-SE and perpendicular to the channel axes (Figure 2.1C). At this scarp, the slope angle is 19-26° and the depth difference is about 100 m over a horizontal distance of 350-450 m.

The slope to the east of the large scarp, in between Theoden Channel and a smaller channel, dips 2-6° to the SSW and is mostly smooth, aside from the pockmark field and five straight to slightly

sinuous seafloor undulations (Figure 2.1C). These undulations are about 1-4.5 km long and run parallel to each other with a general N-S orientation (170-175°N), roughly following the contour lines. They have an average wavelength of 1240 m (range = 730-2020 m) and an average height of 50 m (range = 20-110 m).

2.4.1.2 Gollum Channel System

Within the sections of Bilbo and Frodo channels that are presented here (approximately between the 500 and 1500 m water depth contours; Figure 2.1B), the axes of both channels are generally straight and oriented ENE-WSW. However, in both channel sections, there are two places where the orientation changes by almost 90° that seem to be lined up between the two channels (Figure 2.1D). The walls of Bilbo Channel have an average slope of 14-24°, but in some places can be as steep as 30°. The channel flanks of Frodo Channel have a slope that varies mostly between 10 and 20°, but the steepest sections also have a slope angle of up to 30°. In both channels, the northern channel flank is slightly steeper than the southern flank. Both channels are U-shaped in cross-section and both channel floors slope towards the basin at an average angle of about 2.2°. The depth of Bilbo Channel decreases from about 450 m in the east to about 250 m in the west. Its width is constant at 3.5-4 km, while the width of the channel floor is about 1.5 km throughout. The depth of Frodo Channel also decreases westwards, from about 350 m to 200 m. The width of the channel varies between 3.7 km (in the east) and 2.0 km (in the west), while the channel floor width is more constant varying between 0.8 and 1.2 km. A multitude of slide scars are visible on the channel walls. North of Bilbo Channel, the slope features a long scarp that is about 12.5 km in length and parallel to the channel axis (Figure 2.1D). The difference in water depth over the scarp from north to south is about 40 m over a horizontal distance of about 450 m. The slope angle of the scarp is a bit higher (6-8°) than the slope of the surrounding seafloor (3-5°).

2.4.2 Seismic stratigraphy

The nomenclature for the seismic units and erosional surfaces observed here was adopted from the common nomenclature used on the eastern slope of the Porcupine Seabight (Van Rooij et al., 2007b, 2009).

2.4.2.1 Unit U3

The lowermost unit visible on the seismic profiles in the KCS consists of high-amplitude continuous reflections that are mostly (sub)parallel and dipping towards the basin (Figure 2.4-2.6). The upper 50-100 ms TWT (40-80 m using a P-wave velocity of 1600 m/s; Expedition Scientists, 2005) of the unit is organised in a wavy configuration in several places throughout the KCS (Figure 2.6). The top of unit U3 is truncated by a regional discontinuity called RD2 (Figure 2.4-2.6). In the GCS, unit U3 is not visible on any of the profiles.

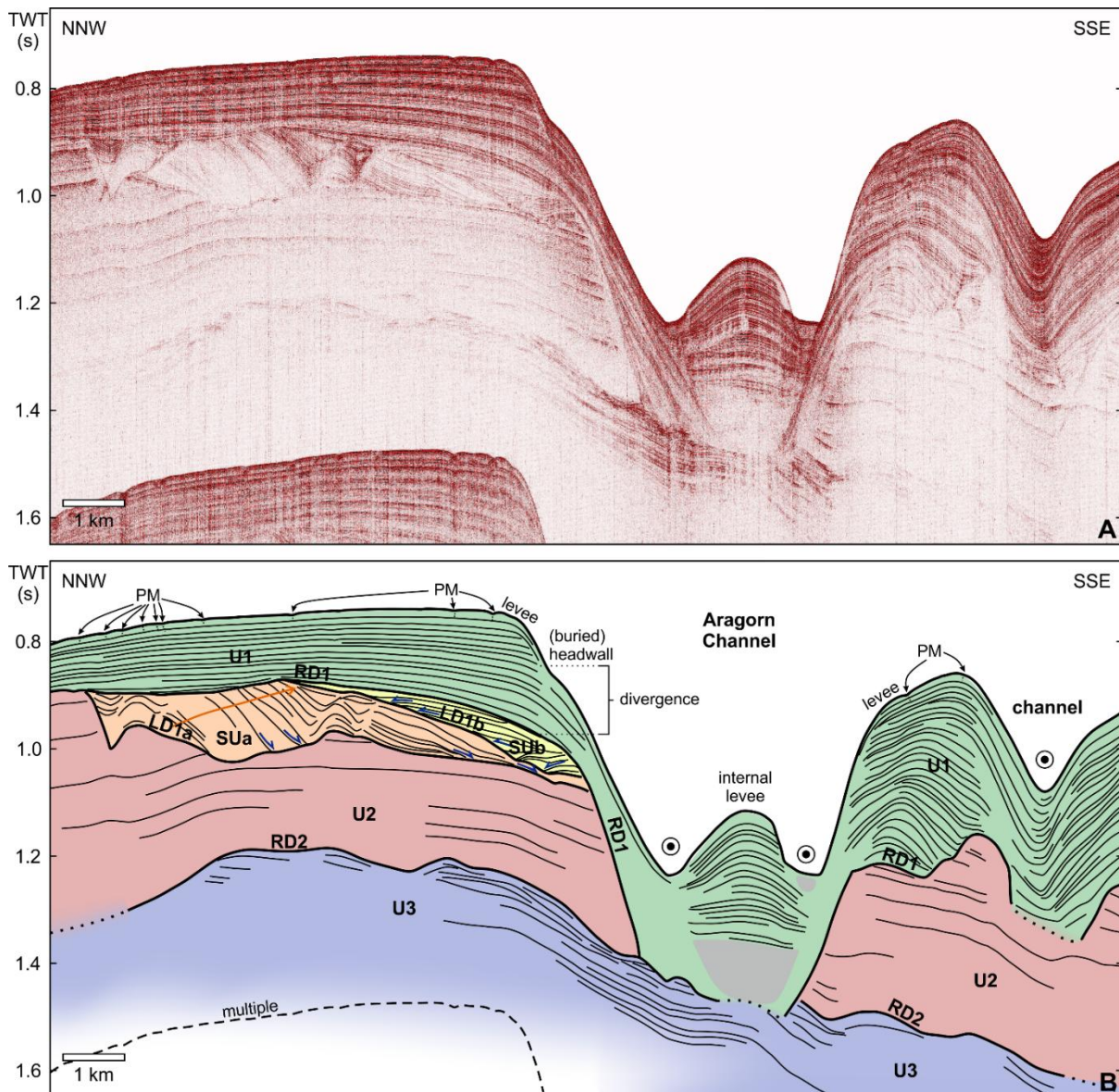


Figure 2.4. (A) Seismic profile (location: see Figure 2.1B) crossing Aragorn Channel and a smaller channel in the KCS with (B) interpretation indicating the seismic stratigraphic units and their boundaries as described in the text. PM indicates the locations of pockmarks recognised on the profile. The orange arrow in unit SUa indicates a southward progradation of reflectors. The blue arrows in units SUa and SUB indicate on- or downlap onto the LD1a and LD1b surfaces, respectively. Grey shaded areas indicate zones with chaotic or transparent seismic facies in the channel infill.

2.4.2.2 Unit U2

The RD2 erosional unconformity is the base of unit U2, which is almost acoustically transparent in the northern part of the study area. Parallel, basinward-dipping reflections can be distinguished, but with a lower amplitude and spaced further apart than those of unit U3 (Figure 2.4, 2.6). Towards the southern end of the study area, however, U2 has fewer transparent sections and the parallel reflections are spaced more closely together (Figure 2.5). The upper boundary of unit U2 is another regional discontinuity called RD1 (Figure 2.4-2.9). Unit U2 has a very variable thickness throughout the study area, with values on average around 300 ms TWT (240 m), but with a maximum of around 450 ms TWT (360 m) on the interfluvial areas and disappearing completely in the channels. North of Bilbo Channel at the top of the unit, a package of U2 reflections is truncated by an internal U2 reflection (Figure 2.8).

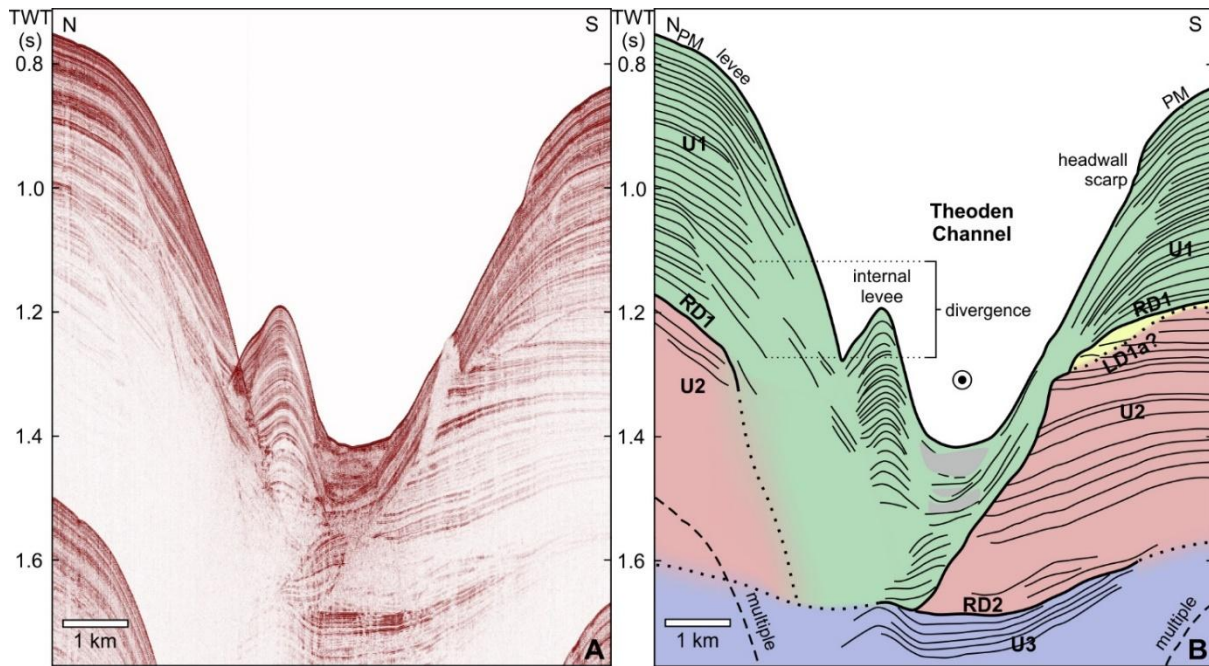


Figure 2.5. (A) Seismic profile (location: see Figure 2.1B) crossing Theoden Channel in the KCS with (B) interpretation indicating the seismic stratigraphic units and their boundaries as described in the text. Grey shaded areas indicate zones with chaotic or transparent seismic facies in the channel infill. PM indicates the locations of pockmarks recognised on the profile.

On the northern flank of Aragorn Channel, two local subunits can be distinguished between unit U2 and RD1 (Figure 2.4, 2.6). They are confined by erosional surfaces and their seismic facies is different from unit U2 below. The lowermost of the two subunits is called SUa. It consists of reflections of varying amplitudes that have a wavy character upslope and a parallel, downslope-dipping character towards the foot of the slope (Figure 2.6). The reflections are down- or onlapping onto the lower erosional bounding surface, which is a local discontinuity called LD1a. In NNW-SSE profiles, the reflections of SUa are dipping to the south with varying angles and prograding in the same direction (Figure 2.4). In WSW-ENE profiles, the wavy reflections can be seen migrating upslope (Figure 2.6). The top of SUa is truncated by either RD1 or another local erosional surface called LD1b. Above SUa, the second subunit SUB is even more laterally restricted (Figure 2.4, 2.6). The reflections in this subunit are subparallel and dipping towards the basin more gently compared to those in SUa. At the bottom of SUB, reflections are on- or downlapping onto LD1b, while at the top they are truncated by RD1.

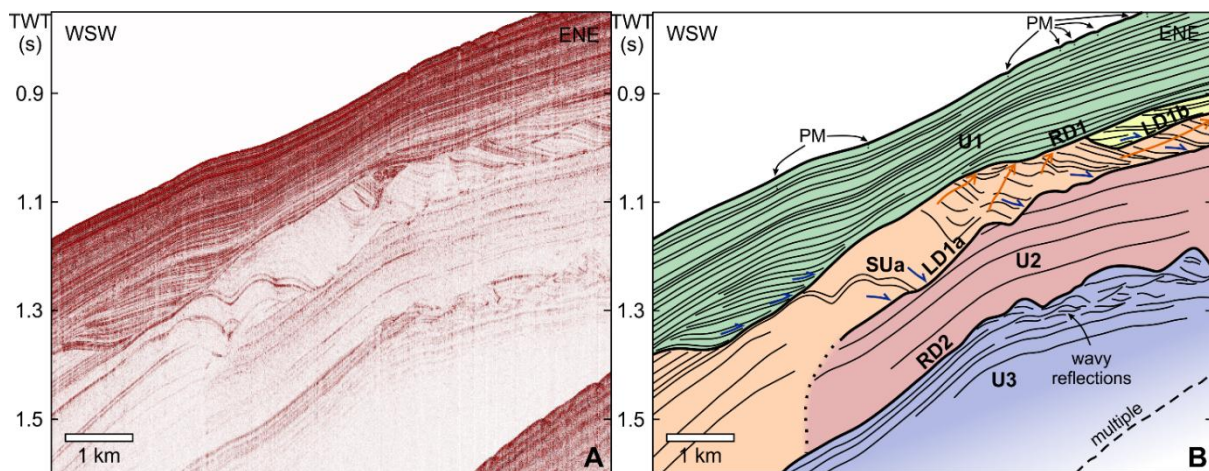


Figure 2.6. (A) Seismic profile (location: see Figure 2.1B) crossing the slope north of Aragorn Channel in the KCS with (B) interpretation indicating the seismic stratigraphic units and their boundaries as described in the text. The orange

arrows indicate upslope migration of wavy reflectors in unit SUa. The blue arrows in unit U1 indicate onlap onto the RD1 reflector. The blue arrows in units SUa and SUB indicate on- or downlap onto the LD1a and LD1b surfaces, respectively. PM indicates the locations of pockmarks recognised on the profile.

2.4.2.3 Regional discontinuity RD1

The upper boundary of unit U2 is the laterally extensive RD1 erosional surface. It has an irregular topography on the interfluvies in the KCS, truncating reflections of unit U2 below (Figure 2.4-2.6). The steepest slopes of the RD1 reflector, however, can be found underneath the channel edges, where it creates 0.4 s TWT (300 m) deep and 2-5 km wide U-shaped incisions in the paleotopography (Figure 2.4, 2.5). In those locations, unit U2 is completely eroded and sediments at the top of unit U3 may have been removed as well. On the slope north of Theoden Channel and on the interfluvies of the GCS, the RD1 reflector is generally less erosional and overlies U2 conformably (Figure 2.7-2.9). The steepest slopes in the paleotopography of RD1 in the GCS are again visible on the channel edges, where the RD1 reflector cuts 0.4 s TWT (300 m) into U2 sediments, creating narrow (0.4-3 km), V-shaped incisions underneath the present-day channels.

RD1 truncates the reflections of the two local subunits north of Aragorn Channel (SUa and SUB) and the local erosional surfaces that bound them (LD1a and LD1b; Figure 2.4, 2.6). Although the lateral extent of LD1b is very limited, LD1a is more extensive and created a rugged topography, potentially limiting the downslope extent of U2 sediments north of Aragorn Channel (Figure 2.6).

2.4.2.4 Unit U1

Above the RD1 unconformity, the uppermost seismic unit is called unit U1. On the slopes in between the channels, it consists of parallel-layered, high-amplitude reflections (Figure 2.6-2.9). The same unit also constitutes the channel fills and in that setting typically includes more chaotic and transparent sections (Figure 2.4, 2.5, 2.7). U1 reflections generally drape the RD1 unconformity conformably, but they can also be found on- or downlapping, especially where the RD1 event created a steep topography (Figure 2.4, 2.6, 2.7).

The stratification within unit U1 is usually very distinct, parallel and continuous, although there are some exceptions. Firstly, at the very top of the unit on the slopes north of Aragorn Channel and north and south of Theoden Channel, the reflections are interrupted by short (sub)vertical successions of small diffraction hyperbolas (Figure 2.4-2.6). Several of these successions join the pockmarks on the seabed. Secondly, on the southern flank of Theoden Channel, unit U1 includes several erosional events (Figure 2.5). Lastly, on the slope just north of Bilbo Channel, a large discontinuity (Figure 2.8) is situated underneath the scarp in the bathymetry parallel to Bilbo Channel (Figure 2.1D). It is related to the truncation of a package of sediments at the top of unit U2 (Figure 2.8).

On the northern channel flank of Aragorn Channel, U1 is relatively thin (100-150 ms TWT or 80-120 m) compared to the southern Aragorn Channel flank (300-400 ms TWT or 240-320 m) and surrounding Theoden Channel (200-400 ms TWT or 160-320 m). North of Bilbo Channel, the thickness of U1 decreases from 300 ms TWT (240 m) in the east to 200 ms TWT (160 m) in the west (Figure 2.7, 2.8).

On the slope north of Theoden Channel, undulating reflections are present from the seafloor down to the penetration depth of the seismic signal (Figure 2.9). They make up the complete U1 sequence and even reach beyond the RD1 reflector, which has a conformable character in these profiles, into unit U2. They are usually laterally continuous, although some onlapping of reflections from the upslope flank of one undulation onto the downslope flank of the next undulation is visible just above and just underneath RD1 (Figure 2.9). Especially underneath RD1 and in the lowest 150 ms TWT of unit U1, the upslope flanks of the undulations are thicker than their downslope flanks, causing them to migrate in an upslope direction. The thickness of both

flanks of each undulation becomes more equal in the more recent deposits, and the undulations transition into a more aggradational configuration (Figure 2.9).

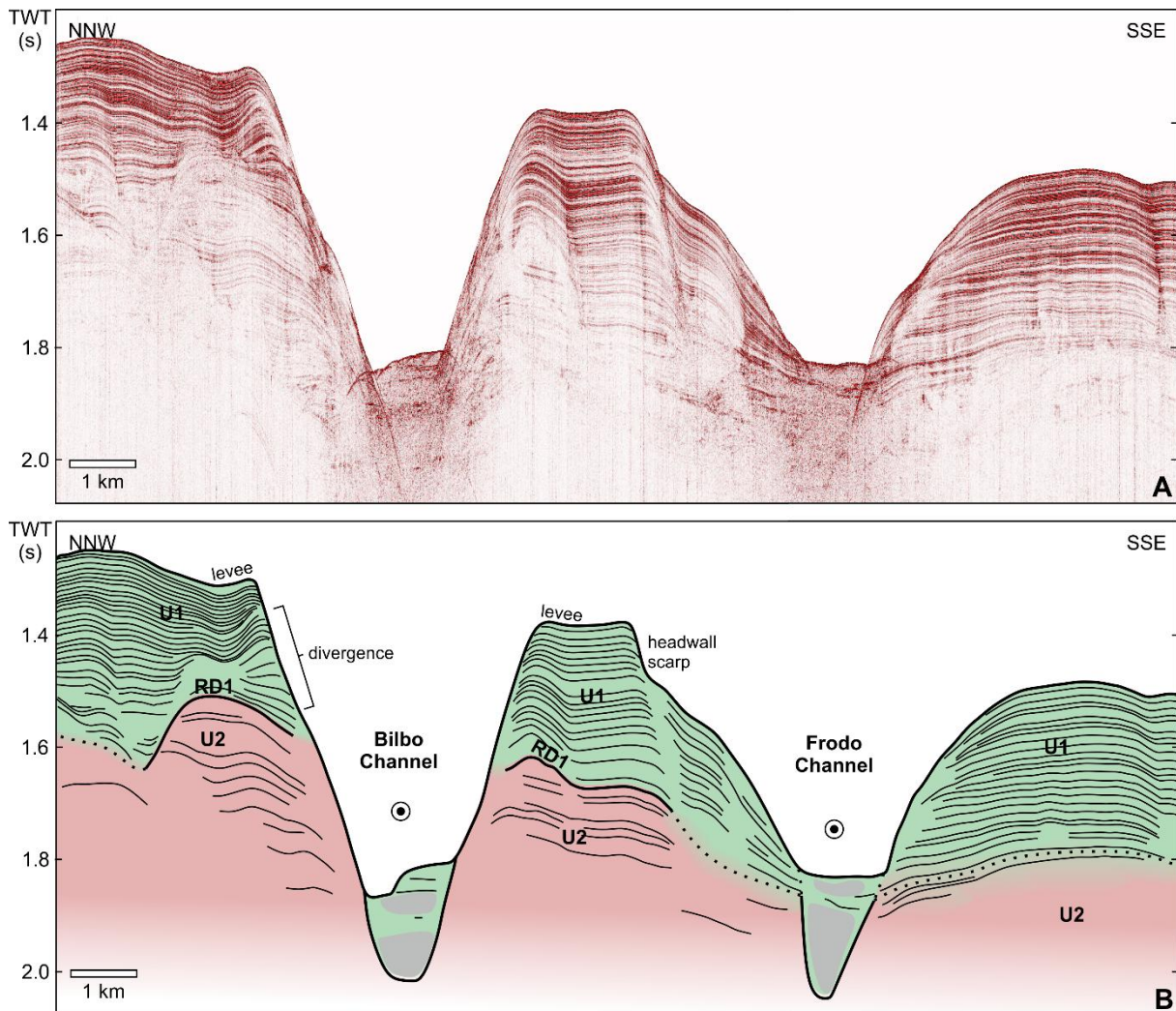


Figure 2.7. (A) Seismic profile (location: see Figure 2.1B) crossing Bilbo and Frodo channels in the GCS with (B) interpretation indicating the seismic stratigraphic units and their boundaries as described in the text. Grey shaded areas indicate zones with chaotic or transparent seismic facies in the channel infill.

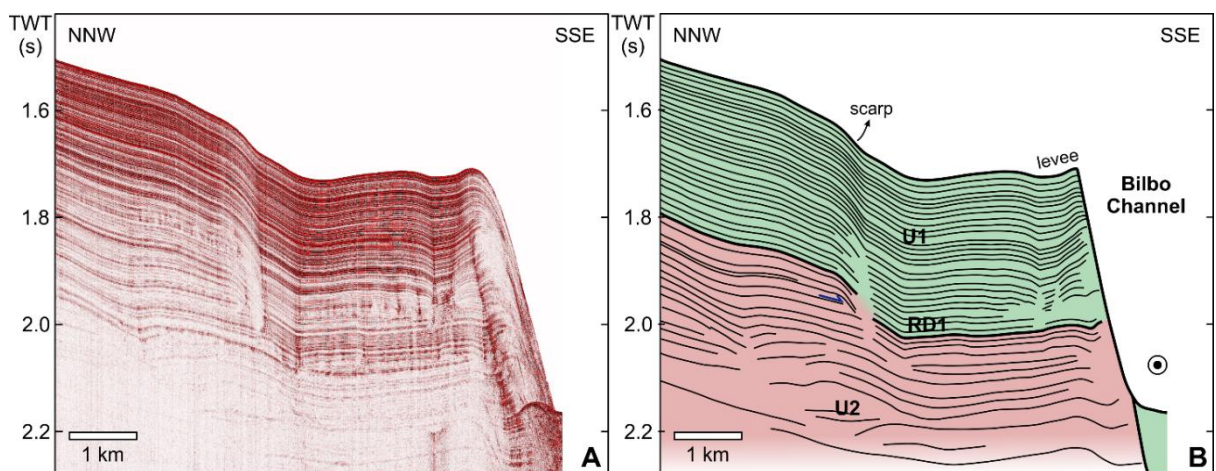


Figure 2.8. (A) Seismic profile (location: see Figure 2.1B) crossing Bilbo Channel in the GCS with (B) interpretation indicating the seismic stratigraphic units and their boundaries as described in the text. The blue arrow in the middle of the figure indicates truncation by an internal reflection.

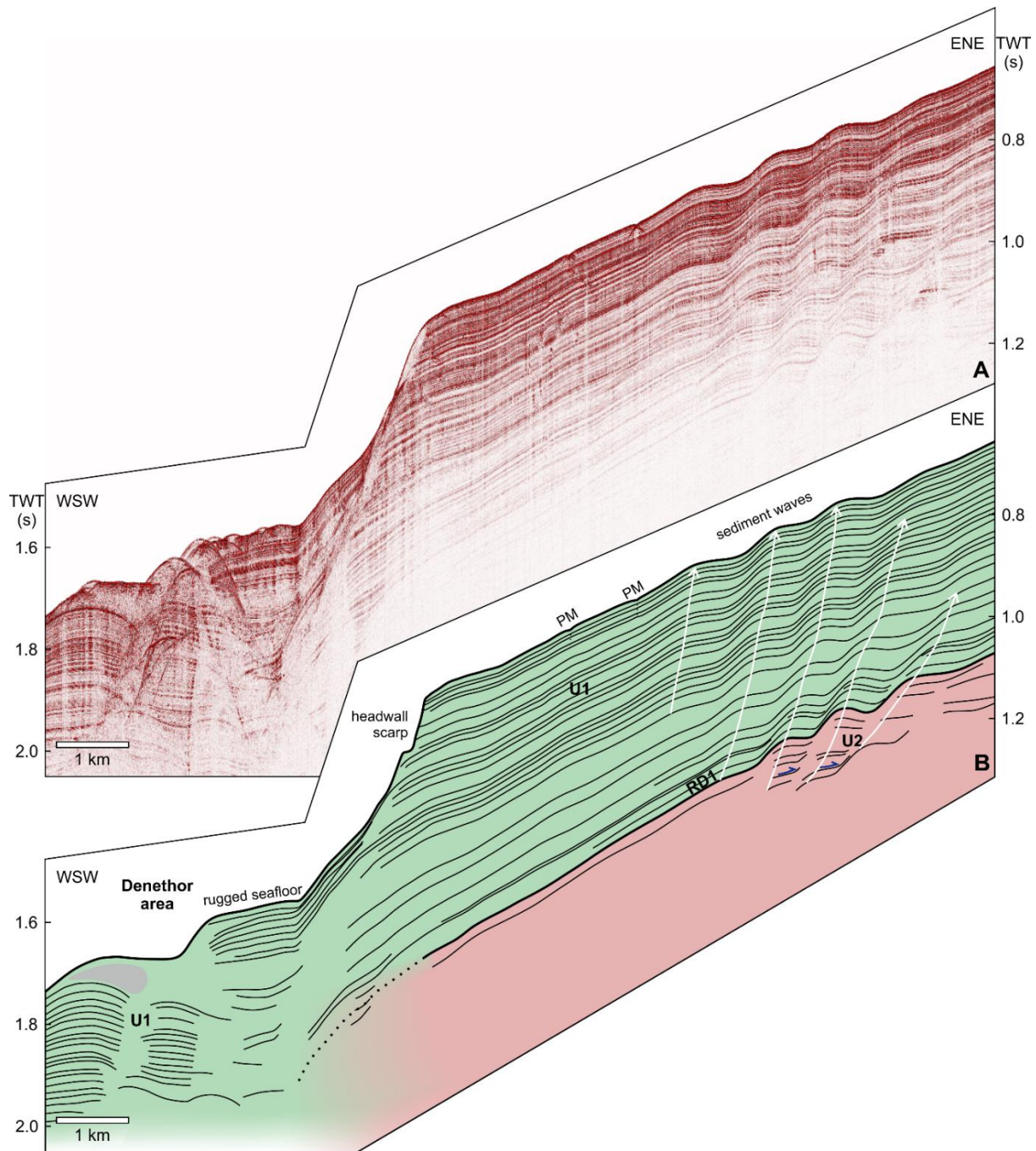


Figure 2.9. (A) Seismic profile (location: see Figure 2.1B) crossing the Denethor area and the sediment wave field north of Theoden Channel in the KCS with (B) interpretation indicating the seismic stratigraphic units and their boundaries as described in the text. PM indicates the locations of pockmarks recognised on the profile. The grey shaded area at the base of the slope indicates a zone with transparent seismic facies. The white arrows indicate the migration of sediment wave crests. The blue arrows in U2 indicate onlapping.

Unit U1 on the slope north of Theoden Channel is not only characterised by the undulating reflections, but also by the truncation of the uppermost 120 ms TWT (100 m) of reflections west of the undulations. This truncation is situated at the location of the largest scarp of the Denethor area (Figure 2.1C), which corresponds to the headwall of a submarine landslide. Underneath the truncated package, the reflections continue to the west and run parallel to the steep slope (Figure 2.9). West of the large scarp, at the base of the slope, most of the terrace-like features in the Denethor area have an internal structure that is well-stratified, with reflections that are (sub)parallel and somewhat wavy or (slightly) dipping downslope, although some chaotic or

transparent intervals are present as well (Figure 2.9). The continuity of the reflections from one terrace to another is interrupted by diffraction hyperbolas, which are linked to the scarps and incisions in the bathymetry that create a rugged seafloor.

2.4.2.4.1 U1 characteristics in channel infill

In the channels of the KCS, part of the channel infill consists of reflections that are continuous from the channel fill to the channel flanks, where they drape the RD1 unconformity. In the GCS, on the other hand, U1 reflections are truncated at the channel edge and do not form part of the channel infill.

The channel infills in Aragorn and Theoden channels are about 250 ms TWT (200 m) and 200-300 ms TWT (160-240 m) thick, respectively. The infills consist mainly of subparallel layered, slightly undulating reflections, with chaotic and transparent sections in between (Figure 2.4, 2.5). Underneath, the base of the channel consists of the typical wavy and/or parallel reflections of relatively high amplitude of unit U3. The mounded features in both channels contain mostly continuous, parallel, high-amplitude reflections that dip towards the basin and have a convex-up shape in an across-slope profile, while at their bases a transparent to incoherent facies is present (Figure 2.4, 2.5).

In Bilbo and Frodo channels, the infills are about 200 ms TWT (160 m) and maximum 300 ms TWT (240 m) thick, respectively. A high-amplitude reflection presumably indicates the bases of the channels, though it is not always clearly defined on the seismic data. The channel infills are composed of a chaotic to transparent facies including a couple of discontinuous, high-amplitude reflections (Figure 2.7).

2.4.3 Hydrography

The time series in Figure 2.10A shows that the daily mean current in the recorded period never exceeded 14 cm/s. The maximum instantaneous current velocity measured 53.7 cm/s, but the average measured current speed was 15.1 cm/s and 70 % of the measurements were lower than 20 cm/s (Figure 2.10B, D). The overall residual current in the channel from June to December 2005 at 8 m above the channel floor was oriented towards 324°N ($\approx$ NW) with a velocity of 4.75 cm/s (Figure 2.10C). Tidal analysis indicates that the M_2 component (principal lunar semidiurnal; 12.42 hour period) dominates. The major axis of the M_2 tidal ellipse has an amplitude of 17.1 cm/s and is oriented along 100-280°N. The minor axis has an amplitude of 0.01 cm/s, hence the M_2 -tide is strongly rectilinear along the channel axis (Figure 2.10C). The tidal ellipses of the S_2 - and the K_1 -tides are both oriented along the channel axis as well (100-280°N and 90-270°N, respectively). The ellipse of the S_2 -tide (principal solar semi-diurnal; 12 hour period) has a major axis with an amplitude of 9.2 cm/s (minor axis: -0.2 cm/s), while the K_1 (lunar diurnal; 23.93 hour period) tidal ellipse has a major axis with an amplitude of 1.0 cm/s (minor axis: -0.2 cm/s). Spring-neap cycles are also clearly visible in the data (Figure 2.10E).

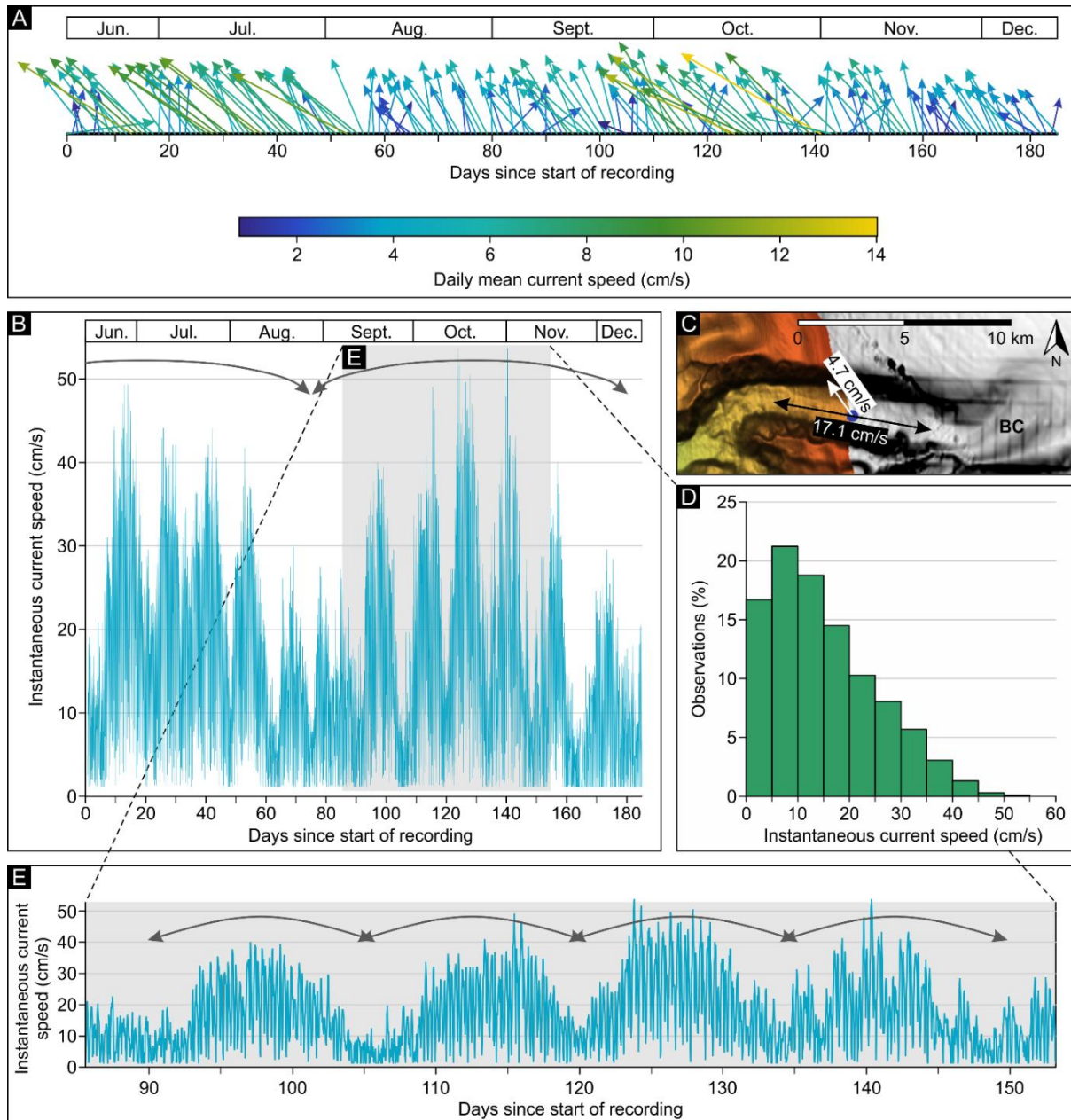


Figure 2.10. Data from the mooring in Bilbo Channel (location: Figure 2.1B) spanning the period from mid-June to mid-December 2005. (A) Time series of the daily mean current speed. (B) Time series of the instantaneous current speed. The grey arrows indicate long-term cyclicity. (C) Close-up shaded bathymetry map of the uppermost section of Bilbo Channel (BC; colour scale and location: Figure 2.1B, D), with indication of the location of the mooring (blue circle), the residual current vector (white arrow), and the major axis of the M2 tidal ellipse (black arrow). (D) Histogram of the instantaneous current speed measurements. (E) Close-up time series of the instantaneous current speed. The grey arrows indicate spring-neap cyclicity.

As the mooring results indicate that M_2 is the dominant tide in the study area, the internal wave frequency σ is set to 0.081 cph in the calculation of the γ/c -ratio that is used to estimate the interaction of internal waves with the seabed (Cacchione et al., 2002). Results from this calculation indicate that principal critical conditions ($\gamma/c \approx 1$) occur between 600 and 800 m water depth and more sporadically between 1000 and 1200 m water depth (Figure 2.11A). In waters shallower than 600 m, subcritical conditions ($\gamma/c < 1$) prevail, while between 800 and 1000 m water depth, supercritical conditions are dominant ($\gamma/c > 1$).

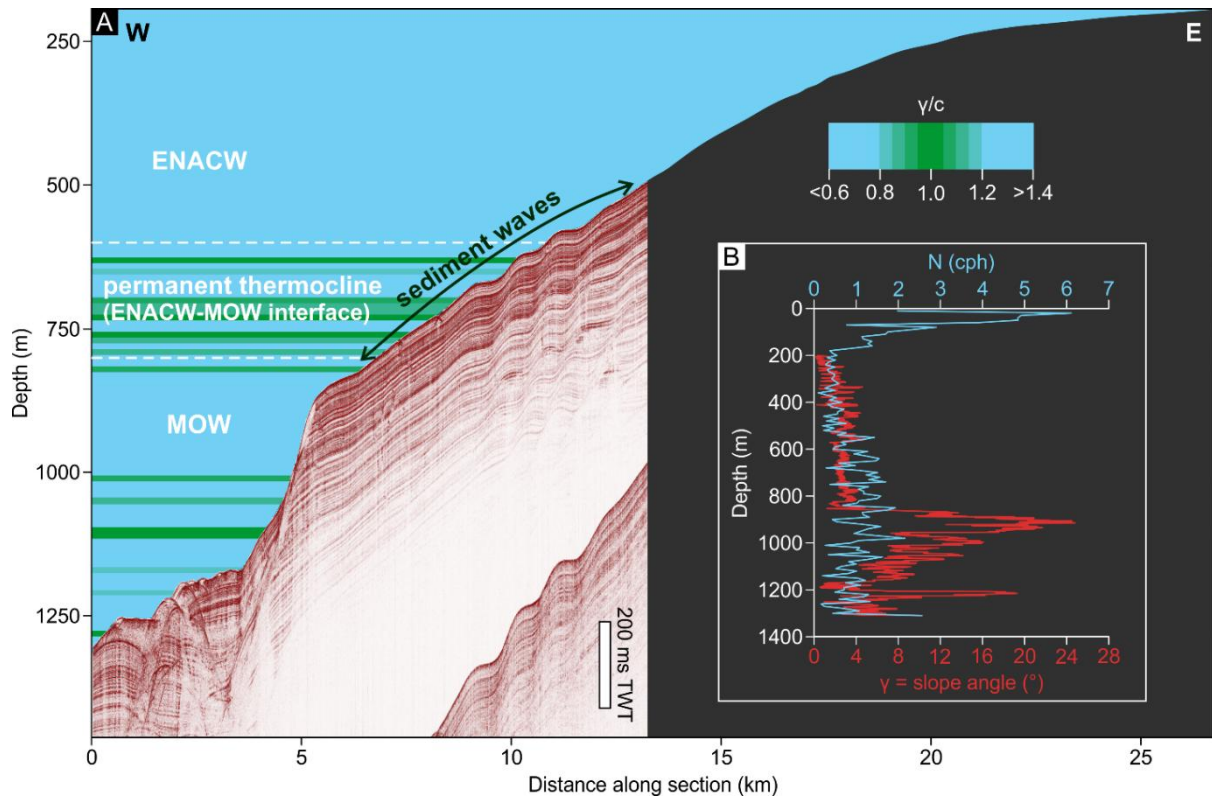


Figure 2.11. (A) Ratio of the slope angle (γ) to the angle of energy propagation of internal waves (c) over the present-day seabed in the sediment wave field at the location of the seismic profile shown in Figure 2.9 (location: see Figure 2.1B). The extent of the sediment wave field in the study area is indicated by the arrow. The permanent thermocline, roughly coinciding with the interface between ENACW (= Eastern North Atlantic Central Water) and MOW (= Mediterranean Outflow Water), is indicated by a white dashed line. (B) Graphs of the Brunt-Väisälä frequency (blue) and the slope angle (red) versus depth from the WOD18 (Boyer et al., 2018) CTD station between the KCS and GCS (location: see Figure 2.1B) and from the bathymetry profile in (A), respectively.

2.5 Discussion

2.5.1 Chronostratigraphic framework

Connecting seismic lines (e.g., the profile shown in Figure 2.4) allow the seismic stratigraphy in the KCS and GCS to be linked to the seismic stratigraphy established in the BMP and Enya CWC mound area (Van Rooij et al., 2007b, 2009). Units U1, U2 and U3, and regional discontinuities RD1 and RD2 described here are the southward continuations of units U1, U2 and U3, and regional discontinuities RD1 and RD2 (respectively) as described in Van Rooij et al. (2007b, 2009). Local discontinuity LD1a is equivalent to RD1b in Van Rooij et al. (2009). Unit U2bis of Van Rooij et al. (2009) in the area of the Enya CWC mounds is here divided into two local subunits SUa and SUB, as local discontinuity LD1b, not recognised before, exists between them. Van Rooij et al. (2007b, 2009) linked the seismic stratigraphy of the Belgica and Enya CWC mounds to the litho- and chronostratigraphy of (the slope surrounding) Challenger Mound. These were determined during IODP Exp. 307 and described by Expedition Scientists (2005), Kano et al. (2007), and Louwye et al. (2008), and are used here to build the chronostratigraphic framework for the upper slope of the KCS and GCS.

Unit U3 is the lowermost unit visible on the seismic profiles. It is present at the base of some of the channels (Figure 2.5) and in the channel flanks (Figure 2.4). The lower boundary of this unit is not visible on any of the seismic profiles in the study area but has been recognised (without age determination) in the BMP (Van Rooij et al., 2007a, 2007b). The sediments within unit U3 are thought to be of early to middle Miocene age (Van Rooij et al., 2003). The upper boundary of unit

U3 is an unconformity created by the RD2 erosional event (Van Rooij et al., 2003), which caused a minor hiatus that was attributed an early middle Miocene age (Louwye et al., 2008).

Unit U2 is bounded at the base by the RD2 unconformity and at the top by the RD1 erosional unconformity, which represents a large hiatus between upper Miocene and upper Pliocene sediments (Kano et al., 2007; Louwye et al., 2008; Van Rooij et al., 2009). The youngest dated sediments in unit U2 were deposited 8.96 Ma (Kano et al., 2007). However, the RD1 event eroded the sediments of unit U2 with varying intensity (Kano et al., 2007; Louwye et al., 2008; Huvenne et al., 2009b), which is also indicated by the lateral thickness variations in the unit (Figure 2.4, 2.5). As a result, seismic unit U2 in the study area is thought to include sediments of middle to late Miocene age, although deposition might have continued longer as it is unknown how much sediment was removed during the RD1 event.

Van Rooij et al. (2009) found that the RD1 event was locally preceded by another phase of erosion and in the data presented here, an additional very local erosion surface is recognised. These three unconformities are only distinguished with confidence north of Aragorn Channel (Figure 2.4, 2.6) and they are indicated as LD1a (oldest, local phase), LD1b (younger, very local phase), and RD1 (youngest, extensive main phase). The LD1a unconformity has been attributed to an early erosional phase within a composite event, possibly masked further south by the more severe RD1 phase (Van Rooij et al., 2009). The same reasoning can be applied to explain the formation of the LD1b unconformity. Units SUa, present between LD1a and LD1b/RD1, and SUB, present between LD1b and RD1, could then consist of the deposits of these early phases within the composite RD1 event (Van Rooij et al., 2009), and thus be between late Miocene and late Pliocene in age. It should be noted, however, that this interpretation was based on the position of LD1a within the seismic stratigraphy and on a composite event with a possible earlier onset creating a better chronological fit between RD1 and regional erosion events (Van Rooij et al., 2009). These erosion surfaces might, therefore, not have any geological relation to the RD1 event, and simply be independent local events.

Sedimentation resumed asynchronously after the RD1 event, thus adding to the lateral variation in the extent of the RD1 hiatus and causing a variation in the age of the base of unit U1. Deposition after RD1 started 2.7 Ma with the onset of CWC mound growth in the BMP (Kano et al., 2007; Huvenne et al., 2009b; Thierens et al., 2013). The CWC framework allowed baffling of sediments probably from around 2.6 Ma (Foubert and Henriët, 2009; Thierens et al., 2010). In locations where no CWC framework was present to promote the entrapment of grains, currents were periodically strong enough to winnow any sediments deposited during periods of weaker current regimes (Huvenne et al., 2009b; Mienis et al., 2009; Thierens et al., 2010). This is exemplified in an off-mound site of IODP Exp. 307, where prolonged sedimentation after the RD1 event did not resume before 1.7 Ma (Kano et al., 2007; Huvenne et al., 2009b). As the youngest dated sediments underneath RD1 were assigned an age of 8.96 Ma, the length of the RD1 hiatus based on IODP Exp. 307 data is 6-7 Myr (Kano et al., 2007; Huvenne et al., 2009b; Thierens et al., 2010). Unit U1 in the study area is therefore suggested to consist of sediments of middle Early Pleistocene and younger ages.

2.5.2 Evolution and regional importance of the Kings and Gollum channel systems

2.5.2.1 Early and Middle Miocene: pre-channel slope sedimentation

Seismic unit U3 mainly consists of siltstone (Expedition Scientists, 2005). Its seismic signature of predominantly subparallel, downslope dipping (3.5-4.5°) reflectors represents relatively calm paleoceanographic conditions with hemipelagic sedimentation during the early Miocene (Van Rooij et al., 2003; Louwye et al., 2008). The wavy reflectors that are present locally (Figure 2.6), however, might represent sediment waves, which would be indicative of the presence of increased bottom currents (Van Rooij et al., 2003; Wynn and Masson, 2008). These structures

have also been observed within U3 in upslope areas in the BMP (Van Rooij et al., 2003, 2007a), demonstrating that these more dynamic conditions in the early Miocene were more widespread than they might seem from the data shown here.

Seismic unit U2 is more acoustically transparent than unit U3, but also contains continuous reflections that are gently dipping towards the basin and are spaced more closely towards the southern end of the study area. IODP Exp. 307 data show that the sediments within U2 are mostly homogenous silty clays, especially at the bottom of the unit, though the clays are more interbedded with clayey silts and fine sands towards the top of the unit (Expedition Scientists, 2005). These factors point towards another period where hemipelagic sedimentation was the dominant process of deposition. Unit U2 also contains sediment waves north of Theoden Channel that are the subsurface expression of the undulations visible on the present-day seafloor (Figure 2.9).

2.5.2.2 Late Miocene-Late Pliocene: incipient channel systems

From similar seismic signatures of unit U2 north and south of individual channels (especially Aragorn and Bilbo channels; Figure 2.4, 2.7), it is clear that the bases of the KCS and GCS are cut into the sediments of unit U2, which were originally deposited more uniformly across the margin. Additionally, some reflectors within U1 are continuous from the channel flanks to the channel axis (Figure 2.4, 2.5), suggesting they were deposited across the slope after the initial incision of the channels. In the KCS, the RD1 event appears to have completely eroded unit U2 at the location of the channel bases, and may have partially eroded unit U3 (Figure 2.4, 2.5). From these findings, it seems plausible that the extensive late Miocene-late Pliocene RD1 erosional event was responsible for initially shaping the KCS. It is likely that the RD1 event also provided the initial seafloor topography on which the GCS was built, as U2 reflections are truncated at the channel edges. However, the channel bases of the two northernmost channels of the GCS are situated at greater depths than those of the KCS, making it more difficult to determine the exact relation between the channels and the RD1 event from the seismic profiles (Figure 2.7). Seismic profiles with deeper penetration are needed to clarify this relationship.

It has been proposed that the RD1 event was a composite event consisting of at least two separate phases (Van Rooij et al., 2009). The data presented here suggest that RD1 is the composite of at least three erosional phases that alternated with periods where local deposition of sediments was possible. Those depositional periods are represented by units SUa and SUB, which both show similarities with highly energetic contourite drift deposits (Stow et al., 2002; Van Rooij et al., 2009); the complex seismic signature with sigmoidal to wavy reflectors of unit SUa indicates a much more dynamic sedimentary environment during their deposition than for the sediments in unit U2 underneath. Within unit SUB, reflections are dipping more gently and are not as wavy as in SUa, but they do display onlap relationships with the LD1b unconformity, suggesting a relatively calmer though still current-influenced depositional environment. This illustrates the clear influence of bottom currents of varying intensities on shaping the initial seafloor topography where the channels developed. It also confirms the presence of high-energy northerly bottom currents in the late Miocene-late Pliocene period as suggested by Van Rooij et al. (2003) in the BMP. In contrast to the RD1 erosional event in the BMP, however, which is caused by bottom current action only (Van Rooij et al., 2003), the RD1 erosion in the study area is due to the combined effect of intensified bottom currents interrupted or followed by downslope flows finally creating the bases of the KCS and GCS. These erosive downslope flows originate from the Irish Shelf, where lowered sea levels in glacial periods generate more proximal sediment sources (Kenyon et al., 1978; Weaver et al., 2000). This accentuates the link between the RD1 event and the onset of glacial-interglacial cycles in the North Atlantic (Van Rooij et al., 2003). It also confirms that these channel systems could contain valuable information on BIC dynamics at its southwestern limit, though further research will be needed to validate this and extract this information.

2.5.2.3 Pleistocene – Recent: channel infill and flank build-up

2.5.2.3.1 Turbidity currents with spatially and temporally varying intensities

The importance of turbidity currents in shaping the channels is demonstrated by multiple fine-grained turbidites found in cores in channel thalwegs and flanks (Kenyon et al., 1998), as well as on the continental rise at the western end of the system (Akhmetzhanov et al., 2003).

In the KCS, several indications exist for turbidity currents changing in intensity throughout the Quaternary. Distinct discontinuities within unit U1 on the flanks of Theoden Channel point to periods of erosion causing repositioning or reshaping (broadening) of the channel alternating with periods of deposition causing channel infilling (Figure 2.5). Furthermore, the three mounds of sediment within Aragorn and Theoden channels (Figure 2.1C, 2.4, 2.5) are interpreted as internal levees (*sensu* Callow et al. (2014)) based on their elongated shape parallel to the channel axis, their location within the channel, and the (sub)parallel, downslope-dipping seismic reflectors they are built from. The presence of internal levees suggests that the Kings channels evolved in such a way that they became capable of accommodating downslope flows. Larger channel dimensions were possibly combined with weaker and thinner flows resulting in the deposition of layered sediments in internal rather than on external levees (Deptuck et al., 2003; Callow et al., 2014). At the base of Theoden Channel, another set of layered reflections possibly hints at the presence of a buried internal levee (Figure 2.5). It could indicate that, although erosional processes were dominant in the very beginning of KCS development creating their U-shaped bases, depositional processes took over soon after. Additionally, it could indicate lateral migration of the channel thalweg, although additional seismic data, and data reaching deeper into the subsurface, are needed to confirm this.

After the initial predominantly erosional to bypassing turbidity current action in the GCS, evidenced by V-shaped cross-sections, turbidity currents became bypassing to slightly depositional and started to fill the channels, creating more U-shaped cross-sections. In the lower section of unit U1 on the channel flanks, divergence of reflectors and a consequential increase in thickness of the unit towards the channels indicate the build-up of levees in very close (1 km) proximity to the channels (Figure 2.7, 2.8). This suggests that turbiditic spillover played a role in shaping the channel flanks. However, the limited extent of the levees both northwards and westwards indicates that spilling was minimal (Mulder, 2011) and concentrated in the upslope section of the channel. The levee geometry diminishes towards the top of unit U1, where sediments are mostly draped onto the already existing topography, with reflections that are conformable, laterally continuous and have very limited thickness variations over large distances (Figure 2.7, 2.8). This suggests a subsequent dominance of hemipelagic processes in the area, which is in accordance with hemipelagic muds devoid of turbidites found in cores on the flank of Frodo Channel (Kenyon et al., 1998). These observations indicate that the sediment load of any turbidity current flowing through the channels more recently than the levee deposition was completely confined within the channel (Weaver et al., 2000). This could be due to either higher channel flanks relative to the thickness of the flow, very infrequent occurrence of turbidity currents, an insufficient amount of suspended sediment within them, or a combination of these (Babonneau et al., 2002; Deptuck et al., 2003, 2007). The transition from a system dominated by turbidity currents to one where hemipelagic sedimentation prevails could be linked to a disconnection from the sediment source and accordingly a diminished sediment supply from the shelf during interglacials (Mulder, 2011). The combination of depositional (levees) and erosional characteristics (steep flanks with numerous slope failure scars; Figure 2.1D), as well as the transition from V- to U-shaped cross-sections, allows for interpreting this upper slope section of the GCS as a canyon-channel transition (Kenyon et al., 1978; Mulder, 2011; Puig et al., 2014).

Whereas the KCS has been dominated by depositional processes interrupted by some erosional events, the turbidity currents in the Frodo and Bilbo channels in the GCS seem to have been predominantly erosional, with some deposition in outer levees (Figure 2.4, 2.5, 2.7). In the GCS,

channel fills are dominated by transparent facies typical for mass-transport deposits with sparse and discontinuous higher-amplitude reflections deposited from coarse-grained turbidity currents (Deptuck et al., 2003; Mulder, 2011). This is in contrast to the channel fills in the KCS, which are dominated by a facies with continuous, subparallel and sometimes undulating reflections deposited from more fine-grained flows running through the channels, intercalated with the transparent facies from mass-transport deposits (Figure 2.4, 2.5). Generally, a more depositional environment is linked to slower currents allowing suspended sediments to settle, while the more erosional environment in the GCS might suggest overall more intense currents that inhibit deposition (Babonneau et al., 2002; Mulder, 2011). Intense currents combined with the strongly incised morphology of the GCS likely prevented large spillover, which may have led to a preservation of the energy of the turbidity currents, allowing them to continue all the way to the abyssal plain. In the KCS, on the other hand, more deposition from currents may have caused more energy loss and currents only reaching the base of the upper slope. This process has been used by Babonneau et al. (2002) to explain morphological differences between the Zaire and Amazon canyon-channel systems. As there is no significant difference in thalweg slope gradient between the two channel systems (about 2° in both systems), the difference in downslope current intensity inferred here must be explained otherwise. The channels of the KCS seem to be oriented towards the southwestern tip of Ireland (Figure 2.1A), possibly indicating more direct access to terrestrial sediments, while the GCS channels are not oriented towards Ireland and may have been fed by material from the Irish Shelf. This possibly led to a difference in the amount and/or nature of sediment supplied to both systems. The higher current intensity and longer extent of the GCS seem to imply a higher sediment load in this system, though the existence of dilute flows causing significant erosion and little channel infill has been suggested as well (Babonneau et al., 2002). If the former hypothesis is true and large amounts of sediment were carried to the abyssal plain, currents there must have been strong enough to either reroute sediment elsewhere, or rework the sediments after deposition, and prevent a significant deep-sea fan from developing. The uncertainty about the nature and amount of sediment supply from the BHS to the channel systems forces this reasoning to remain speculative.

2.5.2.3.2 Significance of bottom currents in a turbidity current-dominated system

Although mass-wasting and turbiditic deposits dominate the channel fills and the lower part of unit U1 on the channel flanks in close proximity to the channel axes, and hemipelagic deposits dominate the rest of the areas in between the channels, several sections of unit U1 are clearly influenced by bottom currents. One such area can be found buried north of Aragorn Channel (Figure 2.6). Another is the area of seafloor and connected subsurface undulations north of Theoden Channel (Figure 2.9). Gravity processes deforming sediments are one of the mechanisms capable of shaping the seafloor into a “wavy” morphology (Faugères et al., 2002; Lee et al., 2002). However, in the study area, the seismic data show no evidence for sediment deformation in or near the undulating features, which are therefore interpreted as a sediment wave field formed by bottom currents. These sediment waves indicate the movement of sediment over this slope area during the late Miocene and throughout the Quaternary. As they are visible on the seabed (and thus not buried by hemipelagic sedimentation), whichever process is responsible for their build-up has likely been active until recently (Faugères et al., 2002).

Several mechanisms are known to form sediment waves: (1) deposition from alongslope-flowing bottom currents (Masson et al., 2002; McCave, 2017), (2) deposition from downslope flowing turbidity currents (Migeon et al., 2000; Normark et al., 2002), or (3) deposition from across-slope oscillating internal waves (Ribó et al., 2016; Collart et al., 2018; Mestdagh et al., 2020). A combination of these processes has been suggested as well (Faugères et al., 2002). The orientation of the wave crests roughly perpendicular to the channel axes and parallel to the isobaths (170-175°N; Figure 2.1C) precludes an origin from alongslope currents or turbiditic spillover. Sediment waves formed by sheet-like unconfined turbidity currents usually display a

trend of decreasing dimensions downslope, are often linked to upslope slump and/or debris flow deposits, and are thought to require regular flow of turbidity currents over the area (Wynn et al., 2000; Ercilla et al., 2002; Kolla et al., 2021). Neither downslope decreasing dimensions nor associated gravity flow deposits are visible here (Figure 2.9). Additionally, turbidity currents in the study area are thought to have been too intermittent (i.e., glacial-interglacial variability; Kenyon et al., 1998; Akhmetzhanov et al., 2003) to have formed such regular and vertically extensive sediment waves. However, the contribution of unconfined turbidity currents to the formation of the sediment wave field cannot be excluded completely and requires further research.

The sediment waves occur on the seafloor at 500-800 m water depth (Figure 2.1C), a range which includes the interface between the ENACW and the underlying MOW (600-800; Figure 2.2) and is roughly the depth of the permanent thermocline (600-900 m; White and Dorschel, 2010). The permanent thermocline is associated with increased baroclinic (internal) energy at tidal periods because of strong vertical density stratification there (Pomar et al., 2012; Shanmugam, 2013). Both semi-diurnal (freely propagating) and diurnal (bottom-trapped) baroclinic waves occur and enhance bottom turbulence (White and Dorschel, 2010; Pomar et al., 2012; Shanmugam, 2013), making both types plausible factors in the sediment wave generation.

The sediment waves are present where an area with little evidence of diurnal waves (at mooring 112; Figure 2.1B) transitions to an area with strong diurnal waves (BMP). In the BMP, the resulting strong diurnal bottom currents were likely crucial for coral mound growth (White, 2007; White and Dorschel, 2010; Wienberg et al., 2020). Some modest enhancement of the diurnal tide might be expected in the study area, hence potentially the sediment waves might result from a localised region of bottom-trapped wave enhancement dependent on stratification, bottom slope and orientation of the barotropic diurnal tide.

Internal tides of semi-diurnal period will freely propagate within a stratified fluid at latitudes $< 75^\circ$ (Morozov and Pisarev, 2002). At the ENACW-MOW interface ($\approx$ permanent thermocline here), freely propagating internal tides have been predicted for a large part of the eastern slope of the Porcupine Seabight (Rice et al., 1990; White, 2007). The interaction of internal waves with a continental slope differs depending on the ratio of the seafloor slope angle γ to the incidence angle of the internal wave characteristics c : the waves are reflected upslope (transmissive conditions) when the slope is subcritical ($\gamma/c < 1$), reflected downslope (reflective conditions) when the slope is supercritical ($\gamma/c > 1$), or reflected parallel to the bottom when the slopes match ($\gamma/c \approx 1$) (Cacchione et al., 2002). In the latter case of a critical slope, the vertical wavelength approaches zero and the wave energy density is significantly enhanced (Lamb, 2014). This leads to increased bottom current velocities and shear stress, which may cause resuspension and transport of sediments, and induce the formation of sediment waves (Cacchione et al., 2002; Lamb, 2014; Ribó et al., 2016). Calculations of the γ/c -ratio using CTD data close to the sediment wave field (Figure 2.1B) indicate that criticality conditions for the dominant M_2 -tide occur within the depth range where the sediment waves are present, near the present-day permanent thermocline between 600 and 800 m (Figure 2.11).

Given the proximity of the mooring 112 location (Pingree and Le Cann, 1990), freely propagating internal waves along the ENACW-MOW interface rather than bottom-trapped baroclinic diurnal tides are suggested as the most probable mechanism for the generation of the sediment waves on the seafloor north of Theoden Channel. Current measurements in the sediment wave field would help validate this hypothesis. Additionally, further research with deeper seismic data is needed to explain the occurrence of the sediment waves underneath RD1 (before the MOW existed; Hernández-Molina et al., 2014). Prior to MOW presence, a permanent thermocline would still have been present, so it is possible that internal waves existed in the depth range of the present-day ENACW-MOW boundary. Furthermore, there is evidence for a vertical shift of the permanent thermocline region over time (Raddatz et al., 2014; Wienberg et al., 2020), hence a

deeper thermocline depth (with enhanced baroclinic energy) would be possible prior to MOW presence. A lack of knowledge on the paleoceanography in the study area before the RD1 event hinders verifying this possibility.

In seismic unit U2 and the lower part of unit U1, a slight offset of reflectors between two adjacent sediment waves points to erosional conditions on the downslope flanks in the period before and just after the RD1 event (Symons et al., 2016). Towards the top of unit U1, the reflectors become laterally continuous (and thus completely depositional) and the configuration of the sediment waves evolves from strongly upslope migrational into more aggradational (Figure 2.9). This suggests a lessened influence of bottom current activity, possibly combined with relatively lower sedimentation rates (Flood, 1988; Wynn and Masson, 2008; Faugères and Mulder, 2011).

A third area that has likely been influenced by bottom currents is the Denethor area, though further research is needed to better understand its formational history. The parallel-layered reflections that are dominant in the internal structure of the individual terrace-like features (Figure 2.9) indicate a possible combined influence of downslope-flowing turbidity currents and alongslope-flowing bottom currents. Sediment brought into the basin by turbidity currents was likely subjected to pirating by northwards flowing contour currents towards the BMP along the steep slope-parallel escarpment in the Denethor area and the alongslope channel just north of the KCS (Figure 2.1B) (Van Rooij et al., 2003). Thus, as they were the conduits for sediment transfer from the Irish Shelf to the Porcupine basin, the channels played an important role in the development of the Belgica CWC mounds and the associated sediment drifts.

Lastly, the slight asymmetry of the Gollum channels (with steeper northern flanks) may be linked to the action of northward-flowing bottom currents. The northern channel flanks create an obstacle for the currents, which causes intensification and erosion, whereas the southern flanks are predominantly areas of deposition (Figure 2.7, 2.8; Marchès et al., 2007; De Mol et al., 2011).

2.5.2.3.3 Slope failures and their origin

Apart from down-, across- and alongslope currents, the channels in the KCS and GCS are also shaped by numerous slope failure events (Figure 2.1C, 1D). Undercutting of the channel walls by erosional downslope flows has been suggested as a major cause of slope failures in this setting (Micallef et al., 2012; Watson et al., 2020). Additionally, the presence of pockmark fields in the KCS (Figure 2.1E) indicates the possible importance of excess pore pressure in the sediments for slope destabilisation (King and MacLean, 1970; Hovland and Judd, 1988; Hovland et al., 2002; Judd and Hovland, 2009), especially for the slope failures on the southern flank of Theoden Channel and at the southeastern edge of the Denethor area (Figure 2.1C). The build-up of excess pore pressure might be linked to elevated sedimentation rates (Judd and Hovland, 2009; Moernaut et al., 2017), which are known to have occurred in the study area in the early period of deglaciation after the LGM (Van Rooij et al., 2007a, 2007b). The long scarp north of Bilbo is linked to a large discontinuity of the reflectors in the subsurface where a package of reflectors is truncated (Figure 2.8), and is therefore suggested to be the seabed expression of a buried landslide headwall.

2.5.3 Present-day in-channel oceanographic processes: tidal forcing and turbidity current action

The presence of the sediment wave field north of Theoden Channel and the general morphology of the study area, including smoothed channel flanks and smooth interfluves (Figure 2.1B, 2.4, 2.5, 2.7), demonstrate the present-day influence of bottom currents on seafloor morphology between the channels (Wheeler et al., 2003; White, 2007). Inside the channels, Tudhope and Scoffin (1995) inferred reversing currents from channel floor ripples and estimated current velocities of 0.5 m/s. Measurements from the mooring in Bilbo Channel indicate a dominance of tidal forcing on the currents (Figure 2.10E) up to a maximum of ~0.5 m/s comparable to the

estimates of Tudhope and Scoffin (1995). Currents are dominated by the M_2 -tide, strongly steered along the channel axis (Figure 2.10C) and spring-neap as well as longer period cycles clearly visible in instantaneous current measurements (Figure 2.10B, E). The currents measured inside the channel are more intense than those on the interfluvium between the KCS and GCS (mooring 112; Figure 2.1B) (Pingree and Le Cann, 1990; White, 2007), further demonstrating the significant influence of the topographically steered tidal motions in the channel. The overall residual current vector (Figure 2.10C) and the daily mean current vectors (Figure 2.10A) suggest the presence of a cross-channel component in the currents. This may be due to highly localised steering from an irregular channel floor topography, although this cannot be confirmed as the mooring is situated just outside the area covered by Beyer et al. (2003) bathymetry and detailed channel floor morphology is not available.

The instantaneous current measurements of the Bilbo mooring show no major (downslope) flow episodes and the mooring remained in the same position during the measurement period. This suggests that no major turbidity currents occurred in Bilbo Channel during that time and reinforces the idea that the channels are currently inactive with respect to sediment gravity flows. However, the possibility remains that turbidity currents occur too infrequently to have been detected by this mooring.

2.5.4 Comparison with Whittard Canyon on the Celtic Margin

Whittard Canyon is the most westerly of the canyon systems on the northern continental slope of the Bay of Biscay (Figure 2.1A). As the GCS, Whittard Canyon is a dendritic system that has been affected by turbidity current activity (Toucanne et al., 2008), and experiences some influence from tides causing present-day up-canyon sediment transport (Amaro et al., 2015) and from contour currents smoothing canyon flanks (Stewart et al., 2014). Canyon incision in the Whittard system started in the Pliocene-Pleistocene and is linked to lowered sea levels (Bourillet et al., 2003), similar to what is inferred here for the GCS. As the drainage system of the Grande Sole drainage basin, Whittard Canyon is thought to have drained the Celtic Sea during the last glacial period, receiving sediment from the BIIS (through the Irish Sea Ice Stream; Figure 2.1A) and feeding the Celtic Deep-Sea Fan (Scourse et al., 2000, 2009b; Zaragosi et al., 2000, 2006; Toucanne et al., 2008). The deglaciation of the most recent British-Irish and European ice sheets (ca. 20-13 ka), in particular, caused increased sediment influx to the Grande Sole drainage basin and consequently enhanced turbidity current activity in Whittard Canyon (Zaragosi et al., 2000; Bourillet et al., 2003; Toucanne et al., 2008). Terrigenous sediment input from glacio-isostatic uplift of the British Isles continued until 7 ka (Lambeck, 1996; Bourillet et al., 2003). Even though Whittard Canyon is now too far from any coastlines and hemipelagic sedimentation influenced by tides and waves dominates the Celtic Margin, sediment gravity flows reworking material on the outer shelf and possibly induced by storm depressions or anthropogenic activity are still active within the Whittard system, although less frequently than during the last deglaciation (Zaragosi et al., 2000; Bourillet et al., 2003; Cunningham et al., 2005; Amaro et al., 2015; Daly et al., 2018).

Even though the Quaternary activity within Whittard Canyon seemingly bears many similarities to the GCS, the geomorphology of the two systems differs. Whittard Canyon is much more like a canyon system *sensu* Shepard (1965) consisting of an extensive dendritic network of conduits with steep walls (commonly 30-40°) incised by numerous gullies, clearly V-shaped cross-sections and large relief, that are in some places cutting into bedrock (Cunningham et al., 2005; Stewart et al., 2014). The present-day V-shaped cross-sections indicate that Whittard Canyon has been governed by erosional processes (Cunningham et al., 2005) more so and until more recently than the GCS. This might be due to the large amounts of sediment that were brought to the Whittard Canyon from the BIIS (Toucanne et al., 2008). As yet, there is uncertainty about 1) the extent of the BIIS before the LGM on the shelf southwest of Ireland and thus the efficiency of the (repeated?) connection to the GCS and KCS (Sejrup et al., 2005), and 2) the proportion of sediment sourced from the BIIS that reached the eastern slope of the Porcupine Seabight

compared to the Celtic Margin. Therefore, it remains difficult to assess the influence of a potential difference in the amount of supplied sediment in the two systems. The absence of turbidity current activity in the GCS in the last 10 kyr (Wheeler et al., 2003) is probably due to a lack of sediment supply and/or external forces driving downslope flow (Mulder et al., 2012; Amaro et al., 2015). This is, however, not likely to be the most significant driver for the geomorphologic difference between the two systems, although very recent activity in Whittard Canyon does flush the canyons, depositing sediments further down the slope rather than in the channel axes (Amaro et al., 2015).

Another factor that might play a role is the difference in the average angle of the continental slope between the two areas: 2-5° on the eastern slope of the Porcupine Seabight versus an average of 8° on the Celtic Margin (Piper and Normark, 2009; Mulder et al., 2012). A higher slope gradient could explain the higher dendricity in the Whittard Canyon system (Harris and Whiteway, 2011), and could also lead to a higher velocity and therefore more erosional character of turbidity currents flowing through the system (Piper and Normark, 2009; Mulder, 2011).

The presence of fault scarps at the seafloor suggests a significant and active tectonic control on the morphology of the Whittard system (Cunningham et al., 2005). In the GCS, geological control on the orientation and morphology of the channels has been hypothesised (Wheeler et al., 2003). Lined-up features such as sudden, sharp (90°) changes in otherwise straight channels (Bilbo and Frodo; Figure 2.1D) could be an argument for that hypothesis, but it should be tested with deeper seismic data.

2.6 Conclusions

The first detailed seismic stratigraphy over the Kings and Gollum Channel Systems on the eastern side of the Porcupine Seabight is presented here. It is used to shed light on the processes involved in the evolution of the two systems, with a focus on their temporal variability, the interplay between them and their importance for regional sediment dynamics.

- The early to late Miocene pre-channel sedimentation is mainly hemipelagic, deposited in calm oceanographic conditions that were interrupted by a period with increased bottom current influence in the late early or early middle Miocene. In the early middle Miocene, erosion caused a small hiatus in the upper slope sedimentation. The margin-wide deposited pre-channel sediments were eroded during the extensive late Miocene-late Pliocene RD1 erosional event.
- The RD1 event is here proposed to have consisted of at least three phases of erosion. The first two phases are only recognised locally and resulted from highly energetic bottom currents that subsequently deposited the eroded sediments as high-energy drift deposits, shaping the initial seafloor topography where the channels developed. The youngest phase in the RD1 event was the most extensive, and, whereas it is attributed to intense bottom currents in the BMP, it is here associated with erosive downslope flows creating broader U-shaped incisions in the KCS and narrower V-shaped incisions in the GCS.
- The RD1 event has been linked to the onset of Quaternary glacial-interglacial cycles in the North Atlantic, and to the introduction of MOW into the NE Atlantic (Van Rooij et al., 2003). The combined effect of both processes (i.e., erosive downslope flows and intensified bottom currents) is visible in the multi-phased characteristics of the event in the KCS and GCS. This indicates that these channel systems could contain valuable information on BLS dynamics at its southwestern limit, but further research will need to confirm this.
- Quaternary sedimentation is notably influenced by both bottom and turbidity current activity. There are strong indications that internal tides are present at the ENACW-MOW interface in at least part of the study area. They are suggested to have occurred throughout the Quaternary, with a diminishing intensity towards the present day. Turbidity currents were most active in the beginning of the channel evolution. They were more

erosional in the GCS, while the KCS has been a predominantly depositional environment. They likely varied in intensity through time but eventually, a disconnection from sediment sources led to a dominance of hemipelagic sedimentation. There are no signs of recent turbidity current activity through the channels. The timing of the sediment source disconnection is uncertain. The upper slope sections of the KCS and GCS are nowadays a predominantly depositional environment where sedimentation is dominated by hemipelagic processes influenced by bottom current activity.

- Sediment carried downslope by turbidity currents was likely pirated and transported northwards towards the BMP by contour currents along slope-parallel escarpments and channels. Therefore, it is here proposed that the channels were of major importance as a source of sediment for the Belgica CWC mounds and the associated sediment drifts.
- The morphological difference between the channels on the eastern slope of the Porcupine Seabight and the Whittard Canyon is likely largely due to a greater sediment supply towards the Celtic Margin during glacial periods combined with a greater slope gradient and possibly an underlying tectonic control.

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CHAPTER 3 Margin processes sculpting a land-detached canyon-channel system: the Gollum Channel System in the Porcupine Seabight

This chapter has been slightly modified from the published version:

Verweirder, L., Van Rooij, D., and Georgiopoulou, A. (2023). Margin processes sculpting a land-detached canyon-channel system: the Gollum Channel System in the Porcupine Seabight. *Front Earth Sci* 11, 1285171. doi: 10.3389/feart.2023.1285171

Abstract

The land-detached Gollum Channel System (GCS) is one of very few large-scale canyon-channel systems on the Northwest European margin and thought to be of high importance in both alongslope and downslope sediment transport processes in the Porcupine Seabight basin. It is, however, unknown when this system was formed and how active it has been throughout its evolution, making it difficult to assess its regional impact. Here, using well data integrated with airgun seismic reflection data, a seismic stratigraphy (Cretaceous to present) is built for the GCS for the first time. We find that, contrary to what was thought before, the GCS was formed before Quaternary glaciations occupied the continental shelf and its origin is tentatively associated to a phase of Northeast Atlantic margin tilting in the early Pliocene. Each of the channels that make up the GCS was initiated by incision from erosive downslope gravity flows originating on the Celtic Sea Shelf. Gravity flows from Quaternary glacial processes reused the channels and mostly bypassed the upper slope or contributed to the channel fill, though some flows were capable of erosion of existing channel flanks and incision of several smaller channels. Additionally, we show that this margin was incised by erosive gravity flows on several occasions through time and that these incisions seem to follow preferential pathways. Interaction with alongslope bottom currents from the start of the Quaternary onwards was crucial to distribute sediments and nutrients to sediment drifts and cold-water coral mounds further north (downstream) along the Irish margin.

Keywords channel, canyon, seismic stratigraphy, bottom current, turbidity current

Author contributions All authors contributed to the conception and design of the study, L. Verweirder analysed and visualised the data and wrote the original draft. All authors contributed to the revision of the manuscript.

3.1 Introduction

Large-scale submarine canyon-channel systems, present on continental shelves and slopes across the world (Harris and Whiteway, 2011), are preferential pathways for the transport of sediment, organic matter and pollutants into the adjacent basins. They are often associated with high terrestrial sediment supply from, e.g., fluvial systems (Khripounoff et al., 2012; Babonneau et al., 2013; Puig et al., 2017) or ice sheets (Zaragosi et al., 2006; Rise et al., 2013). Additionally, several processes, such as trawling (Martín et al., 2014; Daly et al., 2018), dense water cascading (Canals et al., 2006; Puig et al., 2013), earthquakes (Mountjoy et al., 2014), tidal currents (Mulder et al., 2012; Amaro et al., 2016), and storm waves (Puig et al., 2004a) are capable of resuspending shelf and slope sediments and transporting them down (or up; Amaro et al., 2016) canyon axes. This ensures the role of canyon-channel systems in sediment transport even when they are not connected to terrestrial sources. Furthermore, these systems can have a considerable influence on local and regional hydrography (Chapter 2; Voigt et al., 2013; Waterhouse et al., 2017). Considering the significant role large-scale canyon-channel systems have in conducting and redistributing sediment, it is important to understand the processes that shape them.

The Gollum Channel System (GCS) is the only large-scale channel system on the Northwest European margin (Weaver et al., 2000) and stretches over c. 200 km connecting the Celtic Sea Shelf to the Porcupine Abyssal Plain (Figure 3.1). The system was named by Kenyon et al. (1978) and the names of the individual channels were given by Van Rooij (2004) (Figure 3.1B). The GCS presently has no apparent connection to any terrestrial sediment sources and is not thought to be active with downslope gravity flows (Akhmetzhanov et al., 2003; Wheeler et al., 2003). It did have a large influence on regional processes in the past as it is thought to have supplied sediments to the cold-water coral mounds and contourite drifts in the Belgica Mound Province (BMP; Chapter 2; Van Rooij et al., 2007b, 2009). It is, however, unknown when the system was formed and thus when its regional influence started, as well as which processes were involved in creating it. This is in stark contrast to the Celtic and Armorican systems in the Bay of Biscay, for which the history of activity has been thoroughly investigated (Mulder et al., 2012; Amaro et al., 2016; Hall et al., 2017; Aslam et al., 2018; Heijnen et al., 2022) and linked to the activity of the Fleuve Manche palaeoriver (Bourillet et al., 2003; Zaragosi et al., 2006; Toucanne et al., 2008, 2009, 2012).

A causal relationship between erosive downslope currents associated with the start of glacial-interglacial cycles in the North Atlantic and the formation of the GCS has been suggested from Neogene-Quaternary seismic stratigraphy (Chapter 2). However, the dataset used by those authors is not sufficient for this investigation, as the GCS channel bases are situated below the penetration depth of their seismic data. Therefore, the origin of the GCS and the potential link with Quaternary glacial-interglacial cycles needs further examination that is based on seismic data also covering the deeper subsurface. Previous stratigraphic studies in the Porcupine Seabight have focused on the Neogene and Quaternary stratigraphy in the area of the BMP (Van Rooij et al., 2003, 2009; Thierens et al., 2012), on the very general Cenozoic stratigraphy of the basin (McDonnell and Shannon, 2001; Stoker et al., 2005a), and briefly on the morphology of the GCS channel heads (Wheeler et al., 2003), but never comprehensively on the stratigraphy of the GCS channels.

The aim of this study is to investigate which processes were involved in the formation of the GCS and to assess the spatiotemporal evolution of the system. To this end, a seismic stratigraphic framework was built from the channel heads on the upper slope down to the foot of the slope and from the acoustic basement to the present-day seafloor.

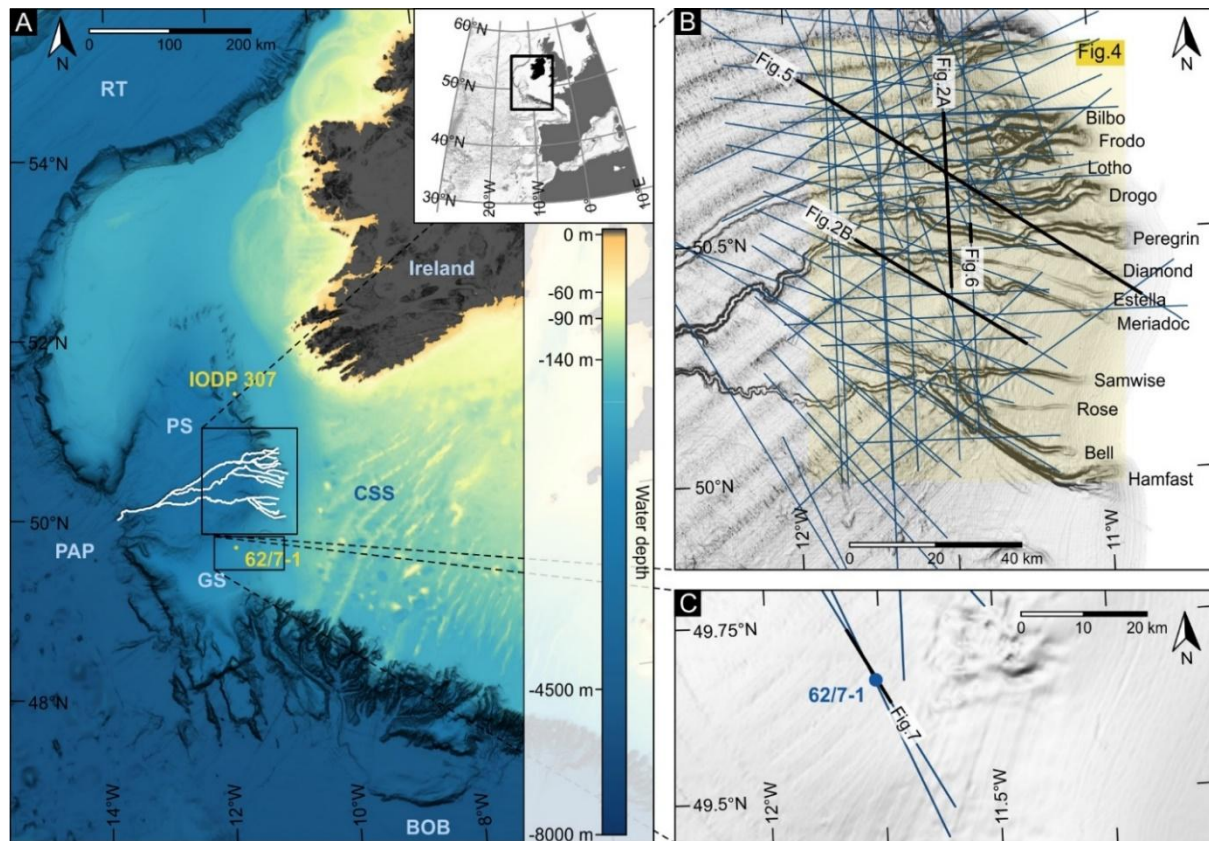


Figure 3.1. (A) Bathymetry map situating the study area in the regional context. The inset shows the location on the West-European margin. The locations of IODP Expedition 307 and well 62/7-1 are indicated in yellow. BOB, Bay of Biscay; CSS, Celtic Sea Shelf; GS, Goban Spur; PS, Porcupine Seabight; PAP, Porcupine Abyssal Plain; RT, Rockall Trough. (B) Slope-shaded map of the study area (yellow shaded area) with the names of the individual channels. Seismic lines used in this study are indicated in blue and profiles used for figures are indicated in black. The location of Figure 3.4 is indicated by the shaded rectangle. (C) Close-up bathymetry map showing the location of the seismic lines going through the location of well 62/7-1 and the location of Figure 3.7. Bathymetry maps are based on Irish Public Sector Data from INFOMAR (Geological Survey Ireland and Marine Institute, 2024) licensed under a Creative Commons Attribution 4.0 International (CC BY 4.0) license.

3.2 Geological Setting

The GCS is situated in the southern part of the Porcupine Seabight, a pear-shaped oceanic basin that forms an embayment on the northeast Atlantic margin (Figure 3.1A). The Porcupine Seabight is the surface expression of the Porcupine Basin, a failed rift basin that formed from the Permian onwards (Shannon, 1991). It owes its north-south elongation to a major Late Jurassic rifting episode with east-west extension associated with North Atlantic seafloor spreading (Shannon, 1991; Naylor and Shannon, 2005; Shannon et al., 2007). The basin experienced thermal subsidence from the Early Cretaceous to the Holocene (Shannon, 1991). Open marine conditions were established for the first time in the Late Cretaceous (McCann et al., 1995; McDonnell and Shannon, 2001), and, after a period of regression in the late Paleocene – late Eocene (Moore and Shannon, 1992), again and finally in the latest Eocene – earliest Oligocene (McDonnell and Shannon, 2001; Stoker et al., 2005b).

A major uppermost Eocene erosional unconformity (regionally identified on seismic data as a high-amplitude reflection called reflector C30) marks the onset of deep-water current circulation in the basin and the change from downslope- to alongslope-dominant sediment transport characterised by drift deposition (Stoker et al., 2001, 2005b; Shannon et al., 2005). The circulation pattern changed in the early to middle Miocene from Tethyan-influenced bottom current activity to Nordic seas-influenced bottom current activity (Stoker et al., 2005b) resulting in an uppermost lower Miocene regional unconformity (regional seismic reflector C20; McDonnell and Shannon,

2001). In the Rockall Trough, the high amplitude of the C20 reflection may be further enhanced by the presence of a diagenetically altered ash layer (Stoker et al., 2001). Another period of extensive contourite drift development in the middle to late Miocene (McDonnell and Shannon, 2001; Shannon et al., 2005) was ended by an episode of regional submarine erosion (regional seismic reflector C10 characterised by a continuous reflection of high amplitude) likely from increased North Atlantic Deep Water input from the Nordic seas along with basin margin tilting in the early Pliocene (Stoker et al., 2001, 2005a, 2005b). The youngest stratigraphic interval is characterised by the presence of coral mounds and contourite drifts, and the upper Quaternary sedimentation is influenced by glacial activity (Tudhope and Scoffin, 1995; Van Rooij et al., 2007b). The Porcupine Basin contains up to 6 km of Mesozoic sediments and the Cenozoic fill is up to 4 km thick (Moore and Shannon, 1992; McDonnell and Shannon, 2001).

Over the eastern upper slope of the basin, previous studies have recognised a discontinuity that was attributed an early middle Miocene age (seismic reflector RD2 of Van Rooij et al. (2003)) (Louwye et al., 2008) and a discontinuity representing a hiatus between upper Miocene and upper Pliocene sediments (seismic reflector RD1 of Van Rooij et al. (2003)) (Kano et al., 2007; Louwye et al., 2008; Van Rooij et al., 2009). They have been proposed as equivalents to the regional latest early Miocene (C20) and early Pliocene (C10) unconformities, respectively (Van Rooij et al., 2003), but have never been tied on any seismic line. In the specific setting of the eastern slope of the Porcupine Seabight, the origin of the RD1 discontinuity has been pinpointed in more detail than the C10 unconformity. The RD1 discontinuity is the result of the combined action of both alongslope and downslope currents associated with, respectively, the influence of the Mediterranean Outflow Water on the oceanographic regime in the Seabight, and the onset of glacial-interglacial cycles in the North Atlantic (Van Rooij et al., 2003). The main sediment source for the Quaternary downslope gravity currents is the Celtic Sea Shelf, with principal supply during lowered sea levels of glacial periods (Akhmetzhanov et al., 2003). The RD1 seismic reflector is very obvious from its erosional characteristics and the distinct differences between the units above and below in the BMP, but is known to become harder to trace to the south, towards the study area (Chapter 2).

3.3 Material and methods

To build the seismic stratigraphy needed for this study, well data were integrated with seismic reflection data, which were then analysed considering the seismic stratigraphy that was established in other parts of the Porcupine Seabight by McDonnell and Shannon (2001) and Van Rooij et al. (2003, 2009). These studies are also the basis for the nomenclature of the seismic units and their bounding surfaces used here (Figure 3.2): C30 from McDonnell and Shannon (2001), and RD2 and RD1 from Van Rooij et al. (2003, 2007a).

The four multichannel seismic reflection datasets used in this work (Nopec Geoventures 1997, Geco-Prakla 1997, Fugro-Geoteam 1997 and Spectrum 1997) were provided by the Irish Department of the Environment, Climate and Communications. All four were recorded in 1997 with airgun sources. They were processed prior to receiving them, with processing steps including dip moveout, stacking, deconvolution, as well as several filters depending on the dataset. A brief overview of the processing flow used for each dataset can be found in Appendix A and some additional details are available on the website of the Irish Department of Communications, Energy and Natural Resources (Department of Communications, 2010). The datasets are used here without any additional processing. The dominant frequency in the seismic data is 5-70 Hz, and their vertical resolution decreases from c. 10 m in the shallow subsurface to c. 20-30 m near the acoustic basement. The profiles were loaded into S&P Global Kingdom software for analysis of the reflectors and integration with well data. Within the Kingdom software, a mistie analysis including time shifts, phase shifts and amplitude scaling was performed to allow good correlation between the four datasets.

Key horizons were first identified in the Kingdom software and then gridded to acquire palaeotopography maps using the Surfer package from Golden Software. Kriging was found to be the gridding method that produced the most realistic maps. The grid spacing was set to 100 m. Thickness maps of specific stratigraphic sections were also created in Surfer.

The drainage networks used to compare the present-day channel network to depressions in palaeotopographical surfaces were generated using the “Generate Watershed” analysis tool in Global Mapper software, which is based on the eight-direction pour point algorithm from Jenson and Domingue (1988) to determine the flow direction at each point of the analysed grid. The horizontal resolution of the watershed was set equal to the resolution of the analysed grid (100 m). As the focus is on the large-scale drainage pattern of the area, the stream cell threshold was set to 500 and the smallest secondary thalweg sections were removed. Minimal smoothing of calculated thalwegs was done manually. To calculate the amount of overlap between the locations of the present-day channel thalwegs and depressions in the palaeotopography, buffers were first created around the main thalwegs of the palaeowatershed. The widths of the buffers correspond to the widths of the palaeodepressions and were measured on the appropriate reflector in the seismic data. Then, the length of the present-day thalwegs intersecting the palaeowatershed buffers was divided by the total length of the present-day thalwegs in the study area. Thus, a percentage value for the amount of overlap between the locations of the present-day watershed and the palaeowatershed was obtained.

Age control of the seismic data was partly achieved through time-depth conversion of the stratigraphy at well 62/7-1. Well data were provided by the Irish Department of the Environment, Climate and Communications and included descriptions of lithological and palaeontological samples, log data, interval velocity data, and the well completion report. Well 62/7-1 is located on Goban Spur (Figure 3.1) at 993 m water depth and reaches a total depth of 3669 m below the seafloor, targeting potential Bathonian and Sinemurian reservoirs. The time-depth conversion of the stratigraphy at well 62/7-1 was performed using data from two airgun vertical seismic profiles covering complementary depth intervals. In these surveys, the travel time between source and receiver is measured and the distance between them is used to calculate average velocities and interval velocities at the location of the well. The results of both surveys were combined and are shown in Appendix B. Since this well is located > 100 km from the most distant (northern) part of the study area, time-depth conversion was only performed at the location of the well. All depths related to well 62/7-1 are reported with respect to the seabed. The well stratigraphy is based on downhole logging data, and palaeontology and lithology obtained from cuttings and side wall cores described in the well reports. In this well, the shallowest sample was taken at a depth of 615 m, so the well data could only be tied to the deepest sections of the seismic data.

The chronostratigraphy of the shallower seismic data was based on the previously established general Cenozoic seismic stratigraphy of the Porcupine Seabight (McDonnell and Shannon, 2001; Shannon et al., 2005) and the Neogene and Quaternary seismic stratigraphy of the BMP (Van Rooij et al., 2003, 2009). These stratigraphies were in turn based on well data from sixteen exploration and appraisal wells (McDonnell and Shannon, 2001), and data from the IODP 307 boreholes (Van Rooij et al., 2009). Tying the seismic data used here to the previously established stratigraphy was possible due to distinct similarities in seismic expression and a direct seismic connection between the study area and the BMP.

3.4 Results

3.4.1 Seismic reflection data

Generally, the sedimentary fill of the Porcupine Basin displays a catenary geometry: horizons are near-horizontal in the centre of the basin and dip towards the basin centre (with a greater dip further away from the centre) on the edges (Figure 3.2B). Five seismic units (U5 – U1; from old to young) bounded by six reflectors (basement, NTC, C30, RD2, RD1, seafloor; from old to young)

have been identified from the acoustic basement to the seafloor (Figure 3.2). A seventh reflector (ECP) is considered as an internal reflector in seismic unit U2, but as it is seen as a key reflector in the evolution of the GCS, it is here described separately.

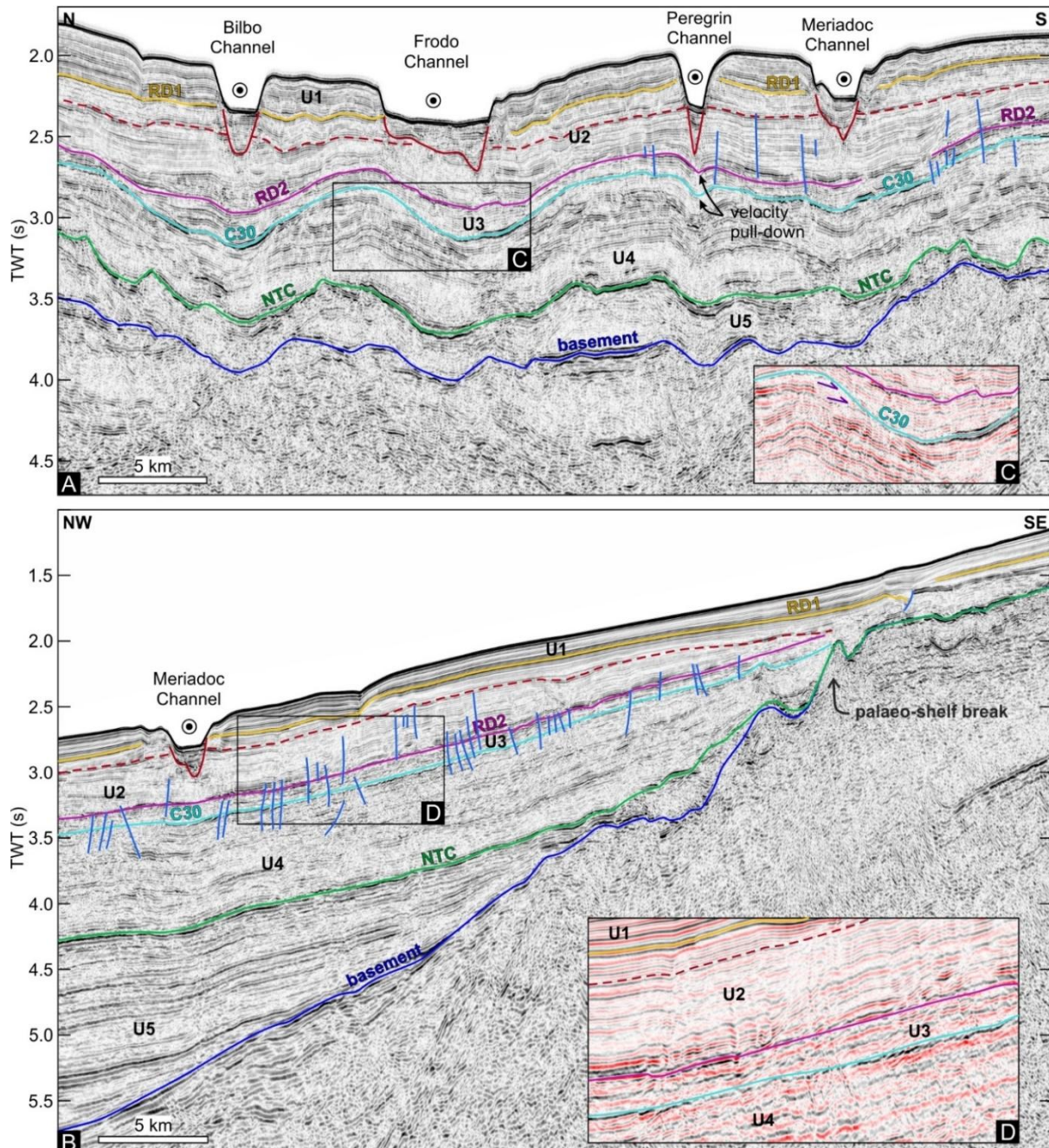


Figure 3.2. (A,B) Interpreted airgun seismic reflection profile showing the five units (U1-U5) and their boundaries (seafloor, RD1, RD2, C30, NTC, basement) that make up the general stratigraphy of the study area. The bases of the present-day GCS are indicated in red. Faults are marked in blue. The ECP reflector is indicated as a red dashed line. The locations are shown in Figure 3.1B. (C) Closeup of the seismic profile in (A) showing the erosion (purple arrows) underneath the C30 reflector at the top of unit U4. (D) Closeup of the seismic profile in (B) highlighting the difference in seismic facies between units U3 and U2.

3.4.2 Key seismic reflectors

The basement reflector is the deepest high-amplitude reflection that is mostly continuous and represents the boundary between the discontinuous to chaotic facies of the acoustic basement and the more continuous facies of the overlying seismic units (Figure 3.2). It is a surface with a very irregular topography and often has an erosional character.

The near top Cretaceous (NTC) reflector is characterised by a moderate- to high-amplitude reflection and is generally continuous in the basin (Figure 3.2). A palaeo-shelf break is inferred from a change in slope gradient in the seismic data at c. 2-2.2 s TWT (Figure 3.3A). The outermost part of the palaeo-shelf has an irregular topography but is generally quasi horizontal to slightly basinward-dipping (2-3°). The steeply dipping (10-15°) palaeo-slope section is smooth to irregular and generally becomes smoother towards the foot of the slope and into the basin, where it displays large-scale undulations of small amplitude and irregular wavelength (Figure 3.2). The NTC reflector is conformable to over- and underlying reflectors in the basin, but erosional on the palaeo-slope and near the palaeo-shelf break. On the palaeo-shelf, it often coincides with the basement reflector. In the absence of obvious seismic terminations the relation to the underlying package is difficult to determine.

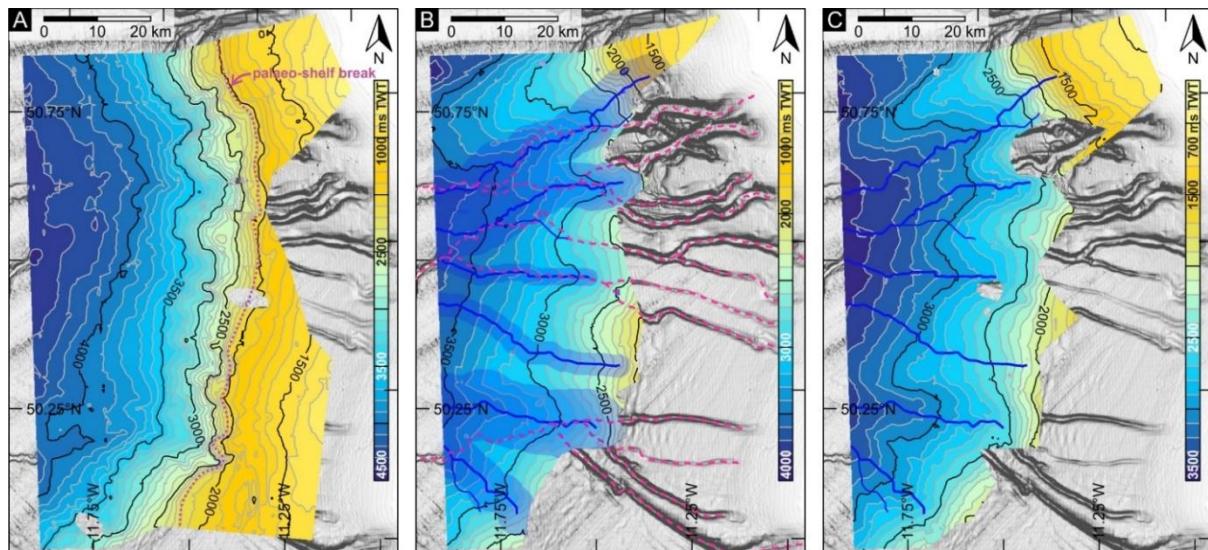


Figure 3.3. Palaeotopography maps of the (A) NTC, (B) C30, and (C) RD2 surfaces expressed in ms TWT from present-day sea level plotted over a slope-shaded bathymetry map of the present-day seabed. (A) The interpreted palaeo-shelf break is indicated by a pink dotted line. (B) The watershed of the C30 palaeochannels is indicated in blue solid lines and the buffers of the main palaeochannels are shaded blue. The thalwegs of the present-day GCS are indicated in pink dashed lines. (C) The watershed of the C30 palaeochannels [identical to (B)] is indicated in blue solid lines. Note that the colour scales of the three maps differ.

The C30 reflector has a wavy to irregular topography and a high-amplitude reflection (Figure 3.3B). It is offset by numerous faults and the identification of this reflector becomes more uncertain towards the south(east). Within the study area, the C30 surface displays six NE-SW to SE-NW oriented channels. Data in the northwesternmost part of the study area suggest more channels are present to the north (Figure 3.3B). The channels observed here are c. 3-13 km wide and c. 100-500 m deep. At the channel bases and edges, the otherwise conformable C30 reflector truncates the reflectors of the seismic unit below (Figure 3.2C). Near the eastern edge of the study area, the C30 reflector is part of a package of reflectors that onlaps the NTC reflector.

The RD2 reflector is erosional to conformable, and, like the C30 reflector, it is offset by numerous faults (Figure 3.2). It has a large-scale undulating appearance in N-S profiles, especially towards the upper slope. The RD2 undulation pattern mimics that of the C30 reflector: RD2 lows, though they are less pronounced than the C30 palaeochannels, are situated over the C30 palaeochannels (Figure 3.3C). The RD2 reflector is only moderately continuous, which makes it difficult to trace towards the south and the east, where it also loses its erosional characteristics. A change in seismic facies, generally from more faulted just below the reflector to slightly more distinct and continuous stratification, sometimes with higher amplitude, just above the reflector, may then help in identifying it (Figure 3.2D). Though the RD2 reflector can be traced towards the

present-day shelf edge in the northernmost part of the study area, it either onlaps one of the underlying units or is eroded by overlying reflectors at the upper slope in the rest of the study area.

The early channel presence (ECP) reflector is characterised by a moderate-amplitude reflection. It is conformable to (small-scale) erosional and has an overall smooth topography without any significant features. It is only present in the northern and central parts of the study area, and is eroded by overlying reflectors in the upslope and southern parts of the study area.

The RD1 reflector is mostly continuous and conformable with the under- and overlying reflectors in the GCS interfluvial areas. It can be truncated and/or truncating at the GCS channel edges. It has an overall smooth topography without the large wavy structures that the RD2 and C30 reflectors have. The main feature in the RD1 palaeotopography are the gaps caused by the incisions at the base of the present-day GCS.

3.4.2.1 Seismic stratigraphy

Unit U5 is the lowermost seismic unit that can be distinguished above the acoustic basement (Figure 3.2). It consists of (sub)parallel reflections that onlap the acoustic basement, filling in the basin. The reflections are mostly continuous, have moderate to high amplitudes, and are sub-horizontal though in some places slightly wavy towards the centre of the basin (Figure 3.2). The NTC reflector bounding the top of the unit truncates U5 reflectors near the basin margin but is conformable in the basin centre. The unit pinches out against the basement at the basin margins and is not present east of the inferred NTC palaeo-shelf break (Figure 3.2B). The thickness of unit U5 increases from the basin margins to the basin centre to 2 s TWT (Figure 3.4A).

Unit U4 is bound by the NTC reflector at the bottom and the C30 reflector at the top. It consists of (sub)parallel to slightly mounded reflections of low to moderate amplitude that are mostly continuous, except for some places where localised internal erosion is visible within the unit (Figure 3.5C). Additionally, the channels of the C30 reflector cut down into unit U4 and remove the upper 200 - 500 ms TWT of the unit. The thickness of unit U4 increases from 0 s TWT on the upper slope, where it onlaps against the steep slope created by the NTC reflector, to 950 ms TWT in the basin, where it fills up the topography of the NTC reflector (Figure 3.4B). Thicker packages in this unit correspond to the interfluvial areas of the C30 palaeochannels.

Unit U3 is bound by the C30 (bottom) and RD2 (top) reflectors. It contains reflections that are generally lower in amplitude compared to unit U4 and are disrupted by a multitude of faults (Figure 3.5A). Reflectors are mostly subparallel to very slightly mounded though the faulting can make them look very discontinuous and almost chaotic (Figure 3.2B). The U3 reflectors infill the inferred C30 channels and drape the C30 reflector where it created topographic highs (Figure 3.2, 5). Like unit U4, unit U3 pinches out towards the eastern edge of the basin. The thickness of U3 varies mostly between 0 and 250 ms TWT, except in the northwesternmost part of the study area, where a thick (500 ms TWT) package is present (Figure 3.4C). This thicker package is elongated in a NE-SW direction and its location and geometry correspond to a C30 palaeochannel just northwest of the study area. In general, unit U3 is thicker where it fills the C30 palaeochannels and thinner on the topographic highs. In the northernmost part of the study area, however, the lower part of unit U3 consists of horizontal reflectors that fill one of the C30 palaeochannels while the top of the unit has a mounded geometry and is preferentially deposited on the interfluvial (Figure 3.5C). The mounded geometry of the upper part of U3 here is also visible as a thicker package in the isopach map (Figure 3.4C).

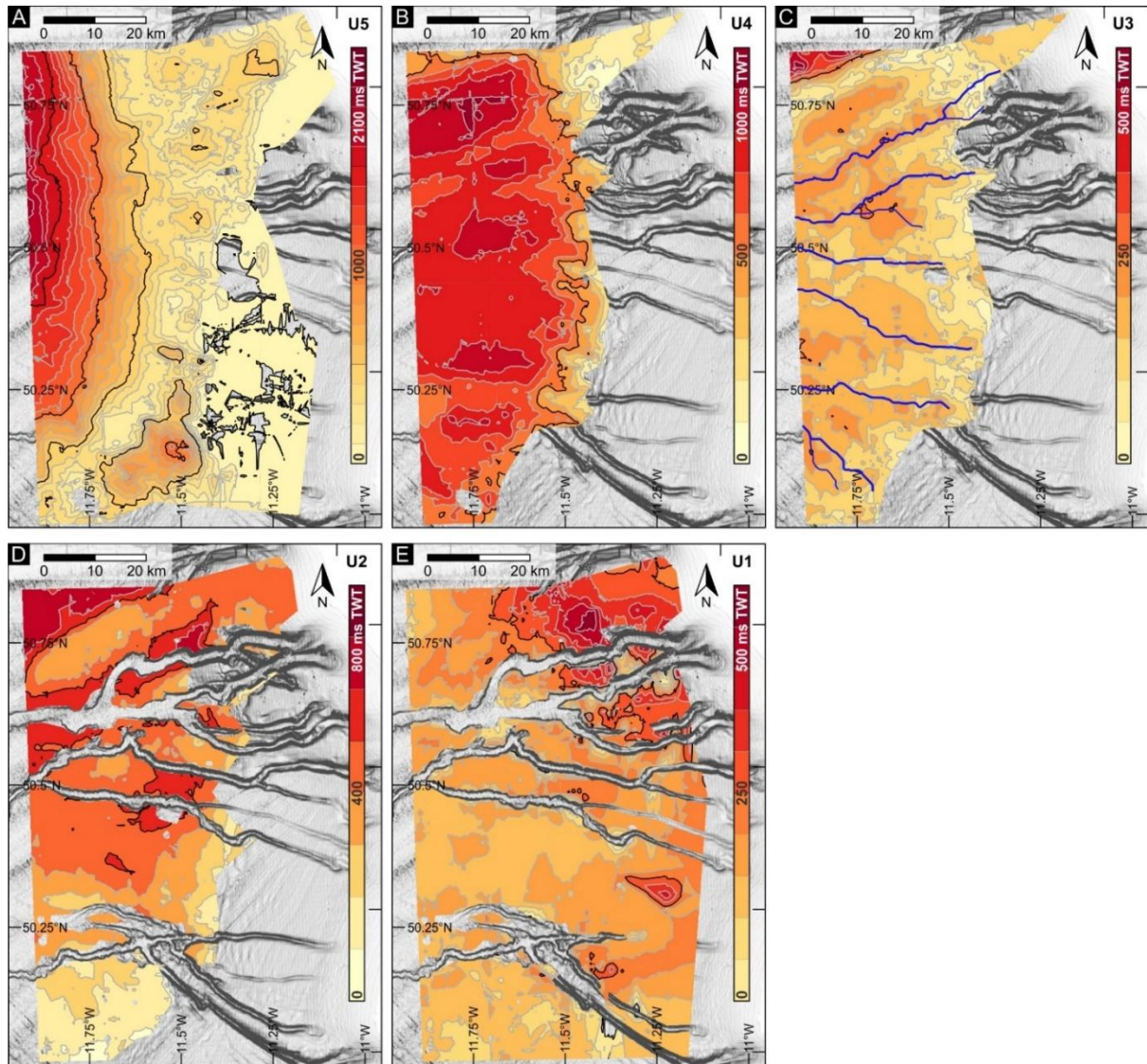


Figure 3.4. Isopach maps (plotted on top of present-day slope-shaded bathymetry maps) of (A) unit U5 between the acoustic basement and NTC reflectors, (B) unit U4 between reflectors NTC and C30, (C) unit U3 between reflectors C30 and RD2, where the thalwegs of the C30 surface are indicated in blue solid lines, (D) unit U2 between reflectors RD2 and RD1, and (E) unit U1 between reflector RD1 and the seafloor. The isopachs are calculated only where both bounding reflectors are present. This creates discontinuities in the maps, especially underneath the present-day GCS where the incisions at the channel bases cut through units U2 (D) and U1 (E). See Figure 3.1B for the location of the study area. Note that the colour scale differs in each panel.

The base of unit U2 is the RD2 reflector, while at the top it is bound by the RD1 reflector. Where present, the ECP reflector is contained in this unit. The reflections within unit U2 have a low to moderate amplitude similar to unit U3 below. The bases of the present-day GCS channels locally remove 50 - > 400 ms TWT at the top of the unit. Although unit U2 is dissected by numerous faults, the reflections are generally laterally traceable over long distances. They dip towards the centre of the basin, and onlap or drape the RD2 reflector. In proximity (1-2 km) to the channel bases of the present-day GCS, U2 reflections may be divergent (Figure 3.6). The thickness of unit U2 generally decreases towards the upper slope and to the south, varying between 0 and 750 ms TWT (Figure 3.4D). When comparing the thickness distribution of unit U2 with the RD2 palaeotopography, it is apparent that the unit heals the RD2 palaeotopography being thicker in the lows and thinner on RD2 palaeo-highs (compare Figure 3.3C and Figure 3.4D).

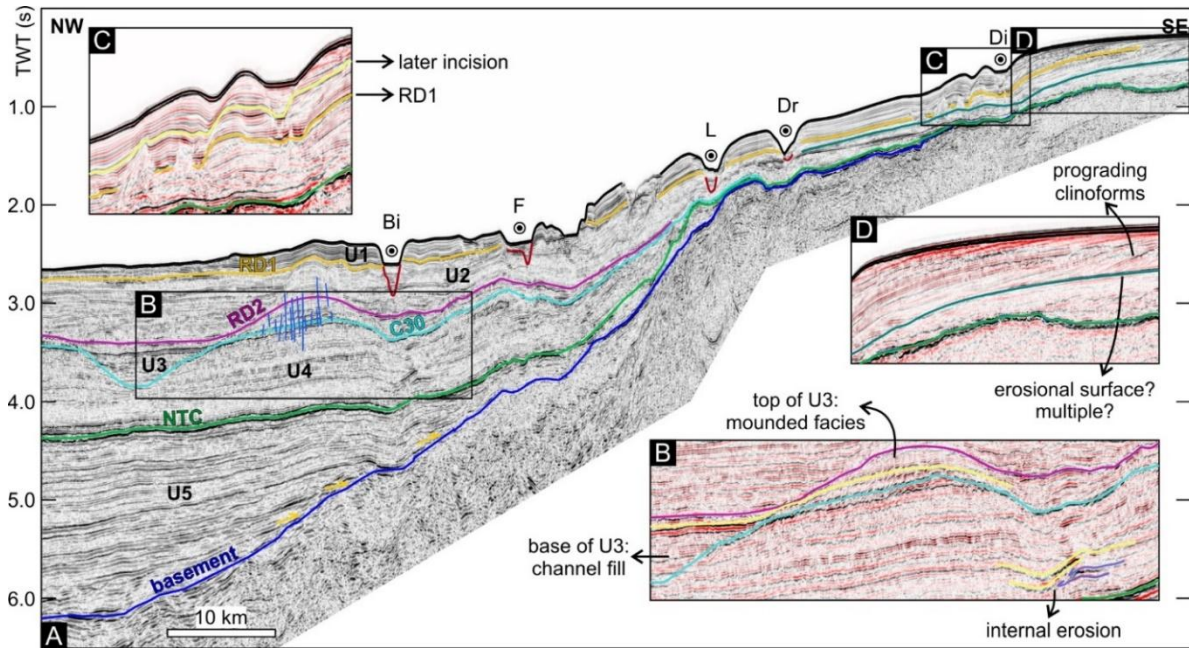


Figure 3.5. (A) NW-SE oriented interpreted seismic profile (see Figure 3.1B for location; colours are as in Figure 3.2). The yellow arrows indicate onlapping onto the basement reflector. Bi, Bilbo Channel; F, Frodo Channel; L, Lotho Channel; Dr, Drogo Channel; Di, Diamond Channel. (B) Close-up of seismic section showing internal erosion in unit U4, basin filling reflections at the base of unit U3 and a mounded facies at the top of unit U3. (C) Close-up of seismic section showing incision after RD1. (D) Close-up of seismic section showing prograding clinoforms as oblique reflections.

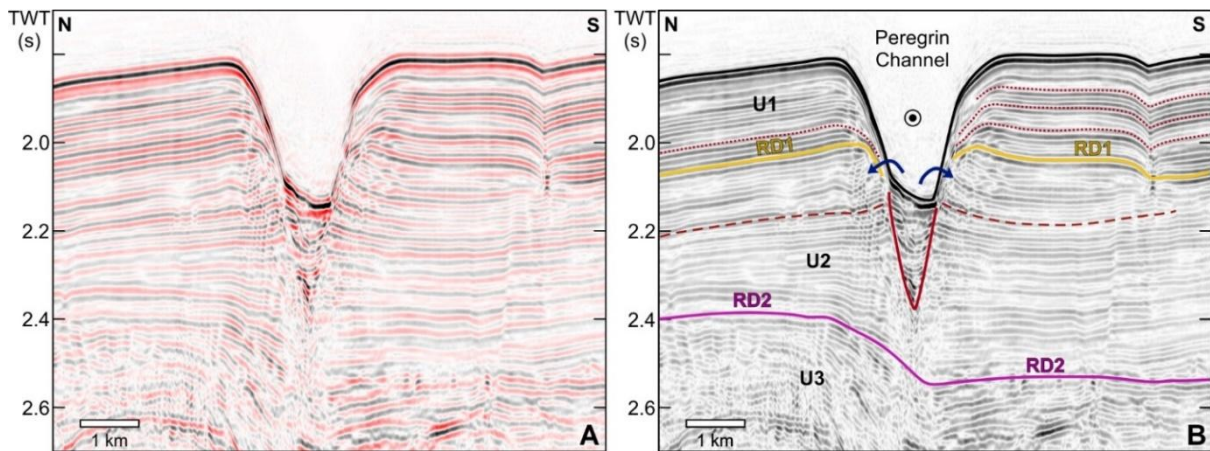


Figure 3.6. (A) Close-up section of a seismic profile (location: see Figure 3.1B) showing the evolution of Peregrin Channel and (B) interpretation indicating several phases of channel incision (red dotted lines). The channel base is indicated by the red solid line. Levee build-up is indicated by the black arrows. The ECP reflector is indicated as a red dashed line.

Unit U1 is the uppermost seismic unit between the RD1 reflector and the seafloor. It is characterised by channels cutting through the unit, sometimes all the way to and through the base. The moderate- to high-amplitude reflections of unit U1 are well-stratified, parallel, dipping towards the basin centre, conformable and continuous, except at the channel edges of the present-day GCS, where they are generally truncated and locally divergent. Wavy reflector configurations can be found in proximity to some of the channels. Unit U1 is thickest on the upper slope present-day interfluvies, especially surrounding the two northernmost channels (Figure 3.4E). The thickness of the unit generally decreases to the south and there is some pinching out of reflectors towards the basin centre, which is reflected in a thickness decrease downslope from 500 ms TWT to 100 ms (Figure 3.4E).

3.4.2.2 Comparison of palaeo- and present-day watersheds

Calculated watersheds show that the thalwegs of the present-day channel system are often situated close to or on top of the C30 palaeochannels. Where the C30 reflector is present (coloured area in Figure 3.3B), 85 % of the present-day thalwegs intersect with a C30 palaeochannel. When also considering the upslope sections of the present-day channels where the C30 reflector is not present due to erosion or onlapping, 34 % of the channel sections intersect with a C30 palaeochannel.

3.4.3 Well 62/7-1 and ties to seismic data

As lithological samples and all logs except for the gamma ray log were retrieved from below 560 m depth, the well data can only be used to link the deeper parts of the seismic stratigraphy (seismic unit U5 and part of seismic unit U4) to the Middle Jurassic to early Eocene sections of the well stratigraphy.

At the location of well 62/7-1 (Figure 3.1), the acoustic basement is not easily identified but is likely situated at 2330 m depth, at the top of a section of calcareous shales of early Bajocian age. It corresponds to a sharp upwards decrease in gamma ray, neutron porosity and sonic velocity (Figure 3.7).

Seismic unit U5 consists of (calcareous) siltstones and claystones at the bottom, followed by a sandstone section with a c. 220 m-thick section of volcanic rocks on top. At 2070-2060 m depth, a package of coal and siltstones is interbedded with the volcanic rocks. From the top of the volcanic section at 1881 m depth up to c. 1040 m depth, the lithology is predominantly cherty limestone, with sandy intervals at the bottom. A chalk interval is present between 1040 and 911 m depth. Generally, unit U5 is characterised by upwards decreasing gamma ray values. There is high variability in the neutron density and sonic velocity logs in the section of volcanic rocks and the chalk section is characterised by an increase in neutron density and decrease in sonic velocity (Figure 3.7).

The NTC reflector, which forms the upper boundary of seismic unit U5, is situated at 925 m depth in Maastrichtian chalks. A sample at 911 m depth is the uppermost sample dated to the late Maastrichtian, while a sample retrieved at 908 m depth was dated to the early Paleocene. This would put the top of the Cretaceous section in the well between 908 and 911 m. It is also identified as a sharp change in the gamma ray, neutron porosity and sonic velocity logs (Figure 3.7).

The section between 911 m depth and the uppermost retrieved sample at 615 m depth, which corresponds to the lower part of seismic unit U4, is predominantly calcareous claystone, with some limestone interbeds near the bottom of the section and occasional sandy intervals throughout (Figure 3.7). The uppermost sample of late Paleocene age was sampled at 882 m depth. The youngest dated sample in the well was taken at a depth of 622 m and was dated to the early Eocene. Generally, the gamma ray values and sonic velocity decrease towards the top of this section, while neutron density increases.

The C30 and RD1 reflectors are both situated above the youngest dated sample, at 260 m depth and 120 m depth, respectively. The RD2 reflector would stratigraphically be placed in between these two reflectors but could not be traced from the study area to the location of the well.

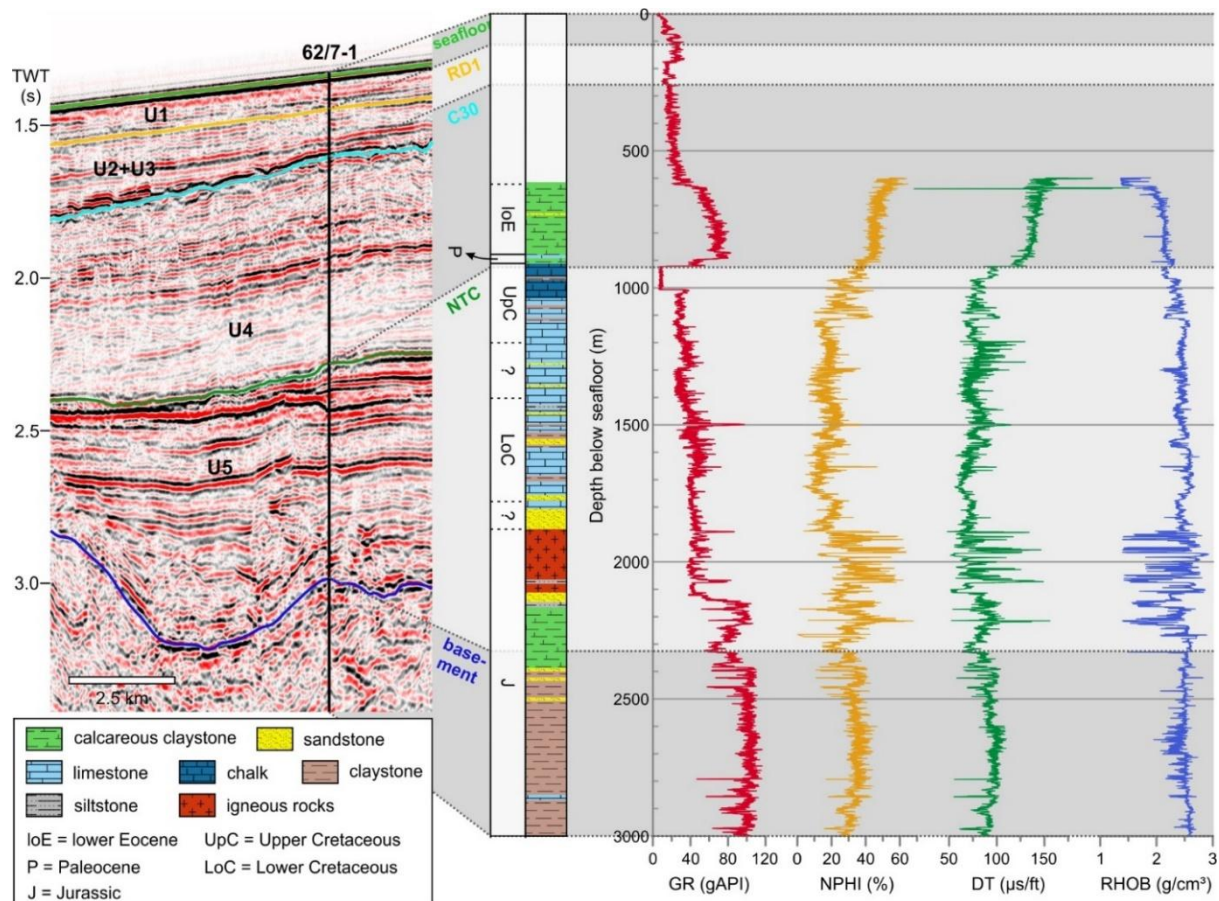


Figure 3.7. A section of a seismic line crossing well 62/7-1 (see Figure 3.1 for location) illustrating the correlation between well (from left to right: lithology, gamma ray log (GR), neutron density log (NPHI), sonic velocity log (DT) and density log (RHOB)) and seismic data. For the definition of the seismic units and boundary reflectors, see text.

3.5 Discussion

3.5.1 Chronostratigraphy

Several seismic lines in the dataset allow the seismic stratigraphy in the study area to be connected to the stratigraphy in the BMP, which was based on data from the IODP 307 boreholes (see Figure 3.1 for location; Van Rooij et al., 2003, 2007a, 2009; Kano et al., 2007). The data from well 62/7-1, as well as additional previously published work (McDonnell and Shannon, 2001; Stoker et al., 2001; Shannon et al., 2005), were used to determine the age of units and reflectors below the coring depth of the IODP 307 boreholes.

At the location of well 62/7-1, the basement reflector is situated in Bathonian-Bajocian rocks (Figure 3.7). The biostratigraphy data from the well demonstrate that the NTC reflector is Maastrichtian in age and situated near the top of the Late Cretaceous chalk section (Figure 3.7). Therefore, at this location, unit U5 is of Middle Jurassic to Late Cretaceous age.

As the C30 reflector is situated above the youngest dated sample in well 62/7-1 (Figure 3.7) and underneath the depth of the IODP 307 boreholes, its age cannot be determined from the well or borehole data. The large-scale undulating geometry and erosional characteristics that it displays (Figure 3.2), however, are distinctive for the C30 reflector in the southern part of the Porcupine Basin (Shannon et al., 2005). Previous studies identifying this reflector were able to date it to the latest Eocene by linking seismic data to wells in the northern part of the basin (McDonnell and Shannon, 2001; Stoker et al., 2001). Consequently, unit U4 sediments are of Paleocene to Eocene age.

The RD2 reflector in the study area is the continuation of the RD2 reflector in the BMP and Kings Channel System (Chapter 2; Van Rooij et al., 2003, 2009). The reflector was dated to the early middle Miocene in IODP 307 borehole U1318 (Louwye et al., 2008), so unit U3 in the study area contains sediments of Oligocene to early Miocene age.

The RD1 reflector identified here is the continuation of the RD1 reflector in the BMP and Kings Channel System (Chapter 2; Van Rooij et al., 2003, 2009). The RD1 reflector was dated at IODP 307 site U1318, where it represents a hiatus between upper Miocene and upper Pliocene sediments (Kano et al., 2007). The erosive phase responsible for forming the RD1 reflector seems to have been extreme in the BMP (Van Rooij et al., 2007a), where strong bottom currents with local enhancement by tidal motions have been inferred from mooring data (Pingree and Le Cann, 1989; White, 2007). They removed (part of) the sediments below the RD1 horizon, causing a steep topography and a large hiatus (Van Rooij et al., 2003, 2009; White, 2007). It is doubtful that the extent of this hiatus is consistent over the whole eastern margin of the Porcupine Seabight (Huvenne et al., 2009b; Van Rooij et al., 2009). There is no indication for the presence of locally enhanced strong bottom currents here (Figure 3.2). Though the RD1 reflector is still erosional in places, its general conformability suggests less extreme current strengths and (non-)deposition rather than (extreme) erosion, which may indicate that it represents a smaller hiatus in the GCS than in the BMP. There is, therefore, no constraint on the exact age of the RD1 horizon (or the associated hiatus) in the study area at present. Due to a lack of precise age information in the study area, the seismic interpretation of the BMP at the IODP 307 sites is adopted here and sediments of unit U2 are tentatively attributed a middle Miocene to late Miocene/Pliocene age.

The resumption of sedimentation after the RD1 erosion is variable, even over short distances (Huvenne et al., 2009b). Cold water coral mound frameworks promoting entrapment of particles allowed sedimentation as early as 2.7 Ma (Kano et al., 2007; Foubert and Henriët, 2009). In locations without such a framework to decrease bottom current intensity, sedimentation may not have resumed before 1.7 Ma (Kano et al., 2007; Huvenne et al., 2009b). Therefore, unit U1 in the study area is thought to mainly consist of Quaternary sediments.

3.5.2 Processes and environments prior to GCS incision

The data at well 62/7-1 indicate that the seismic unit just above the acoustic basement (U5) consists of Jurassic calcareous claystones and igneous rocks underneath Cretaceous limestones and chalks (Figure 3.7). A coal-containing section interbedded with the igneous rocks suggests continental conditions at Goban Spur at the end of the Jurassic and several (very) calcareous sandy sections interbedded with limestones indicate a shallow and/or more dynamic marine depositional environment at the start of the Cretaceous. The upwards decreasing gamma values within the unit along with calcareous lithologies becoming more and more dominant (Figure 3.7) suggest a reducing terrigenous influence on deposition and likely deepening of the basin. This is in accordance with the onset of thermal subsidence from the Early Cretaceous onwards (Shannon, 1991). Underneath the GCS, where the basin was deeper than at Goban Spur (Merlin Energy Resources Consortium, 2020), the (sub)parallel basin-filling reflections of seismic unit U5 (Figure 3.5A) are interpreted as an expression of low-energy deposition in an open marine environment during the relatively warm Late Cretaceous (Figure 3.8A). This is in accordance with the presence of (very) calcareous lithologies on Goban Spur as well as previous work in the basin (McDonnell and Shannon, 2001; Stoker et al., 2017; Merlin Energy Resources Consortium, 2020).

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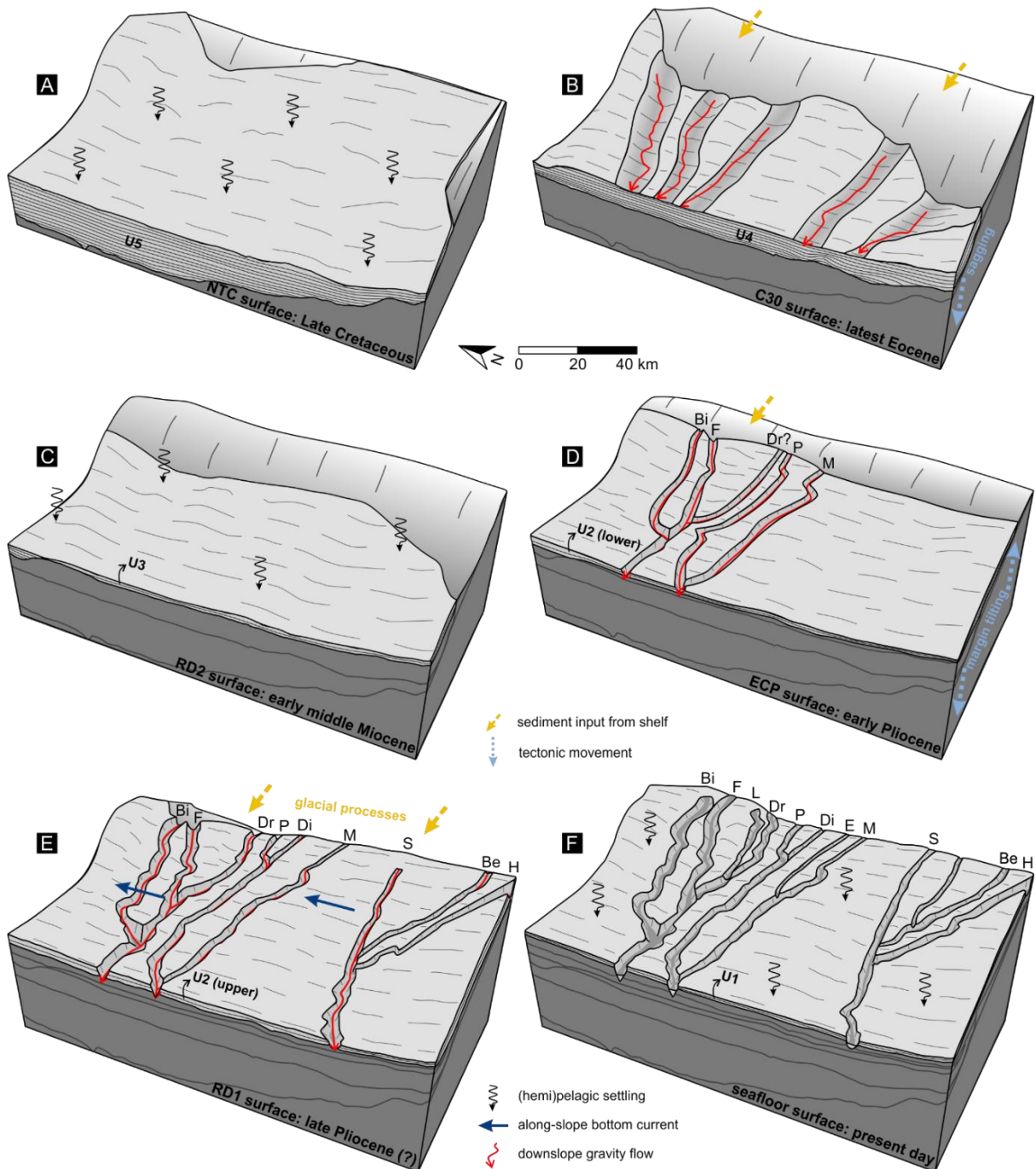


Figure 3.8. Sketches depicting the evolution and most prominent features and processes of the study area in the (A) Late Cretaceous, (B) latest Eocene, (C) early middle Miocene, (D) early Pliocene, (E) late Pliocene, and (F) present day. The names of the present-day GCS are indicated as follows: Bi, Bilbo Channel; F, Frodo Channel; L, Lotho Channel; Dr, Drogo Channel; P, Peregrin Channel; Di, Diamond Channel; E, Estella Channel; M, Meriadoc Channel; S, Samwise Channel; R, Rose Channel; Be, Bell Channel; H, Hamfast Channel.

Shannon et al. (1993) identified a last phase of chalk deposition in the Danian rocks in the Porcupine Basin that are not present in well 62/7-1, where the Cenozoic sequence starts with a section of younger-than-Danian calcareous claystones (Figure 3.7). This suggests the low-energy open marine setting of the Late Cretaceous persisted in the beginning of the Paleocene. The calcareous claystone section of the upper Paleocene and lower Eocene in well 62/7-1 shows that Goban Spur was not influenced by the deltaic sedimentary environment that developed in the northern part of the Porcupine Basin (Moore and Shannon, 1992). The seismic character of unit U4 in the study area with (sub)parallel reflections only occasionally and locally interrupted by

internal erosion (Figure 3.5B) suggests it was predominantly governed by hemipelagic rather than deltaic processes. It was likely a marine basin environment, as also put forward by Merlin Energy Resources Consortium (2020). The large-scale incisions of the C30 reflector bounding the top of the unit indicate a period of major erosional impact from channelised gravity-driven clastic input, probably in the late Eocene as proposed by Shannon (1992) and McDonnell and Shannon (2001). This has been interpreted as a response to rapid differential subsidence (sagging) in the basin centre causing steeper basin margin slopes (Praeg et al., 2005; Shannon et al., 2005; Figure 3.8B). In the study area, the C30 channels likely acted as a drainage system from the Celtic Sea Shelf into the basin as suggested by their orientation (Figure 3.3). They were part of a basin-wide Eocene system also including drainage from the northern and western edges of the basin to the centre (Shannon, 1992; McDonnell and Shannon, 2001). Very similarly, an Eocene channel complex draining the continental margin off northwest Ireland was formed in the northeastern Rockall Basin and attributed to margin uplift (Georgiopoulou et al., 2021).

The upper Eocene to lowermost Miocene sedimentary sequence is above the sampled interval in well 62/7-1 (Figure 3.7) and below the depth that was reached in the IODP boreholes (Kano et al., 2007; Louwye et al., 2008). Well data in the northern Porcupine Basin, though, show that this sequence mainly consists of (calcareous) (silty to sandy) claystones (Merlin Energy Resources Consortium, 2020). In the Rockall Trough (see Figure 3.1 for location), contourite sequences have been recognised in seismic data at this level, which indicates the onset of deep-water circulation in an open marine setting created by a subsiding basin (Stoker, 1997; Praeg et al., 2005). In this scenario, the C30 reflector signifies the transition from a downslope dominated to an alongslope-dominated sedimentation pattern (Stoker et al., 2001). As this reflector was also identified in the Porcupine Basin, the same interpretation was applied there (McDonnell and Shannon, 2001). However, the contourites in the Porcupine Basin are not as well developed and have so far only been described in the northern part of the basin (McDonnell and Shannon, 2001; Shannon et al., 2005). In the study area, seismic unit U3 has a relatively uniform thickness throughout (Figure 3.4C), and unit U3 reflectors are mostly parallel and mirror the C30 topography (Figure 3.2). This suggests draping and a predominantly hemipelagic sedimentary environment (Figure 3.8C). In the northernmost part of the study area though, the lower part of unit U3 is preferentially infilling one of the C30 palaeochannels, onlapping the channel edges, while deposition on the adjacent high at this time was very thin or lacking (Figure 3.5C). This also suggests a hemipelagic environment dominated by the settling of particles, though the preferential deposition in channels indicates some alongslope redistribution of sediments by bottom currents (cf. Georgiopoulou et al., 2021; Rodrigues et al., 2022). Reflectors on top of this C30 high are deposited predominantly after the channel infill (Figure 3.5) and their mounded geometry and low to moderate amplitude suggests bottom-current influence at the top of unit U3 (Rebesco et al., 2014). This concurs with observations in the BMP, where Miocene bottom currents have been inferred from the presence of wavy geometries underneath the RD2 reflector in sparker seismic data (Van Rooij et al., 2007a). A very similar though more developed succession of seismic facies has been described in the Rockall Trough, where interfluvial areas are topographically amplified by the interaction of channel overbank and contouritic deposition (Georgiopoulou et al., 2021). Therefore, it appears that the contouritic signature is more subtle in the study area but increases northwards in the Porcupine basin and along the Irish margin of Rockall Trough. The subdued presence of upper Eocene-lowermost Miocene contourites in the Porcupine Seabight compared to the Rockall Trough has previously been attributed to the shape of the Porcupine Seabight and it being a more confined basin (McDonnell and Shannon, 2001). Thus, it was possibly less easily reached by deep-water currents compared to the open ocean setting of the Rockall Trough (McDonnell and Shannon, 2001).

Data from IODP 307 show that deposition in the middle to late Miocene consisted of homogenous silty clays (Expedition 307 Scientists, 2006), suggesting a low-energy depositional environment.

The reflectors of unit U2 are parallel and horizontal in the basin, suggesting hemipelagic deposition there. On the upper slope, however, reflectors in undulating configurations indicate the influence of bottom currents. A multitude of faults is present within unit U2, especially north of Bilbo Channel and between Meriadoc and Samwise channels (Figure 3.2, 3.5). These are similar to the polygonal faults described at the eastern edge of the Rockall Trough (Georgiopoulou et al., 2021), further accentuating the similarity of the processes occurring along the Irish margin.

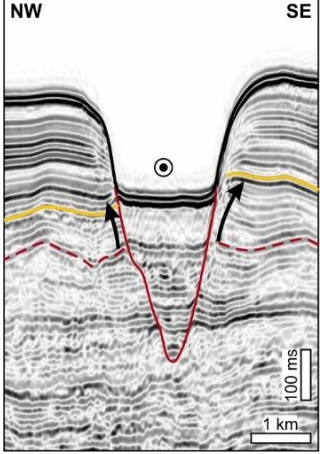
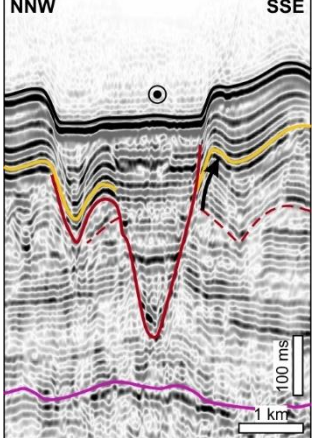
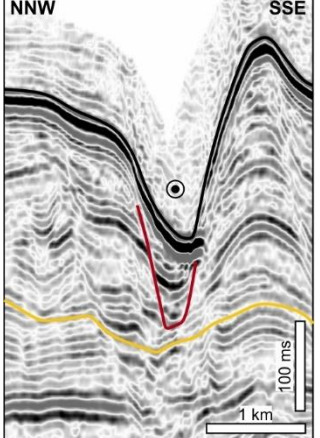
3.5.3 Initial incision of the GCS

The distinct V shape and associated truncation of reflectors at the base of the GCS channels demonstrate that erosional downslope gravity flows are the driving mechanism for the formation of the system. A tentative timing of the initial incision can be deduced from the seismic data. The divergent reflections in unit U2 in proximity to the channel bases (Figure 3.6) are interpreted as levees suggestive of the presence of an adjacent channel. Levees are identified below the RD1 reflector in several places and next to several channels, indicating that the GCS was formed before the RD1 event. There are only a few locations where a channel base cuts through the RD2 reflector, and they do so only slightly, with the erosional edges continuing to a level above RD2. Therefore, the channels were formed sometime between the early middle Miocene (RD2) and the late Pliocene (RD1) (Figure 3.8D). This is contrary to what was previously proposed, and implies that the GCS was formed before the Kings Channel System just to the north (Chapter 2; Van Rooij et al., 2009).

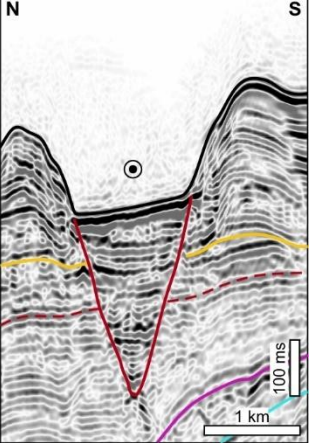
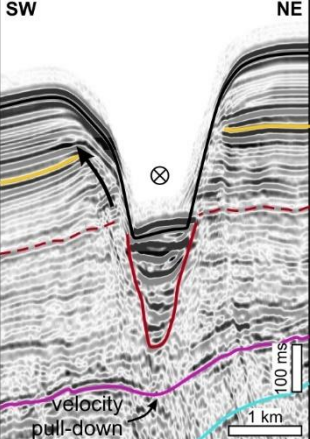
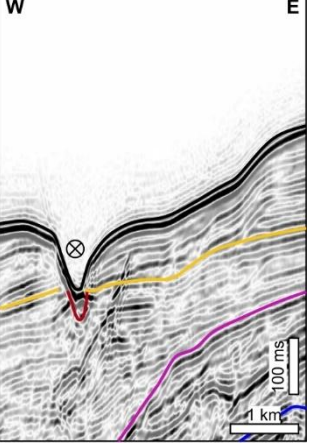
The ECP reflector is interpreted as a reflector representing early channel presence in the study area. It is situated at the base of clear levee formation on both sides of Peregrin Channel, and coincides with potential levee formation next to Bilbo, Frodo, and Meriadoc channels. This suggests synchronous (or near-synchronous) channel formation for these four channels, at least at the resolution of this dataset (Table 3.1). It is unclear whether Drogo Channel was formed at the same time as the four oldest channels, though down-dipping truncating reflectors do suggest it was present before the RD1 event (Table 3.1). The ECP reflector is situated deeper than the deepest point of incision of Lotho and Estella channels, indicating these were formed later (Table 3.1). It is possible that a low in the RD1 topography was influential on the formation of Lotho Channel, but the dataset used here is insufficient to make a definite conclusion on the timing of formation of this channel. This is also the case for Estella Channel, as only one profile crosses the channel, though in this profile, the incision appears above the RD1 reflector (Table 3.1). The same uncertainty arises for Diamond Channel, where the two crossing profiles suggest the RD1 reflector may have been erosional and shaped the channel base (Table 3.1). Evidence for when the southern branches of the system were formed is even less clear. The ECP reflector is eroded by overlying reflectors before it reaches the southern branches, and tracing the RD1 reflector to the southern channels becomes increasingly difficult as well. Nevertheless, on several profiles, the facies underneath the RD1 reflector seems continuous on either side of Samwise Channel and truncating reflectors suggest channel formation at the time of the RD1 event (Table 3.1). However, the presence of potential levees and down-dipping truncating reflectors below the RD1 reflector are arguments for the formation of Samwise Channel before the RD1 event. Bell Channel and Hamfast Channel, are likely younger than Samwise Channel and formed around the time of the RD1 event, shown by the presence of down-dipping truncating reflectors (Table 3.1). The incision of the smallest of the southern branches, Rose Channel, does not reach below the RD1 reflector, indicating this channel is younger than the RD1 event.

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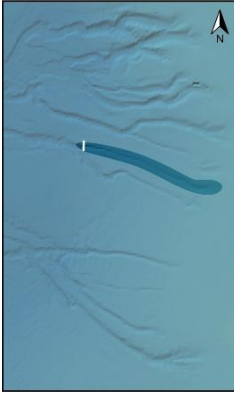
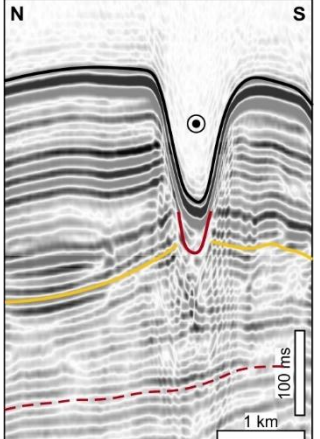
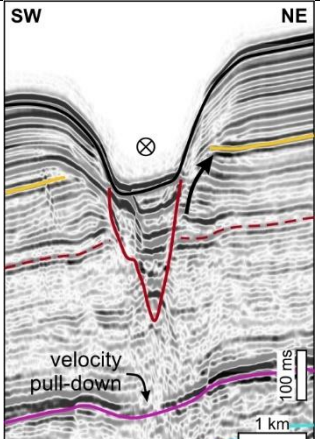
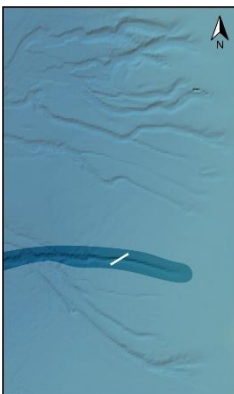
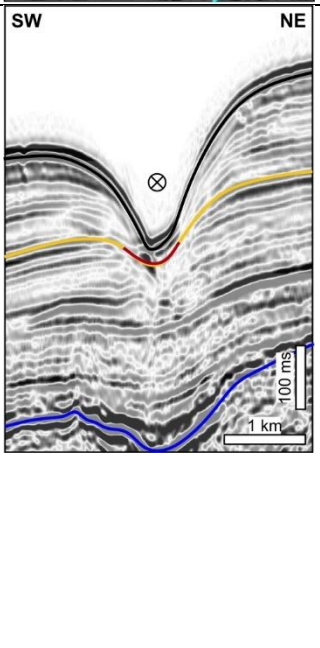
Table 3.1. Table showing the location and seismic expression of the individual channels of the GCS used in this study. Colours used to indicate key reflectors and levee formation in the interpreted seismic data are the same as in Figure 3.2. Also shown are the inferred ages of the individual channels along with the arguments used in determining them. Lastly, the total depth of each channel (D: measured vertically from the shoulder to the deepest incision) and thickness of the channel infill (F: measured vertically from the seafloor to the deepest incision) is given as average [minimum-maximum] values.

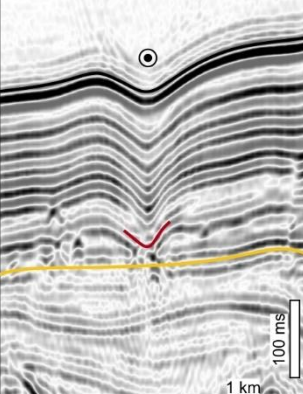
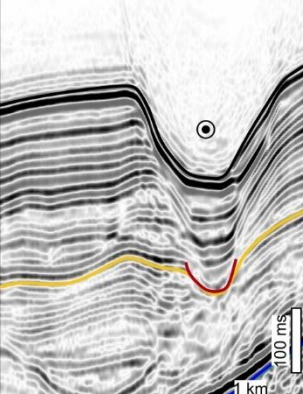
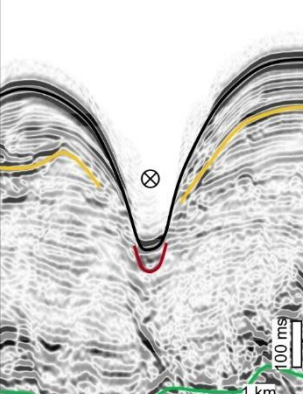
name + location	seismic expression	age + arguments	dimensions (s TWT)
Bilbo 		early Pliocene (ECP reflector) -levee formation -dipping reflectors are present below RD1 that connect to channel base	D: 0.58 [0.43-0.97] F: 0.32 [0.26-0.44]
Frodo 		early Pliocene (ECP reflector) levee formation	D: 0.45 [0.25-0.93] F: 0.27 [0.13-0.45]
Lotho 		-younger than the four oldest channels -younger than late Pliocene (RD1 reflector)? deepest incision does not reach down to the ECP reflector and in some profiles does not even reach down to the RD1 reflector	D: 0.33 [0.25-0.38] F: 0.13 [0.07-0.18]

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Drogo 		older than late Pliocene (RD1 reflector), but very unclear whether formed at the time of the ECP reflector or after -deepest incision barely below the ECP reflector in one profile and completely above it in another, but very much below the ECP reflector in all other profiles -no levee identified though	D: 0.46 [0.41-0.58] F: 0.30 [0.18-0.42]
Peregrin 		early Pliocene (ECP reflector) levee formation	D: 0.45 [0.32-0.63] F: 0.22 [0.13-0.29]
Diamond 		late Pliocene (RD1 reflector)? -continuity of reflectors on both sides of the channel below RD1 reflector -RD1 reflector dips down and truncates reflectors underneath (Figure 3.5) -only two channel crossings, which increases uncertainty	D: 0.23 [0.14-0.33] F: 0.08 [0.00-0.17]

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Estella 		younger than late Pliocene (RD1 reflector) -deepest incision does not reach down below the RD1 reflector -only one channel crossing, which increases uncertainty	D: 0.22 [0.22-0.22] F: 0.06 [0.06-0.06]
Meriadoc 		early Pliocene (ECP reflector) levee formation	D: 0.39 [0.19-0.52] F: 0.20 [0.00-0.28]
Samwise 		-older than late Pliocene (RD1 reflector)? -late Pliocene (RD1 reflector)? -dipping reflectors connecting to the channel base seem present below RD1 in some profiles -RD1 seems to dip down and truncate reflectors underneath that show continuity between both sides of the channel in other profiles	D: 0.31 [0.23-0.41] F: 0.14 [0.00-0.26]

Rose 	NNW SSE 	younger than late Pliocene (RD1 reflector) deepest incision does not reach down below the RD1 reflector	D: 0.04 [0.04-0.05] F: 0.00 [0.00-0.00]
Bell 	NNW SSE 	late Pliocene (RD1 reflector)? -RD1 reflector dips down at channel edges and truncates reflectors underneath that show continuity between both sides of the channel	D: 0.29 [0.27-0.31] F: 0.13 [0.08-0.18]
Hamfast 	SW NE 	late Pliocene (RD1 reflector) -RD1 reflector dips down at channel edges and truncates reflectors underneath that show continuity between both sides of the channel	D: 0.37 [0.21-0.47] F: 0.11 [0.00-0.26]

This discussion implies that at least five main channels (Bilbo, Frodo, Drogo, Peregrin, and Meriadoc) were formed well before the start of the Quaternary (RD1 event; Figure 3.8D). Several other channels (Diamond, Samwise, Bell, Hamfast, Lotho?) were formed later and likely just before or around the time the Quaternary started (Figure 3.8E), while at least some of the smaller channels (Estella, Rose) postdate the RD1 event and the start of the Quaternary (Figure 3.8F).

Quaternary glaciations on the Irish mainland and shelf (and associated lowered sea levels) were previously mentioned as probable sources of the sediment gravity flows that eroded the initial GCS thalwegs (Akhmetzhanov et al., 2003). The channels may indeed have been utilised and re-incised during deglaciation periods but the timing of the initial incision implies an earlier mechanism. The most likely scenario for initial incision of the system stems from passive margin tectonics. A phase of tilting across the NE Atlantic margin (from Norway to Ireland) in the early Pliocene (Praeg et al., 2005; Stoker et al., 2005b) has been associated with channel re-incision in

the Rockall Trough (Elliott et al., 2006; Georgiopoulou et al., 2021), indicating the existence and erosive nature of downslope flows at the Irish margin at this time. Tilting includes uplift of onshore and shelfal areas producing source material, and coeval basin subsidence creating accommodation space (Praeg et al., 2005). Prograding clinoforms overlying an erosional surface, recognised as indicators of margin tilting further north (Dahlgren et al., 2005; Praeg et al., 2005), are here observed in unit U2 on the shelf edge (Figure 3.5). Additionally, prograding clinoforms were identified in middle to late Miocene seismic data on the Celtic Sea Shelf east of the channel heads (Merlin Energy Resources Consortium, 2020). This may indicate margin tilting and subsequent increased sediment supply to the shelf edge between the middle Miocene (RD2) and the late Pliocene (RD1) here as well, likely originating from the same early Pliocene episode. The tilting and likely associated erosive downslope flows are a possible explanation for the incision of the GCS channel bases and the timing of the tilting (between RD2 and RD1) is in agreement with the timing of incision of the first channels suggested before. However, a direct connection between the erosional surface on the outer shelf and the channels has not been observed.

Many of the channel bases of the present-day GCS seen on the seismic profiles are located above (the edges of) large depressions in the NTC, C30 and/or RD2 surfaces (Figure 3.2, 3.3). Within the section of the study area where the C30 reflector is present, 85 % of the present-day GCS channels are located above a C30 palaeochannel. The repetition of depressions in similar locations throughout time suggests that these have been preferential pathways for downslope sediment transport in the basin since the late Eocene and in some places perhaps since the end of the Cretaceous. Reoccupation of palaeochannels leading to stacked channel fills is a known phenomenon in both terrestrial and marine settings (Elliott et al., 2006; Greene et al., 2007; Georgiopoulou et al., 2021; Horozal et al., 2021) and has been recreated in flume experiments (Reitz et al., 2010). A combination of greater compaction and higher erodibility of channel fill sediments compared to interfluvial sediments has been suggested to create topographic lows over the palaeochannels, thus favoring reincision at the same location later (Greene et al., 2007; Straub et al., 2009). However, different from these examples, the channels formed in the study area at the level of the C30 unconformity and subsequent lows at the level of the RD2 horizon are completely filled in during the late Paleogene-early Neogene. New channels are formed between the early middle Miocene (RD2) and the late Pliocene (RD1) by erosional downslope flows that did not simply reuse pre-existing depressions (Figure 3.8E). This is shown by quasi-horizontal reflectors just underneath and adjacent to the present-day GCS channel bases, with the bases cutting into those reflectors (Figure 3.2, 3.6). Therefore, erodibility and compaction of the palaeochannel fills seem less likely to have influenced the location of incision of the present-day GCS. There are also no indications for a long-term tectonic influence on repeated flow pathways. Plenty of faults are present at several stratigraphic levels (Figure 3.2, 3.5), but it is difficult to recognise faults underneath the channels due to artefacts in the data. However, any faults that were tentatively identified only have small offsets and did not create abrupt changes in topography that would lead to preferential channel formation in those locations. Additionally, any large (reactivated) basin-bounding faults are oriented north-south and perpendicular to the orientation of the channel systems (Bailey et al., 2003). A combination of (small) lateral differences in seafloor lithology and topography at the time of incision is thought most likely to have influenced channel locations.

The repetitive presence of east-west oriented channels on the eastern slope of the Porcupine Seabight also indicates that there was a long-lived sediment source on the shelf intermittently providing sediment in gravity-driven flows. Both for the C30 palaeochannels and the present-day channel system, passive margin tectonics (sagging and tilting, respectively) are seen as the main and most likely cause for increased sediment input (Praeg et al., 2005; Shannon et al., 2005). The characteristics of this source area are uncertain. The simplest interpretation would be a direct connection with a fluvial system on the Celtic Sea Shelf. While this is not apparent in the seismic

data presented here or data on the shelf investigated in other studies (Giglio et al., 2022), well data may provide some arguments for this hypothesis. At the end of the Eocene, the outer shelf was a shallow marine environment but the presence of late Eocene-early Oligocene paralic and fluvial facies on the inner Celtic Sea Shelf does suggest a narrower shelf than at present and closer proximity of terrestrial material (Merlin Energy Resources Consortium, 2020). In the early to middle Pleistocene, a fluvial to shallow marine facies with an unconformable base was deposited across the Celtic Sea Shelf (Merlin Energy Resources Consortium, 2020). It passes laterally into the open marine facies of the Porcupine basin, with the transition area on the outer shelf and upper slope near the GCS channel heads.

3.5.4 Channel processes

The evolution of the Gollum Channel System started with a canyon phase, where (possibly margin tilting-induced) erosive gravity flows produced V-shaped incisions (Figure 3.8D). Although the oldest major erosive episode has formed at least four of the channels simultaneously, the existence of younger channels implies multiple phases of incision, of which some around or after the time of the RD1 event (Figure 3.8E). Additionally, truncating reflectors at the channel edges above older channel bases suggest multiple phases of channel flank erosion after the initial incision, and especially after the late Pliocene (RD1; Figure 3.5C, 3.6). Although channel flank erosion could be due to the action of erosional bottom currents (Van Rooij et al., 2010; De Mol et al., 2011), the incision of additional channels points to several phases of re-activation of the source area. Even though glacially-derived turbidity currents are unlikely to have formed the oldest channels, they are the most likely source of sediment for the incision of new pathways after the RD1 event. The less extreme incisions of the younger channels indicate that these flows were less erosive than the flows that created the oldest channels. Some of these younger channels are later filled in and buried or draped and were seemingly active for only a short period of time (too close to larger channels causing spillover, possibly combined with sediment redirection by bottom currents), and some have eroded edges and seem to have been active for longer (Table 3.1).

Although many of the channels (especially the older ones) seem to have experienced repeated periods of activity over 5 Myr (Figure 3.6), the presence of a singular base and no lateral migration or visible re-incision suggests a single incision phase creating each of the channels. The distinct erosional phases after the initial incision suggest there were several moments in the Quaternary where the erosive nature of flows was such that they were capable of local flank erosion. Most of the downslope gravity flows entering any of the channels after their initial incision likely reused the newly created topography to flow into the Porcupine basin and either bypassed the upper slope or added to the channel fill.

Part of the sediment carried by the oldest incising gravity flows was deposited in the unit U2 levees on the channel edges. Levee growth is much reduced though not completely absent (e.g. Bilbo Channel) above the RD1 reflector, where draping dominates over active levee build-up. The (sub)horizontal, laterally continuous reflections of unit U1 suggest a dominance of hemipelagic processes (Figure 3.8F). However, there are several indications for alongslope bottom current action influencing deposition during the Quaternary.

Above the RD1 reflector, asymmetry in the levees is apparent in several channels, with a generally better-developed steeper northern levee and a smoother, less-developed southern levee (Figure 3.6). Levee asymmetry has also been described in several other canyon systems (Fuhrmann et al., 2020; Pandolpho et al., 2021) and was recreated in flume tank experiments (Miramontes et al., 2020). The phenomenon is attributed to alongslope-oriented bottom currents interfering with downslope-oriented gravity flows, with increased bottom current velocities leading to increased levee asymmetry (Miramontes et al., 2020). The levee on the downstream side of the bottom current is better developed (wider, more bedforms) due to more overspill there and inhibition of

overspill by the bottom current on the upstream channel edge (Miramontes et al., 2020). Additionally, the downstream channel flank may act as an obstacle to the bottom currents, causing erosion there and draping on the upstream channel flank (Van Rooij et al., 2010; De Mol et al., 2011). The evolution from symmetry in the U2 levees (especially visible in Peregrin and Meriadoc channels) to levee asymmetry above RD1 suggests little to no bottom current influence on the GCS in its early stages, and slightly higher influence from a northward-flowing bottom current from the RD1 event onwards (Figure 3.8E). This is in agreement with the introduction of the Mediterranean Outflow Water into the Porcupine Seabight at the time of the RD1 event. This water mass is carried northwards along the eastern slope of the basin by the European Slope Current and produces enhanced bottom currents through interaction with over- and underlying water masses (Van Rooij et al., 2003; Toucanne et al., 2021). The levees disappear towards the top of unit U1 (Figure 3.6; Chapter 2), indicating that the channels evolved to accommodate any flows running through them towards the end of the Quaternary. By this time, flows may have been completely contained within the channels, too infrequent, or not carrying big sediment loads.

3.5.5 Regional importance

There was little to no interaction with bottom currents in the initial stages of GCS existence. When the Mediterranean Outflow Water was introduced in the basin, downslope-transported sediment was thieved by northerly alongslope bottom currents. Thus, even though the channel system was formed before, the oceanographic conditions allowed sediment transport to the BMP from the start of the Quaternary at the earliest. During glacial periods, bottom currents slowed down, reducing the potential for sediment to be brought north to the BMP. Periods of deglaciation, on the other hand, caused increased sediment input into the GCS, with some gravity flows capable of eroding new pathways (and formation of a whole new channel system: the Kings Channel System; Chapter 2). Additionally, bottom current strength picked up as climatic conditions changed, potentially allowing thieving of sediments and nutrients from the channels for drift and coral mound build-up in the BMP. This process was likely halted when the channels became deep enough to accommodate flows running through them, inhibiting the thieving of sediments by bottom currents. The increasing importance of hemipelagic processes over bottom current-related deposition throughout the Quaternary visible in the GCS is also seen in the last 0.5 Myr in the BMP; glacial deposition with some muddy contourites is favored over coral growth and mound development (Huvenne et al., 2009b).

The GCS is here proposed to have been the conduit for downslope sediment transport into the Porcupine Seabight since the early Pliocene. If it was indeed fed by a fluvial system on the Celtic Sea Shelf, this might explain its initial likeness to the Bay of Biscay canyons, which were fed by the Fleuve Manche palaeoriver (Bourillet et al., 2003; Toucanne et al., 2009, 2012). A subsequent lower amount of sediment input and/or less erosive nature of downslope gravity flows in the Quaternary is a likely reason for the present-day difference between the GCS and the canyons in the Bay of Biscay. In this regard, oceanographic processes on the shelf (wind-driven currents and downwelling circulation between shelf and slope) and upper slope (internal tides) are potentially crucial (Heijnen et al., 2022).

3.6 Conclusions

From the seismic reflection data and well data presented here, it is inferred that the Gollum Channel System in the Porcupine Seabight was formed by basin margin tilting in the Pliocene. It is also shown that while all of the channels visible in the present-day bathymetry have a long history, they were not all formed at the same time. Some, the smaller ones, were formed in the Quaternary, likely by the erosive action of gravity flows related to glacial processes. The lack of lateral migration and the presence of a singular base in all channels points to one incision phase creating each channel. There were a few subsequent moments where gravity flows were capable

of (re-)incision and less extreme (channel flank) erosion. They likely mostly followed the pre-existing topography and either bypassed the upper slope or contributed to the channel fill.

The present-day channel system is underlain by an infilled and buried latest Eocene channel system, suggesting these conduits have been preferential pathways for downslope sediment transport on different occasions. There does not seem to be any structural control on the pathways, and the main influence on the location of the channels is thought to be (small-scale) pre-existing variability in bathymetry.

The influence of large-scale canyon-channel systems as sediment conduits on depositional systems and habitats elsewhere in a basin is illustrated by the link between activity in the GCS and depositional regimes in the BMP. The interaction with thieving alongslope bottom currents is crucial and enlarges the area of influence of the channel system significantly. These are processes that should also be considered and better understood in light of the distribution of terrestrially-derived pollutants.

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CHAPTER 4 Variations in downslope activity and bottom current dynamics in a land-detached submarine channel system since the Last Glacial Maximum

A modified version of this chapter is in review to be published as:

Verweirder, L., Georgiopoulou, A., Langedock, K., Fourie, F., White, M., Benetti, S., Sacchetti, F., Boone, W., and Van Rooij, D. (2025). Variations in downslope activity and bottom current dynamics in a land-detached submarine channel system since the Last Glacial Maximum. *Sedimentology*.

Abstract

The Gollum Channel System is a land-detached large-scale canyon-channel system situated offshore southwest Ireland on the Northeast Atlantic margin. The system has been considered inactive with downslope gravity flows since the last glacial period, but geophysical data in some of its branches do suggest Holocene activity. Here, the hypothesis of (in)activity under Holocene highstand conditions is tested. High-resolution side-scan sonar, photography and bathymetry data were collected using an autonomous underwater vehicle (AUV) in the upper slope section (350-1000 m water depth) of two channels in the system, Bilbo Channel and Frodo Channel. The AUV results are quantified and validated by current meter data from a mooring station in Bilbo Channel. Additionally, two sediment cores are used to build a record of channel floor processes since the Last Glacial Maximum and detect any temporal variability in channel activity. Turbidites in the sediment cores indicate that downslope sediment transport was happening well after deglaciation of the continental shelf, revealing that the channels were active under modern climatic and oceanographic conditions in the early Holocene. Crescentic bedforms in one of the channel head gullies suggest present-day activity, but their limited spatial extent indicates present-day downslope flows are minimal in frequency and volume. As a whole, the Gollum Channel System can be considered inactive today. Rather, tidally modulated bottom currents oriented along the channel axes are the main mechanism currently transporting and redistributing sediments, forming fields of megaripples and superimposed ripples. This dynamic channel floor environment has allowed the development of cold-water coral mini-mounds since the early Holocene. This study highlights that specific environmental conditions are necessary to allow frequent downslope activity in land-detached canyon-channel systems and these may vary even in systems draining the same shelf, and that inactivity may have important consequences for deep-sea habitats, by allowing or promoting their development.

Keywords bottom currents, turbidity currents, autonomous underwater vehicle, cold-water corals, Porcupine Seabight, Gollum Channel System

Author contributions L. Verweirder wrote the manuscript with input from all other authors, constructed the figures, and analysed and interpreted the data. K. Langedock, F. Fourie and W. Boone helped with AUV data acquisition and processing. M. White helped with current meter data acquisition and processing. S. Benetti and A. Georgiopoulou helped with core data acquisition and processing. F. Sacchetti helped with bathymetry data acquisition and processing.

4.1 Introduction

Submarine canyon-channel systems are important geomorphological features that facilitate the downslope transport of sediments, nutrients, and pollutants from shallow continental shelves to deep-water basins (Canals et al., 2006; Puig et al., 2014; Azpiroz-Zabala et al., 2017; Kane and Clare, 2019; Maier et al., 2019; Pierdomenico et al., 2023; Baker et al., 2024). Generally, land-detached systems are considered inactive in this regard, as they are too far away from terrestrial sources and input from, e.g., fluvial systems or long-shore drift cells in the present-day highstand conditions (Covault and Graham, 2010). On the other hand, several processes are capable of resuspending shelf and slope sediments, transporting them across a shelf and feeding them into a (land-detached) canyon system (Puig et al., 2014). These include earthquakes (e.g., Mountjoy et al., 2014), anthropogenic activities (e.g., Daly et al., 2018), storm wave loading (e.g., Puig et al., 2004), slope failures (e.g., Hsu et al., 2008), internal waves (e.g., Aslam et al., 2018), dense shelf-water cascading (e.g., Canals et al., 2006), hyperpycnal flows (e.g., Li et al., 2018),.... Recent monitoring has showed that in some land-detached systems, these processes enable a frequency and intensity of downslope activity similar to that in land-attached systems (Normandeau et al., 2020; Heijnen et al., 2022; Jaijel et al., 2023). Thus, the idea of inactivity in land-detached systems during sea-level highstands is being challenged more and more. This evokes the question whether frequent downslope activity should be expected in land-detached canyons globally or rather distinct environmental conditions allow or trigger regular activity in specific locations only. Considering c. 80 % of all large-scale canyon-channel systems are land-detached (Harris et al., 2014; Bernhardt and Schwanghart, 2021), and they are spread across all continental margins (Harris and Whiteway, 2011), the impact on our understanding of canyon system dynamics and canyon-associated deep-sea ecosystems, carbon cycle and marine geohazards, is potentially large.

To better understand the constraints on the inactivity hypothesis for land-detached submarine canyons, the Gollum Channel System (GCS) is an ideal study area. It is a large-scale land-detached system situated on the same margin as the also land-detached Whittard Canyon (Figure 4.1), where frequent present-day downslope activity has been documented (Heijnen et al., 2022). The two systems are, however, positioned in different basins and, therefore, potentially affected by different sediment dynamics and linked to different sediment sources. Whittard Canyon has been draining the Celtic Shelf and received input from the Irish Sea Ice Stream and Fleuve Manche paleoriver during the last glacial period (Bourillet et al., 2006; Heijnen et al., 2022). The GCS is thought to have drained the Celtic Shelf also, but this was never proven and the source of sediment supply to the GCS remains enigmatic (Tudhope and Scoffin, 1995; Wheeler et al., 2003). Additionally, whereas present-day downslope activity has been documented in Whittard Canyon (Heijnen et al., 2022), the GCS is considered inactive since the Last Glacial Maximum, when the British-Irish Ice Sheet extended onto the Celtic Sea Shelf (Wheeler et al., 2003; Scourse et al., 2021). Tudhope and Scoffin (1995) observed biological encrustation on dropstones and coral thickets during their submersible dives in one of the GCS tributaries, Bilbo Channel, from which they inferred no turbidity currents occurring for at least several decades before their dive. They also made visual observations of reversing bottom currents, for which they estimated maximum velocities of 0.5 m/s, and identified them as the main mechanism of present-day sediment transport (Tudhope and Scoffin, 1995). Current meter data from a 2005 mooring in Bilbo Channel confirmed the presence of tidally modulated bottom currents and did not record any downslope flows in the 185-day measurement period (Chapter 2). However, bathymetry data from the INFOMAR mapping programme of the Geological Survey Ireland and the Marine Institute (publicly available at <https://www.infomar.ie/>) reveal gullies in the channel heads of Bilbo and Frodo channels, the two northernmost channels of the system (Geological Survey Ireland and Marine Institute, 2024), suggesting possible Holocene activity.

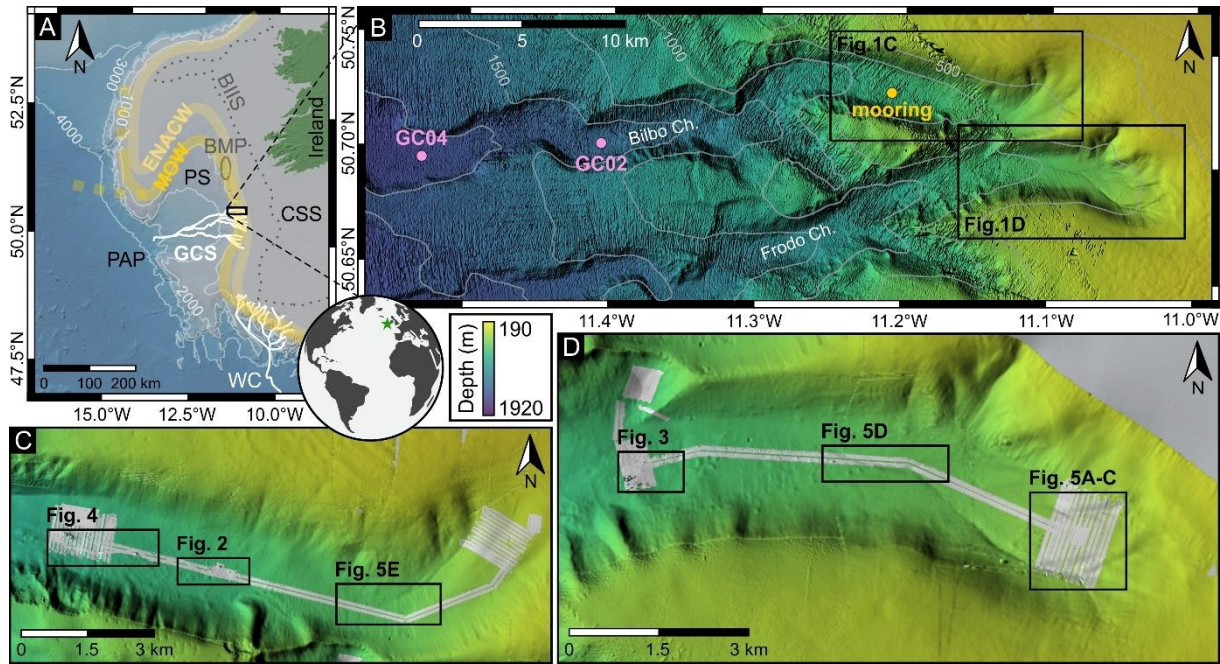


Figure 4.1. (A) Topography map showing the Porcupine Seabight (PS) on the Northeast Atlantic margin. The bathymetry is the GEBCO_2023 Grid (GEBCO Compilation Group, 2023), with contour lines drawn every 1000 m. The Gollum Channel System (GCS) is indicated in white, stretching from the Celtic Sea Shelf (CSS) to the Porcupine Abyssal Plain (PAP). Whittard Canyon (WC) is also indicated in white. The grounding line of the British-Irish Ice Sheet (BIIS) at its maximum extent (c. 26 ka) is indicated as a grey dotted line and based on Clark et al. (2022a). The Belgica Mound Province (BMP) is indicated by a grey oval. The Eastern North Atlantic Central Water (ENACW) and Mediterranean Outflow Water (MOW) are carried northwards by the European Slope Current. (B) Bathymetry map showing the sections of Bilbo Channel and Frodo Channel that are the focus of this study. The location of the mooring from Chapter 2 is indicated by the yellow dot. The locations of the two sediment cores described in this study (B2306-GC02 and B2306-GC04) are indicated by the pink dots. The bathymetry is the INFOMAR 25 m-resolution grid (Geological Survey Ireland and Marine Institute, 2024), with contour lines drawn every 250 m. (C) and (D) Close-up bathymetry maps of the upper slope sections of Bilbo Channel and Frodo Channel, respectively. The background bathymetry is a combination of datasets from campaigns CE21015, B2022-18 and B2023-12. The areas covered during the AUV dives are coloured grey. The locations of Figure 4.4-4.7 are indicated by black rectangles.

The sediment dynamics in two tributary GCS channels, Bilbo Channel and Frodo Channel (Figure 4.1), are the subject of this study. The analysis of channel floor bedforms in their upper slope sections is combined with the bottom current measurements from Chapter 2 to address the first aim of this study: identifying the main processes (if any) transporting material in the GCS at present, thus testing the (in)activity hypothesis in this land-detached system. Sediment core analysis is used to reach the second aim: testing the presence or absence of sediment transport after the Last Glacial Maximum, thus determining whether the downslope activity of the system displays temporal variability. In doing so, this work intends to improve the understanding of the dynamics in land-detached submarine canyons, as well as their potential consequences, specifically on deep-sea ecosystems.

4.2 Setting

The study area is situated on the eastern margin of the Porcupine Seabight (Figure 4.1). This embayment on the northeast Atlantic margin originated as a failed rift basin related to the opening of the North Atlantic (Shannon, 1991). It started developing in the Permian, experienced a major rifting episode in the Late Jurassic, and was then subjected to thermal subsidence from the Early Cretaceous onwards (Shannon, 1991; Naylor and Shannon, 2005; Shannon et al., 2007). It has been a continuously open marine environment since the latest Eocene-earliest Oligocene (McDonnell and Shannon, 2001; Stoker et al., 2005b), with a present-day maximum water depth of c. 3500 m (Figure 4.1A). The GCS is the only large-scale canyon-channel system in the

Porcupine Seabight. It stretches for over 200 km and crosses the basin from east to west, connecting the Celtic Sea Shelf to the Porcupine Abyssal Plain (Figure 4.1A). During the Quaternary, it is thought to have been draining the periodically ice sheet-covered Celtic Sea Shelf (Tudhope and Scoffin, 1995; Wheeler et al., 2003). In the last glacial cycle, the ice margin of the British-Irish Ice Sheet was closest to the GCS channel heads at about 26 ka, at a distance of c. 50 km (Figure 4.1A) (Clark et al., 2022a).

Although the Quaternary glaciations were highly influential in shaping the GCS, the oldest channels of the system were formed before the start of the Quaternary (Chapter 3). The first incisions were likely the consequence of a phase of northeast Atlantic margin tilting in the Early Pliocene (Chapter 3; Praeg et al., 2005; Stoker et al., 2005). Erosional events caused by the Quaternary glaciations were less severe compared to these initial incisions, but still capable of expanding the GCS with several (smaller) channels and keeping the main channels open.

Over the eastern slope of the Porcupine Seabight, the upper part of the water column below the seasonal thermocline (c. 200-700 m water depth) is occupied by Eastern North Atlantic Central Water (ENACW). The Mediterranean Outflow Water (MOW) is present underneath, with a high-salinity core at c. 800-1000 m water depth (Rice et al., 1991; White, 2007). Both water masses are carried northwards by a general poleward margin flow that is concentrated in the European Slope Current at the continental slope (Pingree and Le Cann, 1990; Rice et al., 1991; White, 2007). A permanent pycnocline is present at about 700 m water depth, at the transition between ENACW and MOW (White, 2007; White and Dorschel, 2010). At this depth, intensification of the tidal bottom currents by freely propagating semi-diurnal or bottom-trapped diurnal internal tides can be generated and create a dynamic sedimentary environment (White, 2007; White et al., 2007). As a result, the variability of the MOW's depth and strength is considered the determining factor for the growth and decay of the cold-water coral mounds and associated sediment drift in the Belgica Mound Province (Van Rooij et al., 2007a; Huvenne et al., 2009b; Wienberg et al., 2020; Matossian and Van Rooij, 2024).

4.3 Methods

This work is based on two main datasets: (1) data collected in three autonomous underwater vehicle (AUV) dives, and (2) data from two sediment cores. The processing, measurements and analyses performed on these datasets are described below. The two main datasets are complemented by multibeam bathymetry data covering the study area, and oceanographic data from a mooring located in Bilbo Channel (Figure 4.1B).

Multibeam bathymetry data were acquired in Bilbo and Frodo channels using hull-mounted systems in 2021 (RV Celtic Explorer campaign CE21015), 2022 (RV Belgica campaign 2022-18), and 2023 (RV Belgica campaign 2023-12). The RV Celtic Explorer is equipped with a Kongsberg EM302 1x2 degree deepwater multibeam system (30 kHz central frequency), while the RV Belgica is equipped with the newer EM304 1x1 degree system (20-32 kHz central frequency). Both vessels are also equipped with high accuracy Seatex Seapath 330/380 GNSS-IMU systems to provide positioning, motion referencing and timing to the multibeam systems. Regular sound velocity profiles were acquired during these surveys to compensate multibeam data for acoustic refraction. All datasets were processed and cleaned using a combination of Caris Hips&Sips and QPS Qimera hydrographic processing softwares following a standard data cleaning approach based on manual editing and automatic filtering of noisy data and outliers. The 2021 dataset was gridded to 15 m horizontal resolution, while the 2022 and 2023 datasets were combined and gridded to 25 m horizontal resolution.

The data from the mooring station in Bilbo Channel are used to validate and quantify the results from the AUV data. The mooring data were published in Chapter 2, but as they are reinterpreted

here and integral to the interpretation and discussion of the channel floor processes, a brief overview of the results is presented alongside the results from the AUV data. The mooring sat at 900 m water depth and recorded bottom currents at 8 m above the seafloor between June and mid-December 2005. Further specifics of the mooring setup and processing of the current meter data are described in Chapter 2.

Combining these four datasets allows for the identification of seabed features on the channel floors and analysis of the processes from which they are generated, as well as their temporal variability. All coordinates, maps, and distances are reported based on coordinate reference system WGS84 UTM zone 29N (EPSG: 32629).

4.3.1 AUV data

Decimetre-resolution side-scan sonar (SSS), interferometric bathymetry, and monochrome photography data were collected in three dives during RV Belgica survey 2022-18 using the Flanders Marine Institute AUV Barabas. AUV Barabas is a Gavia AUV from Teledyne and has a depth rating of 1000 m. The AUV is equipped with an inertial navigation system (iXblue C3) aided by a pressure sensor (Keller PA-30X) and a doppler velocity log (Teledyne RDI Pathfinder DVL). Positioning accuracy during the three dives was further improved by mounting an ultra-short baseline (iXBlue Gaps USBL) system on the dropkeel of RV Belgica. For each dive, all raw navigation data were postprocessed using Delph INS Subsea software and exported at 100 Hz. Thus, the navigational error of the postprocessed data was generally kept below 5 m. The operating speed was 1.7 m/s during all dives, while other operational settings were adjusted for each dive individually (Table 4.1).

Table 4.1. Overview of the three AUV dives where SSS data were collected. WD = water depth, LF = low-frequency, HF = high-frequency, bathy = bathymetry.

survey	start location	end location	start WD	end WD	channel	altitude	recorded data
1	11°05'2.53"W 50°44'19.69"N	11°14'18.12"W 50°44'19.34"N	373 m	976 m	Bilbo	20 m	LF SSS bathy
2	11°02'22.05"W 50°41'11.79"N	11°08'31.54"W 50°42'18.38"N	486 m	824 m	Frodo	20 m	LF SSS bathy
						8 m	HF SSS LF SSS bathy
3	11°11'10.17"W 50°43'35.19"N	11°14'12.03"W 50°44'21.55"N	858 m	962 m	Bilbo	4 m	HF SSS LF SSS bathy images

4.3.1.1 Side-scan sonar

SSS data were collected using a Klein 3500 module operating either in 455 kHz low-frequency (LF) or 900 kHz high-frequency (HF) mode (Table 4.1). Processing included updating the real-time positioning data in the sonar files with the postprocessed navigation data, which was carried out using NavInjector Pro software from Chesapeake Technology. Then, the data were postprocessed and compiled into mosaics using SonarWiz software, also from Chesapeake Technology. Postprocessing steps were tailored to each dataset and included a nadir transparency filter, empirical gain normalization, heading and pitch correction, and, when necessary, a destriping filter. Turns in the AUV track were not included in the final mosaic as they contain distorted data. The LF data were gridded to a resolution of 0.25 m, while the HF data were gridded to a resolution of 0.10 m.

4.3.1.2 Bathymetry

Along with the LF SSS data, the Klein 3500 module is capable of recording bathymetry through interferometry. After correcting the navigation information in the raw SSS files, the Klein Batch Processing software was used to transfer the raw transducer stave data into raw bathymetry data that are readable by commercial processing software. During postprocessing in SonarWiz software, the data were corrected for variable sound velocity in the water column obtained through the AUV's onboard sound velocity sensor (Valeport miniSVS), and filters were applied to remove faulty near-range values in the nadir. Afterwards, the bathymetry point cloud was loaded into AutoClean software from BeamworX and more faulty pings were removed with filters and, where necessary, manually. All AUV bathymetry data were gridded to a resolution of 0.5 m.

After processing, the qProf plugin in QGIS was used to construct bathymetry profiles. Bedform crests and troughs were determined visually from the profiles. Bedform morphometries were determined following the method used in Lo Iacono et al. (2020). Wave heights and wavelengths were calculated as the perpendicular distance from the crest to the bedform baseline, and the distance along the baseline between two consecutive wave troughs, respectively (Figure 4.2). The asymmetry index (AI) is the ratio of the difference of the distances between the crest and the upslope and downslope troughs, and the wavelength (Figure 4.2). Bedforms with AI values between -0.2 and 0.2 are considered symmetric, while more negative or positive values indicate upslope or downslope asymmetry, respectively.

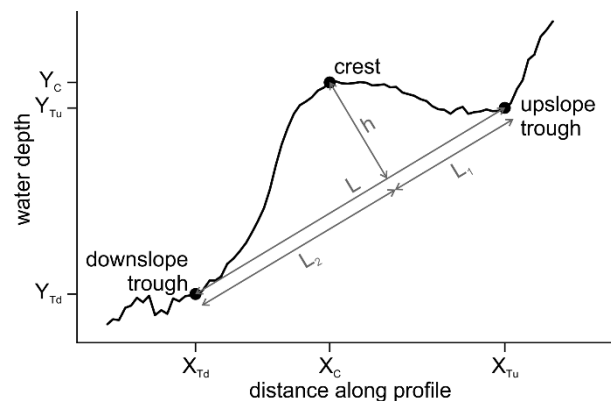


Figure 4.2. Sketch depicting the method for calculating the bedform profile morphometries: wave height (h), wavelength (L), and asymmetry index ($AI = (L_1 - L_2) / L$) (Lo Iacono et al., 2020).

4.3.1.3 Photography

During several sections of the AUV dives, the altitude above the seafloor was low enough to enable seafloor photography using the camera in the nose section of the AUV (Grasshopper GRAS-14S5M-C) (Table 4.1). Pictures were taken at a framerate of 3.75 frames per second. The images were loaded into QGIS by applying the navigation information from the AUV (altitude, depth, heading, latitude, longitude, pitch) and calculating the image size from the flying altitude and a seawater refraction index of 1.33 (calculated according to Quan and Fry (1995)) to ensure correct scaling and measurements of any features.

4.3.2 Sediment cores

Two sediment gravity cores (B2306-GC02 and B2306-GC04) were collected from the floor of Bilbo Channel during RV Belgica survey B2023-12 (Table 4.2). They were cut into 1 m-sections, capped and labelled on the ship.

Table 4.2. Details of the sediment cores used in this study. WD = water depth, L = length.

core name	type	latitude	longitude	WD (m)	setting	L (m)
B2306-GC02	gravity	50°42.155'N	11°24.488'W	1612	channel floor	2.99
B2306-GC04	gravity	50°41.664'N	11°31.853'W	1827	channel floor	3.00

Computed Tomography (CT) scanning was performed on whole cores using the Siemens medical CT scanner at Ghent University Hospital in Belgium. Dragonfly software was used for visualisation and basic analysis of the data. After opening, the cores were photographed at Ghent University using the linescan camera incorporated in the GeoTek multi-sensor core logger (MSCL) system. The MSCL-Standard of Ghent University was also used to measure the gamma-ray density, magnetic susceptibility (point sensor), and colour spectrum (reflectance, XYZ, $L^*a^*b^*$) at 0.5 cm resolution. Afterwards, the cores were described visually for their lithology, colour and structures.

Based on the CT images, visual description and geophysical datasets, the cores were sampled for grain-size analysis. Following the method described in Wils et al. (2021), a chemical pretreatment was used to isolate the terrigenous fraction from the samples by boiling the samples in 10 ml DI water with H₂O₂ (2 ml, 35 %), HCl (1 ml, 10 %) and NaOH (1 ml, 2 M) to remove organic matter, calcium carbonate and biogenic silica, respectively. Lastly, samples were boiled with sodium hexametaphosphate (1 ml, 2 %) to ensure disaggregation of the particles. Grain-size analysis was performed using the Malvern Mastersizer 3000 laser diffraction particle size analyzer at Ghent University. Each sample was measured thrice to ensure robustness and reproducibility of the results. The average distribution of each triplicate was used in further analysis.

Mixed planktic foraminifera were collected from the > 150 µm fraction of a total of five 1 cm-thick intervals over the two cores. Samples were sent to Beta Analytic (Miami, USA) for accelerator mass spectrometry radiocarbon analysis. The rbacon package (v3.3.1; Blaauw and Christen, 2011) was used to construct Bayesian age models for both cores from their respective radiocarbon ages, which were calibrated using the Marine20 calibration curve (Heaton et al., 2020) and $\Delta R = 0$, as per the protocol established by the BRITICE-CHRONO project (Clark et al., 2022b) (Appendix C). Ages for specific horizons mentioned in text and figures refer to median ages extracted from the model at the respective depths.

Both cores were scanned for their geochemical properties using an ITRAX machine at the Austrian Core Scanning Facility of the University of Innsbruck (Appendix D). Selected core intervals were measured thrice to test robustness and reproducibility and only elements where the Pearson correlation coefficient (r) for the repeated measurements was 0.75 or higher were included in the analysis: Si, K, Ca, Ti, Fe, Rb, Sr, Zr and Ba. For each core, the element intensities (in counts per seconds; cps) were centre-log transformed to reduce the unwanted effects of physical properties and matrix effects on the data (Bertrand et al., 2024). Furthermore, a partial least squares regression was performed on a subset of elements (Si, K, Ti, Fe, Rb and Zr) to predict mean grain size at the resolution of the XRF measurements (0.2 mm) across the full depths of both cores. In this procedure (Appendix D), the method described in Liu et al. (2019) was implemented in R using the pls package (v2.8-5) (Mevik and Wehrens, 2007).

4.4 Results

The multibeam bathymetry data show that the upper slope sections of Bilbo and Frodo channels have U-shaped cross-sectional geometries with flanks of 200-400 m high, and channel floors that are about 1.5 km and 0.8-1.2 km wide, respectively. Their northern flanks are slightly steeper than the southern flanks and their floors dip downslope at an average angle of 2.2° (Figure 4.1B). The mooring data, data from the AUV dives, and sediment core data are described in more detail below.

4.4.1 Bottom currents

Present-day bottom currents in the GCS have speeds that are on average 15.1 cm/s and reach maximum values of 53.7 cm/s (Figure 4.3). The current meter record is dominated by tidal currents: the M_2 (principal lunar semidiurnal; 12.42 h period), S_2 (principal solar semidiurnal; 12 h period), and K_1 (lunar diurnal; 23.93 h period) tidal ellipses are all oriented parallel to the channel axis (E-WNW) (Figure 4.3). This results in a continuous up- and downslope movement of the bottom currents in Bilbo Channel (Chapter 2). The overall residual current is oriented towards 324°N ($\approx$ NW) with a speed of 4.75 cm/s.

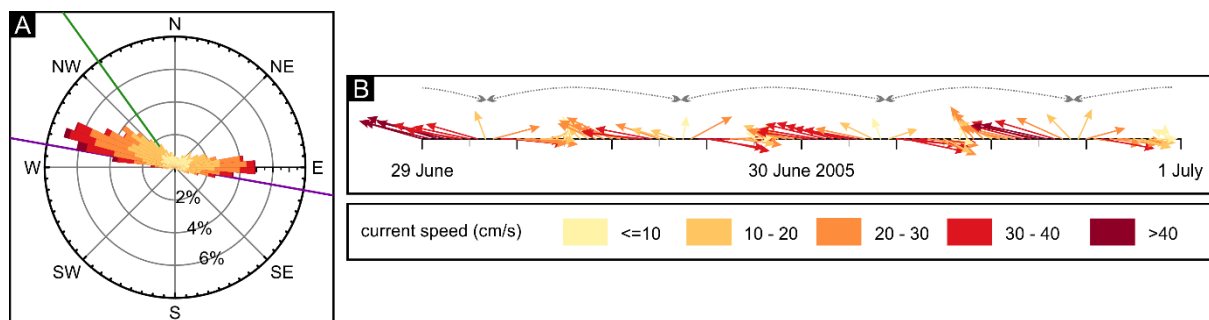


Figure 4.3. (A) Rose diagram showing the bottom current directions and speeds as a percentage of the total number of measurements recorded by the mooring in Bilbo Channel described in Chapter 2 (for location, see Figure 4.1). The orientation of Bilbo Channel is indicated in purple. The direction of the overall residual current is indicated in green. (B) Vector diagram showing a representative two-day time series of current directions and speeds just after spring tide. The grey arrows indicate semi-diurnal cyclicity.

4.4.2 Channel floor bedforms and structures

4.4.2.1 Channel head gullies

Down-channel oriented gullies were observed in the heads of both channels. They are most pronounced in Frodo Channel, where they form a dendritic network (Figure 4.4A-B). They are 10-25 m deep, 100-200 m wide, and are identified down to water depths of about 750 m. Some gullies (Figure 4.4B, D) are recognised as areas of higher backscatter intensity in the SSS data, indicating an increased roughness of the seabed there compared to the surrounding channel floor. Most gullies, however, have a SSS intensity that is similar to the surrounding seabed (Figure 4.4B).

The gullies in Bilbo Channel were not covered during the AUV dives. The SSS data just downslope of where they are located, however, display high-reflectivity linear features (ripples?) that are in line with the upslope gullies (Figure 4.4E).

In one of the gullies in the head of Frodo Channel, a train of wavy bedforms was identified. They are crescentic in shape with heights of 1-2 m and wavelengths of 10-15 m (Figure 4.4C). Another section of this gully further downslope contains two scarps that are each about 1 m high (Figure 4.4D).

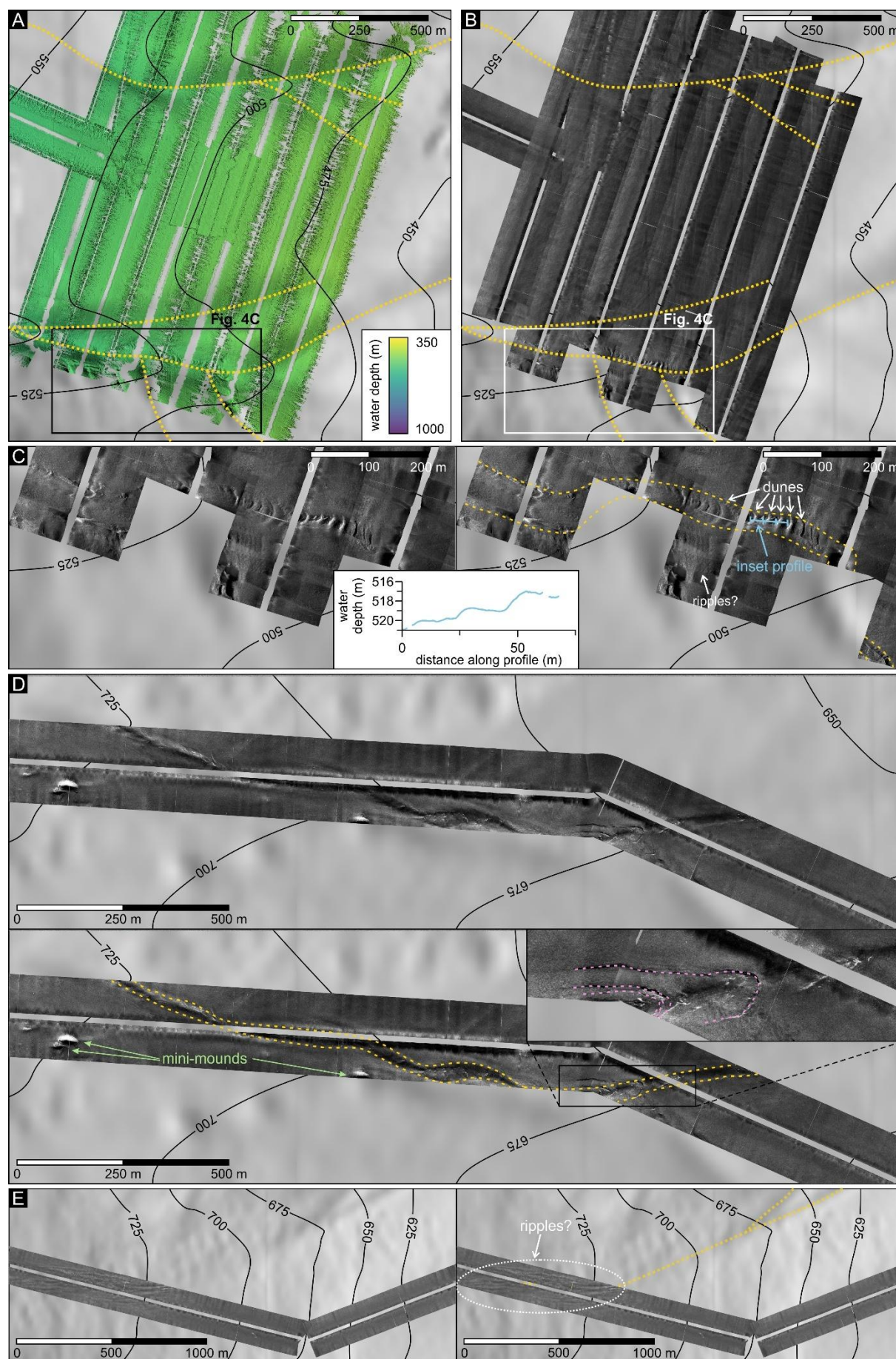


Figure 4.4. (previous page) Close-up maps showing gully features in Frodo (A)-(D) and Bilbo (E) channels. Yellow dotted lines indicate the locations of gullies. In SSS data, whiter values indicate higher backscatter intensities. The background bathymetry is the 25 m-grid from the RV Belgica 2022-18 and 2023-12 campaigns, from which depth contours were extracted every 25 m and indicated in black. (A) Bathymetry and (B) low-frequency SSS data in the head of Frodo Channel. The location of Figure 4.4C is indicated by a black/white rectangle. (C) Close-up of the low-frequency SSS data in (B) (left) and their interpretation (right). A train of dunes and potential ripples are indicated in white. The inset shows a bathymetry profile through the train of dunes. (D) Low-frequency SSS data showing a section of a gully in Frodo Channel between c. 650 and 730 m water depth (upper panel) and its interpretation (lower panel). Cold-water coral mini-mounds are indicated with green arrows, and three scarps are indicated as pink dashed lines. (E) Low-frequency SSS data of Bilbo Channel between c. 600 and 740 m water depth (left) and their interpretation (right). Features (ripples?) at the downslope extent of this gully are indicated in white.

4.4.2.2 Wavy bedforms

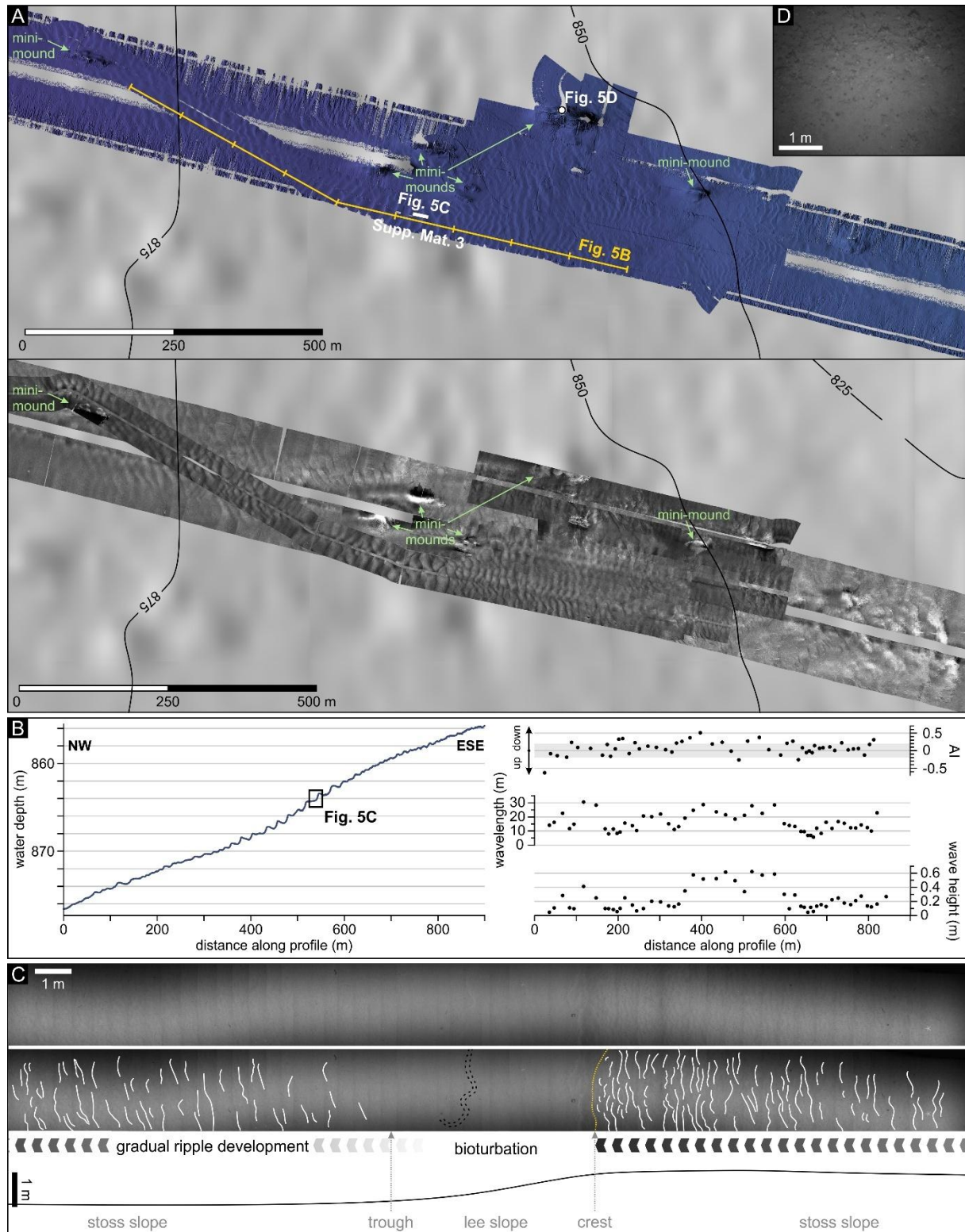
Wavy bedforms were identified on the channel floors of both channels, in water depths of 840-885 m (Bilbo Channel) and 750-850 m (Frodo Channel) (Figure 4.5, 4.6). Based on their dimensions (Table 4.3), they are classified as megaripples (van Dijk et al., 2020). The megaripple crests are straight to sinuous and some have developed bifurcation (Figure 4.5). Most megaripples in Bilbo Channel are symmetric. About a third of the measured bedforms is downslope asymmetric, whereas upslope asymmetry is rare (Figure 4.5B). The gentler (stoss) side of the asymmetric megaripples is characterised by higher backscatter compared to the steeper (lee) side (Figure 4.5A). In Frodo Channel, ripplefield A is characterised by megaripples with upslope asymmetry (Figure 4.6C) and stoss sides that have a lower intensity in the SSS data (Figure 4.6B). The megaripples in ripplefield B, on the other hand, are mostly symmetric or downslope asymmetric and upslope asymmetry is rare (Figure 4.6C). There is no difference in backscatter character for either flank (Figure 4.6B).

Table 4.3. Dimensions of the megaripples in Bilbo and Frodo channels shown as [average (minimum; maximum)]. AI = asymmetry index.

channel	wavelength (m)	wave height (cm)	crest orientation	AI
Bilbo	15.9 (5.7; 30.6)	23 (5; 62)	N-S	0.1 (-0.6; 0.5)
Frodo	10.2 (4.5; 19.3)	21 (3; 59)	N-S/NNW-SSE	-0.1 (-0.6; 0.5)

In the photographs, smaller ripples were identified superimposed on the Bilbo Channel megaripples (Figure 4.5C). They have wavelengths of about 20 cm and their straight to sinuous crests are oriented parallel to the crests of the megaripples. Wave heights cannot be measured from the top-down photographs, but the ripples do look most pronounced when located on a megaripple crest. They often disappear for 0-5 m just over the crest on the megaripples' lee side, after which they show up again over the remainder of the lee side, the megaripple trough and the stoss side of the next megaripple (Figure 4.5C).

Figure 4.5. (next page) (A) Close-up map showing AUV bathymetry (upper panel) and SSS data of the Bilbo Channel axis between 830 and 880 m water depth. A field of megaripples is visible, as well as cold-water coral mini-mounds indicated by green arrows. The bathymetry profile shown in (B) is indicated in yellow, with markers every 100 m. The strip of images shown in (C) is indicated as a white rectangle. The location of the photograph in (D) is indicated as a white dot. The background bathymetry is the 25 m grid from the RV Belgica 2022-18 and 2023-12 campaigns, from which depth contours were extracted every 25 m and indicated in black. (B) Graphs showing the water depth of the bathymetry profile indicated in (A), and the wave height, wavelength and asymmetry index (AI) of the megaripples it crosses. Megaripples of AI between -0.2 and 0.2 (indicated in grey) are considered symmetric. (C) A strip of stitched-together monochrome images recorded by the AUV over the megaripple field in Bilbo Channel (upper panel) and their interpretation (lower panel). The interpreted figure shows superimposed ripple crests in white, a megaripple crest in yellow, and a track produced by deep-water fauna in black. (D) Black-and-white photograph over the flank of one of the cold-water coral mini-mounds.



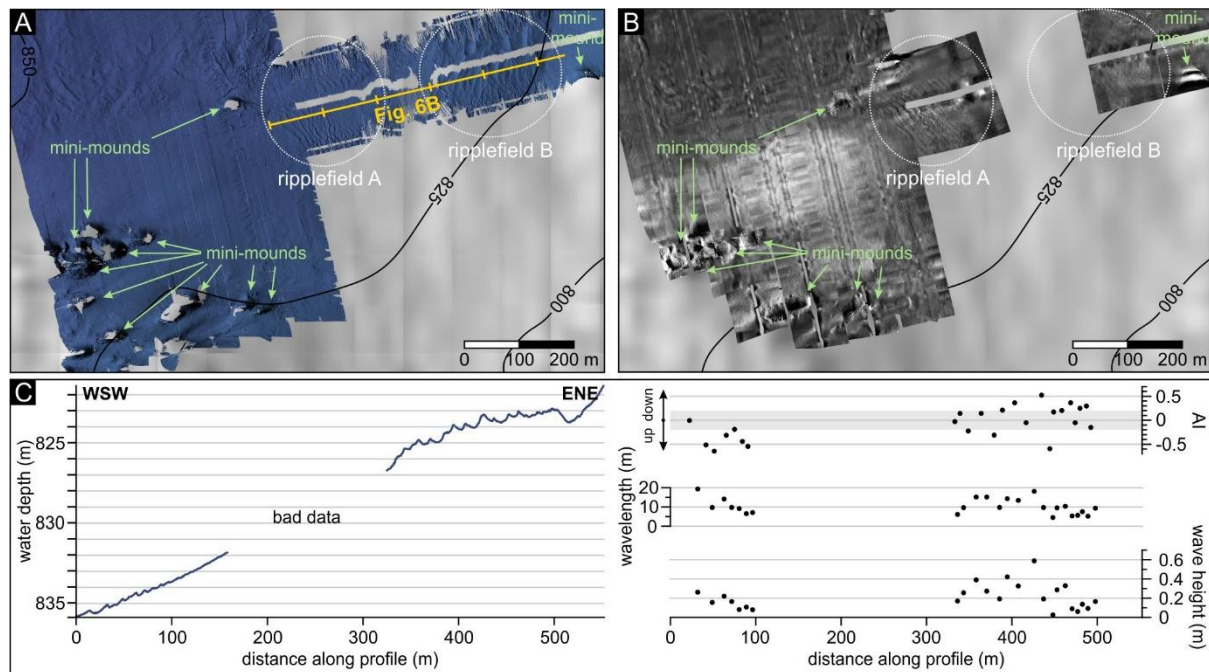


Figure 4.6. Close-up maps showing AUV bathymetry (A) and low-frequency SSS (B) of the Frodo Channel axis between c. 820 and 850 m water depth. In the SSS data, whiter values indicate higher backscatter intensities. Two fields of megaripples are indicated in white, and cold-water coral mini-mounds are indicated by green arrows. The bathymetry profile shown in (C) is indicated in yellow, with markers every 100 m. The background bathymetry is the 25 m-grid from the RV Belgica 2022-18 and 2023-12 campaigns, from which depth contours were extracted every 25 m and indicated in black. (C) Graphs showing the water depth of the bathymetry profile indicated in (A), and the wave height, wavelength and asymmetry index (AI) of the megaripples it crosses. Megaripples of AI between -0.2 and 0.2 (indicated in grey) are considered symmetric.

4.4.2.3 Cold-water corals

In the bathymetry data, mounded features are identified in water depths of 840-1000 m and 680-840 m in Bilbo and Frodo channels, respectively (Figure 4.5 and Figure 4.6, respectively). They are oblong structures oriented along the channel axis that are on average 56 m long, 27 m wide, and 7 m high (ranges 30-105 m, 12-48 m, and 2-17 m, respectively). Their backscatter character shows high reflectivity in the SSS data, indicative of coarse texture (Figure 4.6, 4.7). The seabed photos show that these are cold-water coral mounds (Figure 4.7D-G). Based on their dimensions, they are called mini-mounds, similar to examples from other locations (Masson et al., 2003; Foubert et al., 2011; Stewart et al., 2014; Collart et al., 2018). In the black-and-white images, white parts of the coral framework are interpreted as living coral, whereas the darker (greyer) parts are interpreted as dead framework or rubble (Hebbeln et al., 2014).

At c. 1000 m water depth in Bilbo Channel, a considerably bigger coral structure is present (Figure 4.7A-C). It is elongated along the channel axis, c. 600 m long, a little over 100 m wide, and about 30 m high. The corals in this particular structure are organised in subparallel patches oriented perpendicular to the channel axis (north-south). The patches are c. 0.4 m high and spaced about 9.8 m apart on average (ranges 0.1-1.3 m and 4.4-20.1 m, respectively).

Additional parallel, coarse-textured, high-backscatter areas are present on the southern edge of the SSS data in Bilbo Channel (c. 930-985 m water depth; Figure 4.7A). They are recognised as elevated areas in the bathymetry data. As they were only partially covered by the AUV data, however, extracting any meaningful dimensions is not possible.

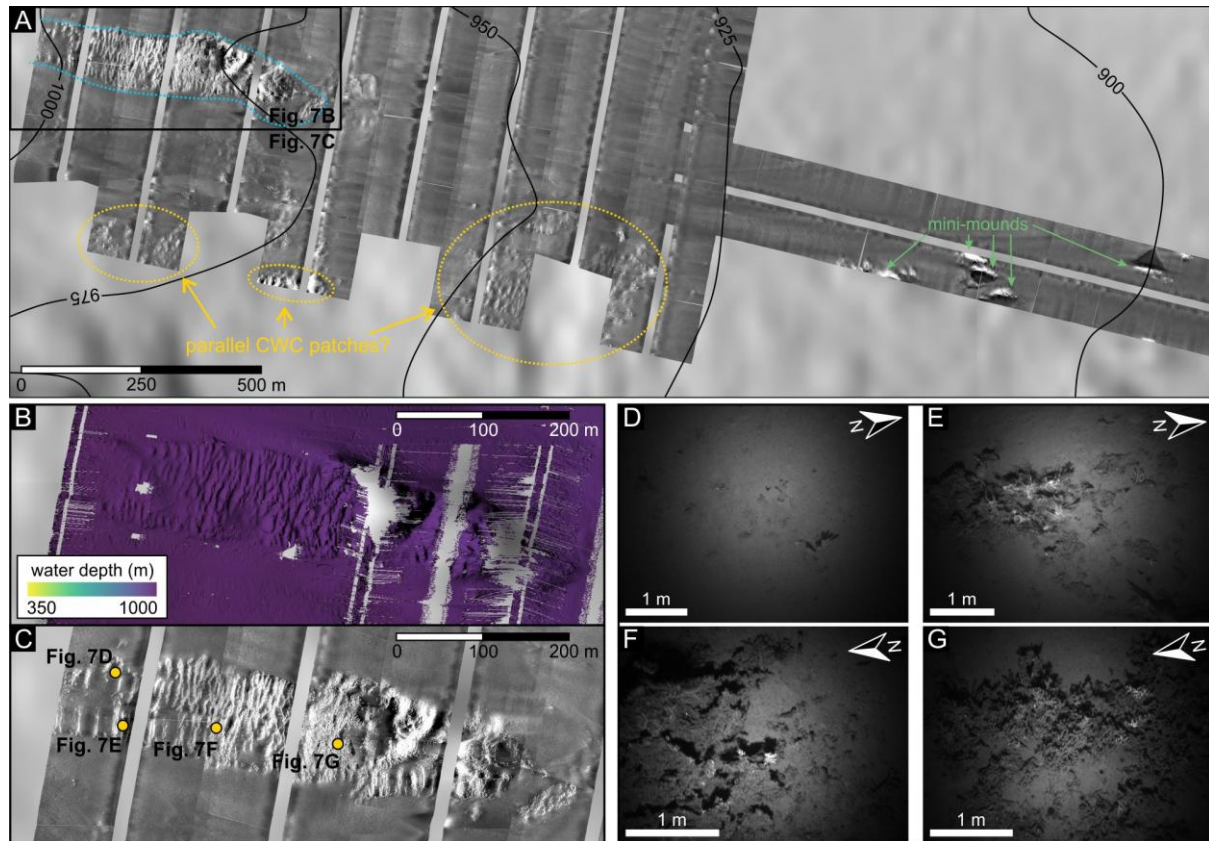


Figure 4.7. (A) Close-up map showing low-frequency SSS data of the Bilbo Channel axis between c. 890 and 1000 m water depth. Whiter values indicate higher backscatter intensities. The large cold-water coral structure is indicated in blue; cold-water coral mini-mounds are indicated by green arrows; lined-up parallel cold-water coral patches are indicated in yellow dotted circles. The location of Figure 4.7B and Figure 4.7C is indicated as a black rectangle. The background bathymetry is the 25 m-grid from the RV Belgica 2022-18 and 2023-12 campaigns, from which depth contours were extracted every 25 m and indicated in black. (B) AUV bathymetry of the large cold-water coral mound in Bilbo Channel. (C) Low-frequency SSS of the large cold-water coral mound in Bilbo Channel. Whiter values indicate higher backscatter intensities. The locations of the images shown in (D)-(G) are indicated in yellow. (D)-(G) Monochrome images recorded by the AUV flying over the large coral mound shown in (B) and (C).

4.4.3 Sediment cores

The results of the accelerator mass spectrometry analysis are given in Table 4.4.

Table 4.4. Results of the accelerator mass spectrometry radiocarbon analysis, with conventional radiocarbon ages given alongside the calibrated ages reported as [average (95 % range minimum – 95 % range maximum)].

Lab code	Core	Depth (cm)	Material	Analysed weight (mg)	¹⁴ C age (yr BP)	Calibrated age (cal yr BP)
Beta - 699043	B2306-GC02	78-79	Mixed planktic foraminifera	5.9	2810 ± 30	2410 (2172 – 2674)
Beta - 699044	B2306-GC02	237-238	Mixed planktic foraminifera	5.1	8370 ± 30	8760 (8526 – 9005)
Beta - 699045	B2306-GC04	104-105	Mixed planktic foraminifera	4.8	5720 ± 30	5960 (5703 – 6236)
Beta - 699046	B2306-GC04	168-169	Mixed planktic foraminifera	8.6	14540 ± 40	16490 (15521 – 16928)
Beta - 699047	B2306-GC04	232-233	Mixed planktic foraminifera	9.4	17300 ± 50	19860 (18891 – 20379)

4.4.3.1 B2306-GC02

The 299-275 cm interval in core B2306-GC02 has a very dark greyish brown colour and appears chaotic, with a grey (very dark greenish grey) mud clast and dipping laminations (inset Figure 4.8). Shell fragments are present as well, and this section is bioturbated. According to the age model, the age at the top of the disturbed interval is c. 10400 cal yr BP (Figure 4.8). From 275 cm to 254 cm core depth, < 1 mm to 1 cm thick intervals with distinct upper and lower boundaries are visible both to the naked eye, as darker (very dark grey) and coarser sections than the background mud, and in the CT scan as sections of higher intensity. They have a very fine sand lithology that shows a fining-upward trend in the thickest layers (Figure 4.8) and most layers are deformed or interrupted by bioturbation. The three thickest layers can be correlated with a peak in magnetic susceptibility and density values, as well as increased Zr, Si, Ca and Sr values in the XRF data (Appendix E). According to the age model, the uppermost layer was deposited c. 9600 cal yr BP (Figure 4.8). Most of the core (254-8.5 cm core depth) consists of bioturbated foraminifera-rich silt (Figure 4.8). Within this unit, the colour gradually changes from dark grey at the bottom to dark greyish brown, greyish brown, and finally light brownish grey at the top. The upper 2.5 cm of core B2306-GC02 (8.5-6 cm core depth) are greyish brown in colour, with an orange hue, and are silty with a fine sand component.

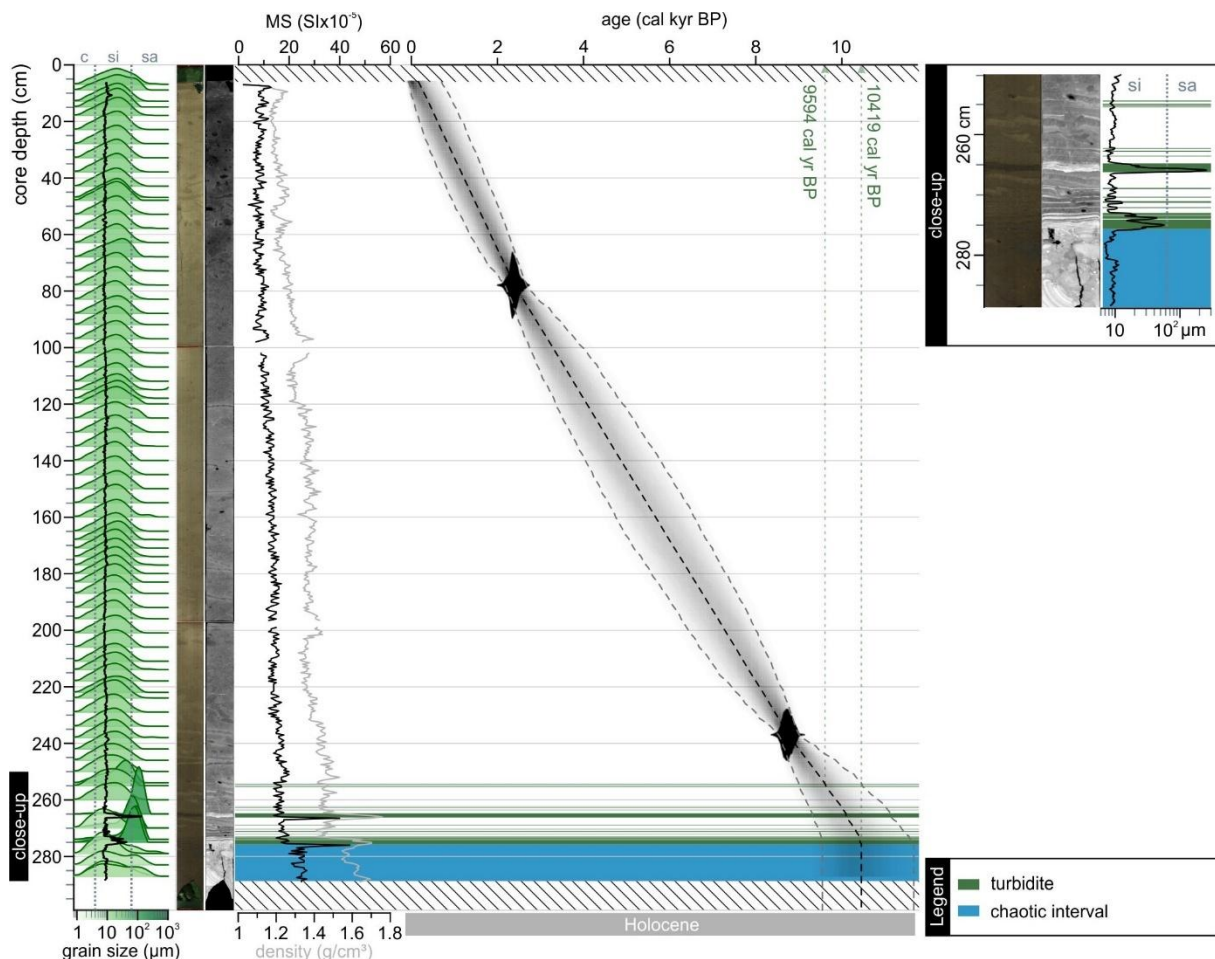


Figure 4.8. Core B2306-GC02. From left to right: grain-size distribution plots coloured according to the mode (with darker green representing a coarser mode) and mean grain size predicted from XRF data plotted in black; core photograph; CT scan; magnetic susceptibility; gamma-ray density; age-depth model. The inset represents a close-up of the photograph, CT scan, predicted mean grain size and identified facies for the 250-289 cm interval.

4.4.3.2 B2306-GC04

Between 290 cm and 172 cm core depth, core B2306-GC04 is characterised by horizontal to distorted and dipping layers, with pieces of gravel scattered throughout. Most of the layers are c. 1 mm to several millimetres in thickness, and, where visible to the naked eye, sandier and darker coloured than the background sediment, except for two greenish grey layers at 262-260 cm and 177 cm core depth (Figure 4.9). These two layers have very slightly coarser lithologies and coincide with peaks in the Ca curve and dips in the magnetic susceptibility data (Figure 4.9). The other layers have varying characteristics in the magnetic susceptibility, density, CT, and XRF data (Figure 4.9, Appendix F). The interval at the very bottom (290-263 cm) appears chaotic. Above the layered interval, a structureless very dark greyish brown silty sand crosscut by four burrows is present between 172 and 161 cm core depth. It transitions into a chaotic-looking sandy silt interval (161-152 cm core depth; inset Figure 4.9) that contains sediments of grey, black, beige, and brown colours in distorted layers and pieces of gravel, and is capped by a layer of very fine sand. At 152 cm core depth, a distinct but irregular erosional boundary is present that represents a hiatus of c. 5300 years between c. 16.0 and c. 10.7 cal kyr BP (Figure 4.9). The upper part (152-0 cm core depth) of core B2306-GC04 is made up of bioturbated foraminifera-rich beige silt (Figure 4.9). Within this section, the colour changes from dark greyish brown at 152 cm core depth to greyish brown, grey, and eventually light brownish grey at the top of the core.

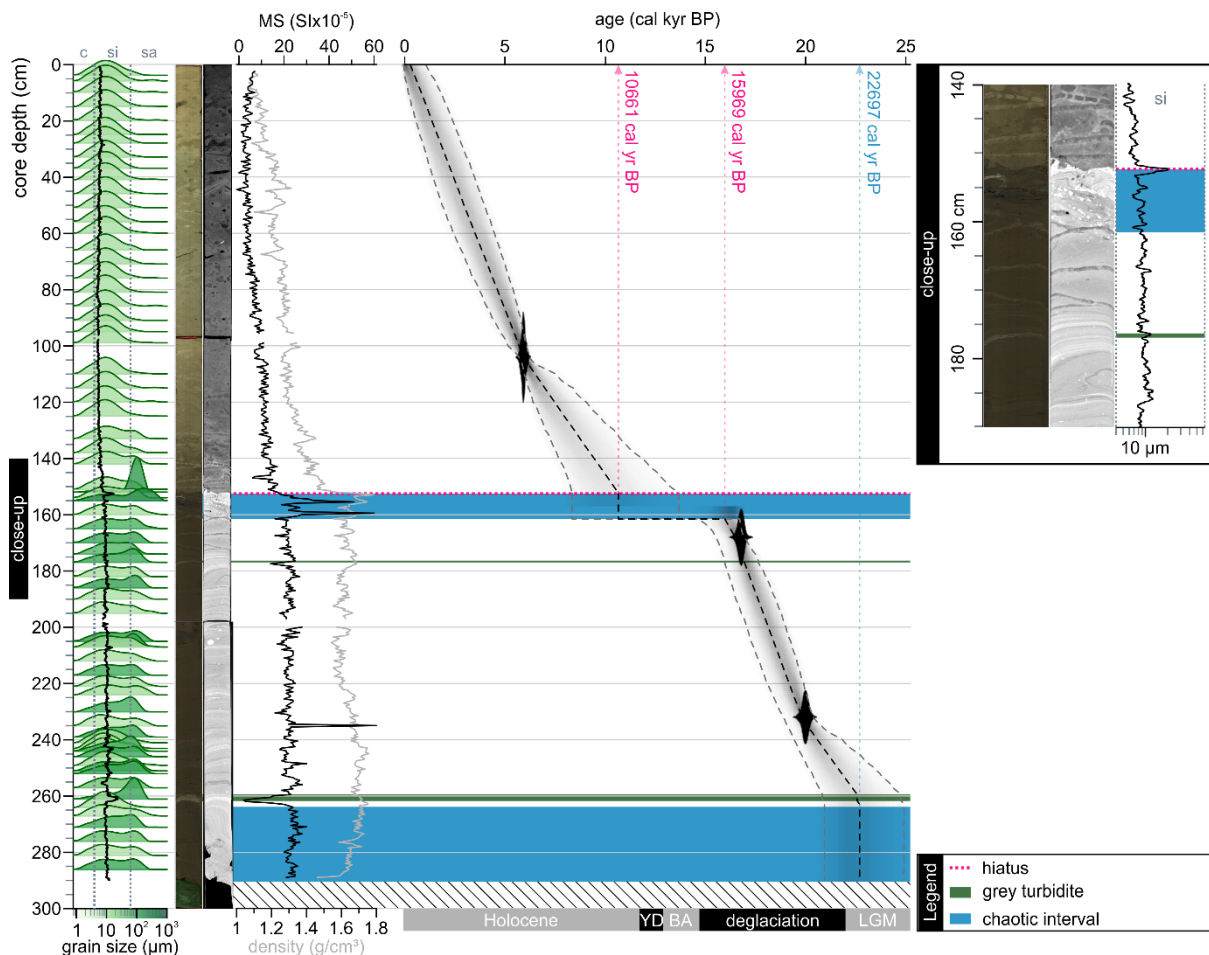


Figure 4.9. Core B2306-GC04. From left to right: grain-size distribution plots coloured according to the mode (with darker green representing a coarser mode) and mean grain size predicted from XRF data plotted in black; core photograph; CT scan; magnetic susceptibility; gamma-ray density; age-depth model. The inset represents a close-up of the photograph, CT scan, predicted mean grain size and identified facies for the 140-190 cm interval. YD = Younger Dryas; BA = Bølling-Allerød; LGM = Last Glacial Maximum according to Clark et al. (2022b).

4.5 Discussion

4.5.1 Present-day predominance of tidally modulated bottom currents

4.5.1.1 Indications from (mega)ripples

The Bilbo Channel current meter (for location, see Figure 4.1) recorded tidally modulated bottom currents that are strongly oriented along the channel axis (E-WNW), and, therefore, perpendicular to the N-S orientation of the megaripple crests in both channels and the N-S orientation of the superimposed ripples in Bilbo Channel (Figure 4.3, Figure 4.5). Flume tank experiments where a unidirectional current (or combined wave-current regime) is imposed onto an initial setup of a flat, sandy bed for a prolonged time (several hours) result in an equilibrium configuration where large-scale (asymmetric) wavy bedforms (bars, sandwaves, or dunes) are covered by superimposed ripples and both the primary structures and the superimposed ripples have crests perpendicular to the flow direction (Cataño-Lopera and García, 2006). The symmetric to slightly downslope-asymmetric geometries of most of the megaripples observed here (Figure 4.5, Figure 4.6) suggests the bidirectional (rather than unidirectional) current action oriented along the channel axis is influential for megaripple geometry (Cataño-Lopera and García, 2006). Considering the channel floor sediments are mostly silty sands (Tudhope and Scoffin, 1995), the measured bottom current velocities are capable of transporting the channel floor sediments and building the straight to undulatory ripples visible in the AUV images (Figure 4.5) (Stow et al., 2009). The superimposed ripples are likely the ripples that Tudhope and Scoffin (1995) observed during their submersible dives and attributed to reversing bottom currents. Higher energies are probably required for megaripple formation and migration (Passchier and Kleinhan, 2005; Stow et al., 2009; Foubert et al., 2011; Lo Iacono et al., 2020). This implies temporal variations in the sediment dynamics in the channels, where the megaripple fields are formed and active only during periods of peak velocities of the tidal bottom current. Small-scale spatial variations, on the other hand, are suggested by the upslope-asymmetric megaripples in ripplefield A, which likely represent a local phase of strengthened up-canyon currents and sediment transport.

Despite the current measurements suggesting that the megaripples were not actively migrating during the measurement period (cf. Stow et al., 2009), the present-day channel floors can be considered a dynamic environment. Tudhope and Scoffin (1995) noticed formation of new ripples in Bilbo Channel during an experiment that spanned a period of 21 days. Active migration of the superimposed ripples is also supported by the current meter measurements and the ripples distribution pattern over the megaripples presented here. The superimposed ripples are larger and more abundant on the megaripples crests than in the troughs and may be absent on the megaripples lee sides (Figure 4.5C). This is consistent with the SSS intensities increasing crestwards over the megaripples stoss sides and lower intensities on their lee sides. This pattern has been recognised in both experimental and field data (Cataño-Lopera and García, 2006; Li and King, 2007; Guerrero and Guillén, 2020) and is caused by the topography of the primary bedform producing a larger velocity over its crest (Cataño-Lopera and García, 2006). Flow separation may occur over the crest of the primary bedform depending on the lee side angle (Kwoll et al., 2016; Zomer et al., 2021), leading to the creation of a high-turbulence, low-horizontal velocity eddy and the absence of bedforms there (Venditti et al., 2005; Naqshband et al., 2014; Herbert et al., 2015). Superimposed bedforms are known to contribute to the growth and migration – and, therefore, maintenance – of their primary bedform (Venditti et al., 2005; Reesink and Bridge, 2007, 2009). As the superimposed ripples migrate over the stoss side of the primary bedform, they eventually reach its crest and produce grain flows over its lee side. Therefore, migration of the superimposed ripples in the study area facilitates (slow) migration of the megaripples.

The downslope asymmetry of the superimposed ripples may be explained by the asymmetry in the bottom current direction and strength. About 40 % of all current measurements were oriented towards the west to northwest (downslope), while about 30 % were directed towards the ESE to

ENE (upslope) (Figure 4.3). Additionally, over 70 % of the currents with speeds > 30 cm/s were directed downslope (Figure 4.3). It is worth noting, however, that current intensities in the Porcupine Seabight are known to experience seasonal differences in strength (Rice et al., 1991). Thus, the current meter record of the Bilbo Channel mooring is potentially not long enough to capture the full cycle of current variations and the asymmetry in bottom current direction and strength might even itself out over time. Tudhope and Scoffin (1995) observed both upslope- and downslope asymmetrical ripples, as well as symmetrical and interference ripples. This variation in geometries suggests that small-scale processes may determine the (a)symmetry of the superimposed ripples.

4.5.1.2 Indications from cold-water corals

Based on their dimensions, the oblong structures containing cold-water corals observed in Bilbo Channel and Frodo Channel are classified as cold-water coral mini-mounds such as those reported as part of the Belgica Mound Province (Moirs Mounds; Foubert et al., 2011), on the interfluvial next to the Explorer and Dangeard canyons in the upper reaches of the Whittard Canyon System (Stewart et al., 2014), surrounding the head of Ferrol Canyon on Ortegal Spur (Collart et al., 2018), as part of the Darwin mound area in the northern Rockall Trough (Masson et al., 2003; Huvenne et al., 2009a), in the Mingulay Reef Complex (De Clippele et al., 2017), in the Træna Deep Coral reef field (Cathalot et al., 2015), on Galicia Bank (Somoza et al., 2014), offshore Morocco (Glogowski et al., 2015), and on Great Bahama Bank (Correa et al., 2012). They tolerate a large range of oceanographic conditions, and their occurrence seems mainly controlled by food supply, which in turn depends on primary production near the water surface and processes that bring the food particles to the corals living at greater depths (Freiwald et al., 2004; Duineveld et al., 2007; Wienberg and Titschack, 2015; Hennige et al., 2021). The very high-resolution AUV data presented here allow to observe these structures both in detail and within their channel floor context for the first time. Tudhope and Scoffin (1995) observed small (0.1-1.0 m) colonies of living corals and patches of dead corals (c. 25 m diameter) in their submersible dives in Bilbo Channel but do not mention any mounded structures. Within canyon-channel systems, (framework-forming) cold-water corals are often found as patches or communities on steep and/or rocky outcrops in canyon flanks (Mortensen and Buhl-Mortensen, 2005; Orejas et al., 2009; Tyler et al., 2009; De Mol et al., 2011; Huvenne et al., 2011; Morris et al., 2013; Brooke and Ross, 2014; Sánchez et al., 2014), and rarely, as seen here, in mounded structures on the channel floor (Sánchez et al., 2014). Rocky canyon walls provide hard substrates for settling and intersect with the dilute sections of gravity currents that provide food particles and where energies are not so high that they sweep away any growing colonies (Huvenne et al., 2011). The channel floors of the GCS are littered with dropstones from Quaternary glaciations (Tudhope and Scoffin, 1995) that can serve as points for settling and initial attachment of coral polyps (Huvenne et al., 2005; Roberts et al., 2006). The presence of living sections of coral (Figure 4.7) suggests the absence of erosive gravity currents in that part of the channel at present.

Coral growth is further aided by the oceanographic conditions. The depth interval at which the mini-mounds occur in the GCS (c. 700-1000 m) corresponds to the water depths where the upper part of the MOW is situated and transitions into the ENACW above. At these depths in the Belgica Mound Province further north in the basin (Figure 4.1A), large-scale cold-water coral mounds are present in dynamic bottom current regimes created by bottom trapped diurnal waves rectified by the topography (Rice et al., 1991; White and Dorschel, 2010; Wienberg et al., 2020). Similarly, bottom currents are here suggested to create favourable conditions for settling and growth of the cold-water corals observed in the GCS channel axes. Settling substrates such as hardgrounds, dropstones or coral rubble are often created or maintained by enhanced bottom currents. The bottom currents also prevent any (hemi)pelagic sediments from draping over them. Once settled, the filter-feeding corals need a steady supply of food particles to grow (Dullo et al., 2008). Here,

a consistent transport of particles over the seafloor is likely promoted by the continuous up- and downslope movement of the bottom currents. These currents may also prevent sediment from accumulating on and burying the corals. Although it is possible that the bottom currents (with semi-diurnal signal) originate from the freely propagating internal waves that are present at the interface between the ENACW and the MOW (e.g., Mulder et al., 2012; Aslam et al., 2018; Antsey et al., 2024), further measurements should be undertaken to verify.

Typically, in an environment dominated by a unidirectional bottom current, cold-water coral mounds and patches tend to grow into the dominant current, which often results in an elongated to raindrop shape (Cathalot et al., 2015; Corbera et al., 2022). Flume-tank experiments show this may be due to a high particle encounter rate on the up-current side of the structure, after which the flow is deflected over the structure, leaving corals in its wake devoid of food particles (Corbera et al., 2022). This is indeed the case for the mini-mounds in Bilbo and Frodo channels, where the long axis of the mounds is oriented along the orientation of the tidal bottom currents (Figure 4.3, 4.5).

The same mechanism may lead to self-organisation within cold-water coral colonies and explain the formation of the large coral mound consisting of parallel coral ridges in Bilbo Channel (Figure 4.7B-C) (Mienis et al., 2019; Corbera et al., 2022; van der Kaaden et al., 2023). At a certain distance behind one coral structure, the original flow conditions are re-established and a new structure can grow, which could lead to a series of parallel coral patches at regular distances from one another (Corbera et al., 2022). This has been observed on the scale of fields of coral reefs (Correa et al., 2012), but is here shown on the scale of a single mound. In the case of Bilbo Channel, the higher mound relief on the upslope side, and the higher number of living corals observed there, then imply that the downslope-directed current is dominant for the growth of this mound (Messing et al., 1990). From their flume tank experiments, Corbera et al. (2022) estimate that flow re-establishment downstream occurs at a distance of 15-20 times the height of the coral structure, which fits with the heights (average 0.4 m) and distances (average 9.8 m) measured on the parallel ridges here. Another hypothesis for the formation of parallel ridges of cold-water corals is the corals occupying ripples or sandwaves (Wheeler et al., 2005; Foubert et al., 2011). However, neither the images, nor the bathymetry show any indication of bedforms underneath or surrounding the large coral mound (Figure 4.7). Hence, the first hypothesis is preferred. The partly-imaged coral patches to the south and southeast of the large coral mound (Figure 4.7A) seemingly form comparable parallel, north-south-oriented structures and may represent a similar but less developed stage of coral mound formation.

4.5.2 Temporal variation in downslope activity

All layers described in core B2306-GC02 and the two grey-coloured layers described in core B2306-GC04 are interpreted as turbidites based on their very fine sand (i.e., coarser than background) lithology, fining-upwards trend in grain size, distinct lower boundaries, and distinct to gradual upper boundaries (Figure 4.8, 4.9). Additionally, their colour differs from the colour of the background sediments, indicating a different sediment source.

For the other layers in core B2306-GC04, the interpretation of the depositional process is less straightforward. Their limited thicknesses (1 to several mm) and dipping geometries makes grain-size analysis challenging, but where measurable, the layers have a coarser lithology than the background brown muds. The high-resolution grain-size model does not display clear fining-upwards trends within the layers, and no identifiable erosive lower boundaries were present in the CT data. Some of the layers are coloured differently than the background sediments, they show up as either higher or lower intensities in the CT data, and they do not have a consistent signal in the XRF data, implying a sediment source that is different from the background sediments, but not consistent for all layers. Bottom currents are a known process transporting

channel floor sediments capable of producing layered deposits (Normandeau et al., 2024). During a large part of the time interval where the layers were deposited (> 22.7 kyr BP-16.0 kyr BP: end of Last Glacial Maximum and deglaciation), however, it is thought that bottom currents over the eastern slope of the Porcupine Seabight were too sluggish to transport sand-sized grains larger than 100 μm (Wienberg et al., 2020). Additionally, core B2306-GC04 is located at a water depth of 1830 m, which – even when accounting for lowered sea levels – is well below the depths where intensified (tidal) bottom currents occur in the basin (c. 700 m) (White, 2007; White et al., 2007). Therefore, although no clear sedimentological arguments can be found to link the layers in core B2306-GC04 to downslope gravity flows, they are thought the most likely depositional process producing these layers as well.

This implies that downslope gravity flows were active during the Last Glacial Maximum and early deglaciation and reached > 35 km down into Bilbo Channel. They remained active until c. 9.6 kyr BP, as implied from the age of the youngest turbidite according to the age model of core B2306-GC02 (Figure 4.8). Since the gravity flows carried enough sediment to deposit layers of 1 mm-2 cm thickness at 25-35 km along the channel, it was likely not possible for cold-water corals to survive – let alone build up mounded structures – in the upslope channel sections during the time they were active. It is, therefore, thought that the mini-mounds are younger than 9.6 kyr, which is consistent with other records from the NE Atlantic region (Wienberg and Titschack, 2015). For the corals to build mini-mounds of 2-17 m high (7 m average), this would require average aggradation rates of c. 21-177 cm/kyr (73 cm/kyr average). This is within the range of aggradation rates calculated for several mounds in the Porcupine Seabight (Frank et al., 2009).

There is no evidence for downslope flows during the c. 6.5 months for which the Bilbo Channel mooring current meter record is reliable. The main indicators for present-day downslope activity are the branched gully systems in the heads of both Bilbo and Frodo channels. In the southernmost branch of the gully system in Frodo Channel, the train of crescentic bedforms (Figure 4.4C) along with the presence of the scarps further downslope (Figure 4.4D) are indicative of present-day downslope activity (Paull et al., 2010; Normandeau et al., 2014).

Crescentic bedforms with similar shapes, orientations and dimensions located on canyon floors are often interpreted as cyclic steps (Cartigny et al., 2011; Babonneau et al., 2013; Normandeau et al., 2014; Symons et al., 2016; Chen et al., 2021). These upstream-migrating bedforms are the result of hydraulic jumps in downslope flows, with deposition in subcritical flow on the upstream side of the step and erosion in supercritical flow on the downstream side of the step (Kostic et al., 2010). Thus, cyclic steps can be identified based on the presence of hydraulic jumps within downslope flows, or – more commonly, since flow analyses are more challenging than bedform analyses – their upstream-migrating character (Kostic et al., 2010). Here, however, without any flow measurements or information about the internal structure of the described features, it is impossible to confirm their origin as cyclic steps.

Alternatively, if the flows responsible for creating the crescentic bedforms are subcritical, and the bedforms' internal structure shows downstream migration, these features would be interpreted as barchan dunes rather than cyclic steps (Symons et al., 2016; van Dijk et al., 2020). Barchan dunes are generally associated with flow velocities of 0.6-1.2 m/s (Stow et al., 2009), which is within the range of values recorded by direct measurements of downslope gravity flows in submarine canyon-channel systems but lower than the maximum velocity of the most powerful flows (Khripounoff et al., 2012; Azpiroz-Zabala et al., 2017; Paull et al., 2018; Normandeau et al., 2020; Heijnen et al., 2022; Hill and Lintern, 2022; Liu et al., 2023).

Most of the gullies, however, have backscatter characteristics that are not distinguishable from the surrounding seafloor (Figure 4.4B). This is interpreted as a sign of present-day inactivity and burial of all but one gully by hemipelagic sediments (with a potential bottom-current reworking

component). Along with the limited downslope extent of the gully systems, as well as the presence of cold-water coral mounds containing living coral, and the absence of turbidites in the 9.6-0 kyr BP record of the two channel floor cores, this indicates that present-day downslope activity is low in volume and frequency. The Gollum Channel System can be considered inactive with downslope flows at present. Thus, the GCS can be considered inactive with downslope flows at present.

Due to the spatial limits of the datasets presented here and the lack of data available on the outer Celtic Sea Shelf (Giglio et al., 2021, 2022; Arosio et al., 2023), it is difficult to establish the origin of the downslope flows. Two of the chaotic facies in both cores (299-275 cm in core B2306-GC02 and 161-152 cm in core B2306-GC04; Figs. 8, 9) may be interpreted as slump deposits, and slide scars are visible on the channel flanks (Fig. 1; Chapter 2), which would point to slope failures of the channel flanks or in the channel heads as potential sources for sediment transport. Another potential source mechanism that could be explored further are channelised systems that were mapped on the inner and mid-shelf and interpreted as conduits for glacial meltwater from the British-Irish Ice Sheet (Giglio et al., 2021, 2022; Arosio et al., 2023). A connection between the meltwater features and the shelf edge has not been established. Lastly, oceanographic processes may create turbidity near the seafloor and promote sediment resuspension, such as near-inertial waves created by wind stress at the ocean surface (e.g., Chinn et al., 2012; Jordi et al., 2016). This calls for further research on outer shelf glacio-marine and oceanographic processes, both in past cold periods and the present-day environment.

4.6 Conclusion

In this study, bedforms in the upper slope sections of the two northernmost channels in the land-detached Gollum Channel System were analysed to identify which, if any, processes are actively transporting sediment over the channel floors. Tidal bottom currents, possibly intensified by internal waves at the transition between the ENACW and the underlying MOW, are the main active process and create ripples superimposed on megaripples. Present-day downslope activity is inferred from a train of crescentic bedforms in one of the channel heads. However, the resulting bedforms (gullies, scarps, crescentic bedforms, ripples) have a very limited spatial extent in the upper parts of the channels (< 700 m water depth), and, therefore, downslope flows are thought to be a low-frequency, low-volume process. Thus, the Gollum Channel System can be considered generally inactive at present.

This contrasts with the channel activity during the Last Glacial Maximum and deglaciation, and into the early Holocene, which was recorded as turbidites in two sediment cores on the channel floor. They indicate that gravity flows extended at least 35 km into the channels to > 1830 m water depth from > 22.7 cal kyr BP to 9.6 cal kyr BP. The present-day combination of the absence of downslope flows and the continuous movement of sediments and food particles provided by the tidally modulated bottom currents allows the development of cold-water coral mini-mounds on the wide channel floors. The onset of mini-mound growth is thought to coincide with the Holocene inactivity of the Gollum Channel System c. 9.6 cal kyr BP. This demonstrates the importance of (in)activity and bottom current dynamics in canyon-channel systems for deep-sea habitats.

The combination of the very high-resolution AUV data and sediment core records presented here reveals that the Gollum Channel System was active under modern climatic and oceanographic conditions in the early Holocene, after the deglaciation of the continental shelf and the detachment of the system from terrestrial sources. The processes that allowed this early Holocene activity are yet unknown. The contrast between the Gollum Channel System (currently inactive) and the Whittard Canyon (currently active) situated less than 200 km apart and draining the same continental shelf indicates that, although land-detached canyon-channel systems can

indeed experience frequent present-day activity, specific environmental controls are required and should be investigated for each individual system.

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CHAPTER 5 Open-slope contourites record variability of Mediterranean Outflow Water during the last glacial cycle in the Porcupine Seabight, offshore Ireland

Abstract

The European Slope Current (ESC) is an integral part of the Atlantic Meridional Overturning Circulation as it carries warm and saline Mediterranean Outflow Water (MOW) from the Strait of Gibraltar northwards along the West-European continental margin (< 2000 m). Hence, variability in ESC flow strength and MOW output have been an essential part of the ocean-ice interactions during Quaternary glacial-interglacial climatic changes. This study investigates the record of MOW variability through the last glacial cycle at the distal end of the ESC in the Porcupine Seabight offshore southwest Ireland. Five sediment cores were acquired from open slopes between the branches of the presently inactive Gollum Channel System, where bottom currents are known to have influenced sedimentation. A combination of four mid-slope cores (water depth 1200-1300 m) and one upper-slope core (water depth 700 m) allows bottom current reconstruction at the core of the MOW flow as well as its upper boundary. Three lithofacies representing different bottom current energies are identified through geophysical, geochemical, and grain-size analyses. These indicate that a low-energy environment with hemipelagic deposition was present across the slope at the end of the last deglaciation (c. 17-16 kyr BP). It was followed by intensified bottom currents producing silty contourites at mid-depth in the Bølling-Allerød and Younger Dryas. Deposition on the upper slope remained hemipelagic for longer and gradually transitioned to muddy and then silty contourites by the Younger Dryas. This is interpreted as the introduction of the MOW in the basin by the Bølling-Allerød, and a subsequent shoaling of the water mass until it reached its present-day water depth in the early Holocene. From then on, deposition is stable with muddy contourites on the middle slope and silty contourites on the upper slope, where the density difference between the MOW and the overlying Eastern North Atlantic Central Water produces internal waves and enhanced bottom currents. This study highlights the importance of the ESC on sedimentation across the Irish continental margin and shows that for palaeoceanographic reconstructions, open slope sediment records may be complementary to contourite drifts, where enhanced bottom currents may create hiatuses or condensed sections. The results fit in basin-wide findings showing that the Atlantic Meridional Overturning Circulation strengthens and weakens during interstadial and stadial periods, respectively.

Keywords contourite, European Slope Current, internal wave, sortable silt, Gollum Channel System

Author contributions L. Verweirder analysed and interpreted the data, produced the figures and wrote the manuscript after discussions with her supervisors and S. Benetti.

5.1 Introduction

Bottom-current action greatly affects sedimentation on continental slopes through erosion, entrainment, transport and deposition of particles (Cacchione et al., 2002; Rebesco and Camerlenghi, 2008; Stow et al., 2009; Rebesco et al., 2014; McCave, 2017). Bottom currents can be enhanced at the core of geostrophic currents or at the interface between two water masses by processes such as internal waves and tides or other instabilities created by the density contrast there (Cacchione et al., 2002; Pomar et al., 2012; Rebesco et al., 2014; Zhang et al., 2016), and further by seabed topography (Puig et al., 2004b; Hernández-Molina et al., 2008; White and Dorschel, 2010; Hanebuth et al., 2015; Aslam et al., 2018). As the variability in bottom current strength is therefore inherently linked to sedimentary processes, the resulting sediment deposits are often studied to reconstruct past water mass movements (i.e., palaeoceanography) (Øvrebo et al., 2006; Raddatz et al., 2011; Hanebuth et al., 2015; Petrovic et al., 2019; Vandorpe et al., 2023).

In a palaeoceanographic context, the Mediterranean Outflow Water (MOW) has received a lot of attention as it forms a significant contributor to the global thermohaline circulation and has had a particularly varied history with its connection to the thermohaline circulation interrupted and re-established several times (Hernández-Molina et al., 2014b; Bahr et al., 2015; Ng et al., 2021; Toucanne et al., 2021). This intermediate water mass is formed at the outlet of the Mediterranean Sea at the Strait of Gibraltar and carried northward along the West-European continental margin (< 2000 m water depth) by the European Slope Current (ESC) (Toucanne et al., 2021) (Figure 5.1A). As the ESC is an integral part of the Atlantic Meridional Overturning Circulation, the variability of its flow strength and, by consequence, the variability in MOW output throughout the Quaternary is an essential part of the ocean-ice interactions during glacial-interglacial climatic changes (Rogerson et al., 2012b; Hernández-Molina et al., 2014a; Toucanne et al., 2021). Due to its importance for ocean circulation, the Quaternary evolution of the MOW has been studied extensively in sediment drifts close to its origin in the Gulf of Cádiz, along the Iberian margin, and in the Bay of Biscay (Llave et al., 2006; Toucanne et al., 2007, 2021; Rogerson et al., 2012b; Hanebuth et al., 2015; Kaboth et al., 2016; Lebreiro et al., 2018; Petrovic et al., 2019; Sierro et al., 2020; Moal-Darrigade et al., 2022; Depuydt et al., 2024).

Further north along the ESC flow path over the eastern slopes of the Porcupine Seabight (offshore Ireland; Figure 5.1B), bottom current energy at intermediate depths is still strongly influenced by the presence of the MOW. More specifically, internal waves and tides created at the interface with the overlying Eastern North Atlantic Central Water (ENACW) enhance bottom current dynamics (Pingree and Le Cann, 1990; White, 2007; White and Dorschel, 2010). The variability in bottom current strength has been critical in allowing or restricting the development of cold-water coral mounds and associated sediment drift bodies in the basin (Van Rooij et al., 2003, 2007a; Huvenne et al., 2009a; White and Dorschel, 2010; Wienberg et al., 2020; Matossian and Van Rooij, 2024). Additionally, internal wave action in the ENACW-MOW transition zone is thought to have created a field of sediment waves south of the Kings Channel System (Chapter 2), and the northward-directed residual flow over the branches of the Gollum Channel System (GCS) influences their geometry: southern channel edges are characterised by sediment draping creating gentle shoulders, whereas currents impinging on the northern channel edges may cause erosion there (Chapter 3; Wheeler et al., 2003). Moreover, channel floor bottom currents in the GCS are not only strengthened and tidally modulated at the MOW-ENACW interface but also forced in an across-slope (i.e., along the channel axis) orientation due to the channel topography, allowing cold-water coral habitats to establish there as well (Chapter 4). Any variations in the strength of the MOW flow as well as the density contrast with the ENACW are thus of major importance for the sedimentary environment and associated benthic ecosystems in the Porcupine Seabight.

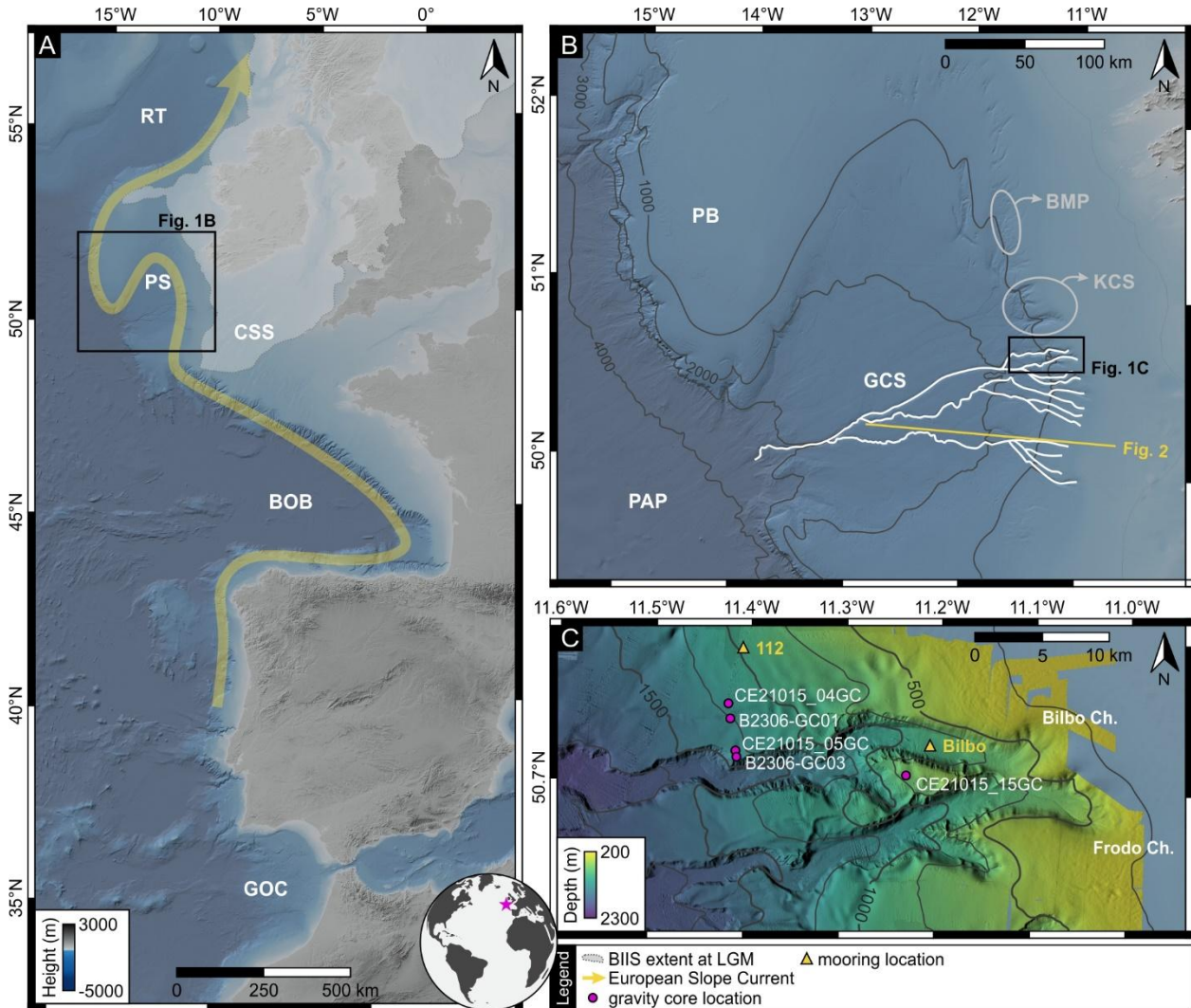


Figure 5.1. (A) Bathymetry map of the Northeast Atlantic margin, showing the European Slope Current in yellow. The maximum extent of the British-Irish Ice Sheet at the Last Glacial Maximum (LGM; ca. 26 ka) is indicated in white based on Clark et al. (2022). The bathymetry map is from GEBCO Compilation Group (2023). (B) Location of the study area offshore southwest Ireland. The Gollum Channel System (GCS) is indicated in white. The location of the profile in Figure 5.2 is indicated in yellow. The bathymetry map is from GEBCO Compilation Group (2023), with contour lines drawn every 1000 m. (C) Close-up bathymetry map (Chapter 4) of the study area: the upper slope sections of Bilbo Channel and Frodo Channel. Cores used in this study are indicated as pink circles. Two moorings are indicated as yellow triangles: one in Bilbo Channel (Chapter 2) and mooring 112 of Pingree and Le Cann (1990) on the open slope north of Bilbo Channel. Contour lines are drawn every 250 m. BMP = Belgica Mound Province, BOB = Bay of Biscay, CSS = Celtic Sea Shelf, GCS = Gollum Channel System, GOC = Gulf of Cádiz, KCS = Kings Channel System, PAP = Porcupine Abyssal Plain, PB = Porcupine Bank, PS = Porcupine Seabight, RT = Rockall Trough.

Previous palaeoceanographic studies in the Porcupine Seabight have looked into reconstructing fluctuations in ESC flow strength and vertical shifts of the MOW at the start of the Quaternary (Raddatz et al., 2011) and the last deglaciation (Wienberg et al., 2020) in the Belgica Mound Province, and during the late Quaternary on the western margin of Porcupine Bank (Øvrebø et al., 2006). Bottom current energy in the Belgica Mound Province, however, is known to be enhanced and erosion has dominated since 22.1-12.2 kyr BP (depending on the water depth) (Wienberg et al., 2020), which created hiatuses that restrict palaeoceanographic reconstructions. Similar limitations are present in the cores on Porcupine Bank, which contain condensed post-Last Glacial Maximum sections and an erosion surface at the base of the Holocene where the c. 17-12 kyr BP interval is removed (Øvrebø et al., 2006). Thus, a detailed palaeoceanography from the end of the last glacial phase until now is missing on this margin.

This study, therefore, focuses on the open slopes surrounding the northern branches of the Gollum Channel System, where (palae)oceanography is known to influence deposition (Chapter 2; Chapter 3; Wheeler et al., 2003), but bottom currents are markedly slower than in the Belgica Mound Province (cf. mooring 112 of Pingree and Le Cann, 1990; mooring Bilbo of Chapter 2) and sedimentary records are expected to be more complete. Hence, the goal of this work is to assess MOW variability in the last glacial cycle and improve knowledge of the late Quaternary palaeoceanography for this distal part of the ESC, and accordingly, the broader thermohaline circulation in the North Atlantic.

5.2 Study area

The study area is part of the continental slope between Bilbo and Frodo channels, the northern branches of the Gollum Channel System on the eastern side of the Porcupine Seabight basin. This basin is situated on the Northeast Atlantic continental margin off the Irish southwest coast (Figure 5.1) and is a failed rift associated with the opening of the North Atlantic that started forming in the Permian (Shannon, 1991; Shannon et al., 2007). Rifting ended in the Late Jurassic, after which thermal subsidence from the Early Cretaceous onwards continued to deepen the basin, which is now filled with up to 10 km of sediments (Shannon, 1991; McCann et al., 1995; Naylor and Shannon, 2005; Shannon et al., 2007).

In the late Quaternary, the sedimentary environment on the slopes surrounding the Gollum channels is dominated by hemipelagic settling influenced by the action of northward-directed bottom currents (Chapter 3; Wheeler et al., 2003). Within the channels, gravity flows are active during glacial periods and subsequent deglaciations (Wheeler et al., 2003). Turbidites have been identified from the last glacial phase until the early Holocene (c. 9 kyr BP) but since then, gravity flows are much reduced in volume and frequency, and currently limited to the channel heads (Chapter 4; Kenyon et al., 1998). Tidal bottom currents create bedforms at 700-1000 m water depth and hemipelagic deposits are dominant below 1000 m water depth (Chapter 4).

The water column in the Porcupine Seabight is characterised by Eastern North Atlantic Central Water (ENACW) at 200-700 m below the seasonal thermocline, and a high-salinity core of Mediterranean Outflow Water (MOW) at 800-1000 m (Figure 5.2). There is a permanent pycnocline at c. 700 m water depth, where high-intensity bottom currents are created by internal waves and tides (Pingree and Le Cann, 1990; White, 2007). Underneath the MOW, Labrador Sea Water (LSW) is identified as a low-salinity, high-oxygen water mass centred around 1700-2100 m water depth (Figure 5.2). At present, the European Slope Current (ESC) transports water masses at depths up to c. 2000 m northwards along the European margin between c. 40°N and 60°N (Toucanne et al., 2021). This contour current is responsible for creating thick contourites on the continental margins along its path (Laberg et al., 2005; Hernández-Molina et al., 2011). During glacial periods, an equivalent slope current (Glacial Eastern Boundary Current) was present and brought recirculated northern-sourced waters along the West-European margin (Toucanne et al., 2021). The Glacial Eastern Boundary Current is thought to have been most vigorous during Dansgaard-Oeschger stadials and slower during interstadial periods (Toucanne et al., 2021).

The study area is situated on the continental slope north of Bilbo and Frodo channels at c. 700-1300 m water depth. According to the present-day water mass distribution, the shallower part of the study area is located at the transition between the MOW and the overlying ENACW, while the deeper part is bathed by the lower part of the MOW (Figure 5.2). Although the study area is located next to the two northernmost branches of the Gollum Channel System, the large channel depth (300-350 m) and minimal recent downslope activity (Chapter 4; Tudhope and Scoffin, 1995; Wheeler et al., 2003) makes any influence of channel-related downslope processes on the surrounding slope sediments unlikely. Additionally, the seafloor topography in both locations is smooth and escarpments, landslide scars or other features indicative of downslope transport or

slope failure are absent (Figure 5.1). The sediment record on the slope is thus thought to represent deposition from bottom currents and hemipelagic settling of grains and will be used to reconstruct the bottom current strength from the MOW core flow as well as the ENACW-MOW transition zone.

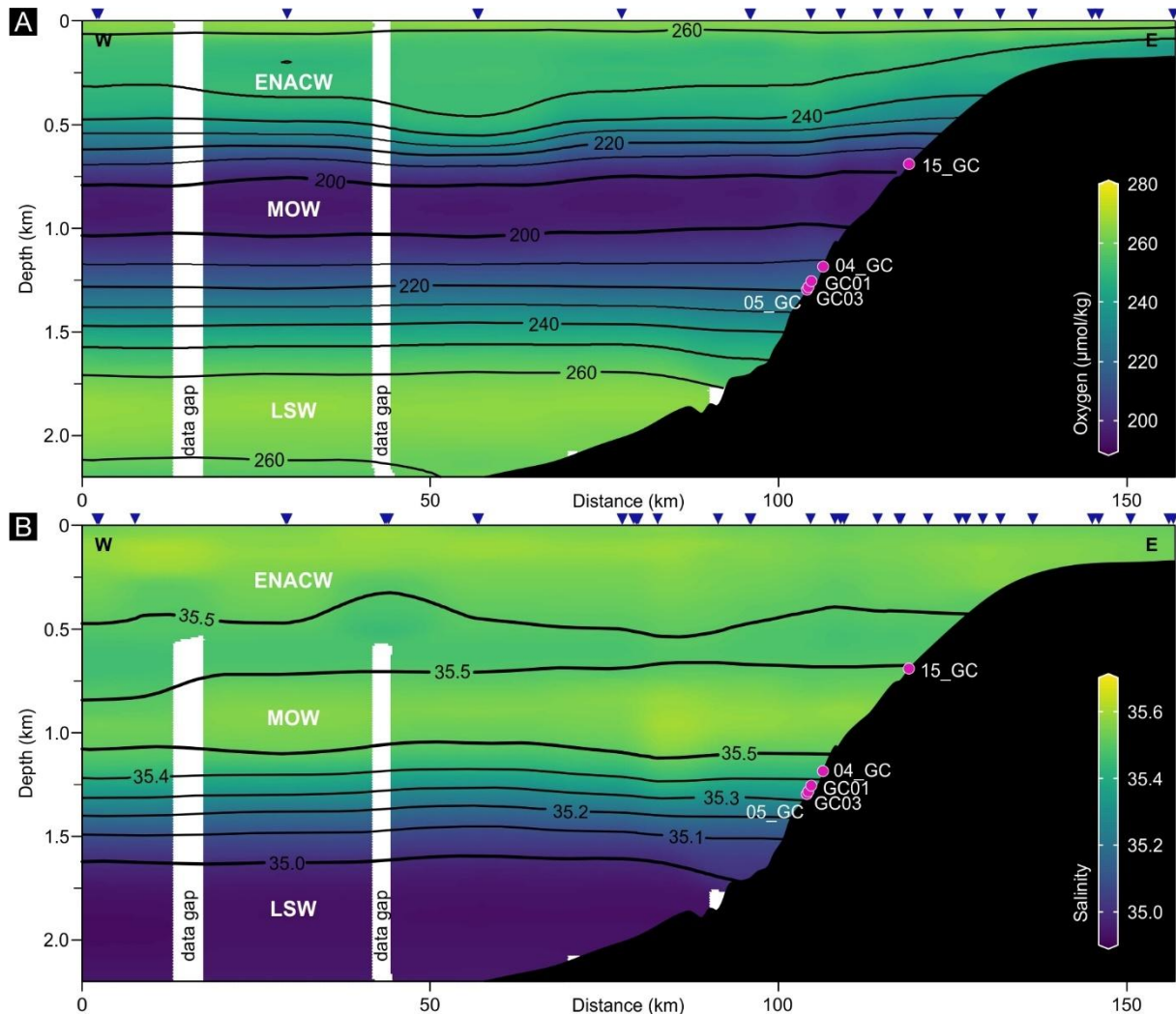


Figure 5.2. Gridded (A) oxygen and (B) salinity data from World Ocean Database CTD stations (indicated as blue triangles) (Boyer et al., 2018) along an E-W profile through the Porcupine Seabight (for location, see Figure 5.1) generated with Ocean Data View (Schlitzer, 2020). The sediment cores are indicated as pink circles at their respective water depths. Contours are drawn every 10 $\mu\text{mol/kg}$ in (A) and every 0.1 in (B). ENACW = Eastern North Atlantic Central Water, LSW = Labrador Sea Water, MOW = Mediterranean Outflow Water.

5.3 Material and Methods

A total of five gravity cores were analysed, of which three were acquired during RV Celtic Explorer survey CE21015 and two during RV Belgica survey B2023-12 (Table 5.1). The cores were cut into 1 m-sections, capped and labelled on the ship. The imaging and geophysical measurements of the CE21015 cores were performed in the British Ocean Sediment Core Research Facility (BOSCORF) at the National Oceanography Centre in Southampton, UK. For the B2023-12 cores, these measurements were completed at Ghent University and Ghent University Hospital, Belgium. All grain-size analyses were carried out at Ghent University. XRF measurements for the CE21015 core were carried out at BOSCORF, while the B2023-12 cores were scanned at the Austrian Core Scanning Facility of the University of Innsbruck. All samples for accelerator mass spectrometer (AMS) radiocarbon dating were sent to Beta Analytic in Miami, USA.

Four of the sediment cores in this study are situated in a north-south transect to the north of Bilbo Channel at c. 1200-1300 m water depth and referred to as “the mid-slope cores”, while the fifth sits at c. 700 m water depth on the interfluvium between Bilbo and Frodo channels and is referred to as “the upper-slope core”.

Sedimentary facies were defined based on the sedimentary structures and bioturbation recognised in the X-ray or CT images as well as the grain-size measurements and associated sortable silt calculations. Correlation between the five cores was based on radiocarbon dates, the trends in their magnetic susceptibility curves, visually identified ice-rafted debris (IRD), and overall lithology, along with the Ca/(Ca+Fe) curves of cores GC03 and 05_GC (Figure 5.4).

Table 5.1. Overview of the sediment gravity cores used in this study. WD = water depth.

core name (named here)	latitude	longitude	WD (m)	setting	length (m)
CE21015_04_GC (04_GC)	50°45.203'N	11°25.344'W	1185	open slope	4.49
CE21015_05_GC (05_GC)	50°43.336'N	11°24.822'W	1295	open slope	4.32
CE21015_15_GC (15_GC)	50°42.566'N	11°14.041'W	680	interfluvium	4.19
B2306-GC01 (GC01)	50°44.625'N	11°25.200'W	1255	open slope	3.01
B2306-GC03 (GC03)	50°43.123'N	11°24.750'W	1283	channel levee	3.12

5.3.1 Core imaging

Core imaging was performed to identify sedimentary structures that are not visible to the naked eye, which improves sediment facies definition and allows appropriate selection of sampling locations for subsequent analyses (e.g., grain size, radiocarbon). The CE21015 whole cores were scanned using a GeoTek ScoutXcan X-Ray imaging scanner, and ImageJ software was used for visualisation. Computed Tomography (CT) scanning was performed on the B2306 whole cores using a Siemens medical CT scanner and here, Dragonfly software was used for visualisation and basic analysis of the data. After splitting, the cores were photographed: CE21015 cores using a GeoTek MSCL Core Imaging System, and B2023-12 cores using a linescan camera incorporated in a GeoTek multi-sensor core logger (MSCL) system. Afterwards, the cores were described visually for their lithology, colour (according to the Munsell chart; Munsell Color, 2012), and structures.

5.3.2 Geophysical measurements

All cores were scanned using a GeoTek MSCL. The CE21015 cores were scanned at 1 cm resolution through an in-line setup of P-wave velocity sensor, gamma-ray density sensor, magnetic susceptibility loop sensor, and resistivity sensor. The cores were also scanned at 0.5 cm resolution using a MSCL-XYZ, which measures the colour spectrum and the magnetic susceptibility through a point sensor. The B2023-12 cores were scanned for gamma-ray density, magnetic susceptibility (point sensor) and colour spectrum at 0.5 cm resolution. Centimetre to sub-centimetre resolution was chosen to capture lithological variability at the highest possible resolution and allow for core correlation, which is what the magnetic susceptibility measurements are here used for. Other geophysical measurements are not further discussed but their results can be found in Appendix G.

5.3.3 Grain-size analysis

Grain-size analysis was carried out to characterise different sedimentary facies, and to allow calculation of sortable silt parameters that can be used for palaeocurrent strength reconstructions from the rationale that stronger currents winnow fines and allow deposition of coarser grains (McCave et al., 1995). A sampling interval of 5-10 cm was aimed for, except when trying to avoid bioturbation (as identified in X-ray or CT images and visual descriptions), or near lithological boundaries and coarser-grained intervals, where the resolution was higher. To isolate

the terrigenous fraction from the grain-size samples, a chemical pretreatment was performed according to the method described in Wils et al. (2021): the samples were boiled in 10 ml DI water with H₂O₂ (2 ml, 35 %), HCl (1 ml, 10 %) and NaOH (1 ml, 2 M) to remove organic matter, calcium carbonate and biogenic silica, respectively. Lastly, samples were boiled with sodium hexametaphosphate (1 ml, 2 %) to ensure disaggregation of the particles. Grain-size analysis was performed using a Malvern Mastersizer 3000 laser diffraction particle size analyser and each sample was measured thrice to ensure robustness and reproducibility of the results. The average distribution of each triplicate was used in further analysis. GRADISTAT v8.0 (Blott and Pye, 2001) was used to calculate grain-size characteristics as well as sortable silt parameters, the latter according to the procedure described in McCave and Andrews (2019). Percentage sortable silt (SS%) is calculated as the ratio of the percentage of 10-63 µm grains to the percentage of < 63 µm grains and mean sortable silt ($\overline{SS}$) is calculated from the weight distribution of grains in the 10-63 µm size interval (McCave et al., 1995; McCave and Andrews, 2019). According to McCave and Andrews (2019), a downcore running correlation between $\overline{SS}$ and SS% of > 0.50 is considered a good indication that the sortable silt record can be interpreted as a proxy for bottom current intensity. This correlation was here evaluated over a 9-point window. Differences in flow speed (ΔU) are calculated based on the $\overline{SS}$ values as a change of 1.33 cm s⁻¹/µm, according to the calibration for measurements from Malvern laser sizers (McCave and Andrews, 2019).

5.3.4 Radiocarbon dating

Radiocarbon dating of solitary corals from cores GC01, GC03, and 04_GC was performed to establish an age model for the cores. The corals were extracted using tweezers, cleaned using deionised water, airdried, placed in vials and sent for AMS radiocarbon analysis at Beta Analytic, Miami. The obtained radiocarbon ages were calibrated using CALIB v8.2.0 (Stuiver and Reimer, 1993) and the Marine20 calibration curve, which has an inbuilt reservoir correction of 400 years (Heaton et al., 2020). No additional local reservoir age correction was applied.

5.3.5 XRF measurements

Cores GC03 and 05_GC were scanned for their geochemical properties using an ITRAX X-ray fluorescence (XRF) machine. For core GC03, the Cr X-ray tube was set to a voltage of 30 kV and a current of 55 mA, while the exposure time was set to 8 s. For core 05_GC, the Mo-tube was set to a voltage of 30 kV and a current of 30 mA with an exposure time of 15 s. All core sections were scanned at 1 mm resolution to capture small-scale lithological variabilities and to allow correlation between cores (which is here based on the Ca/(Ca+Fe) ratio). The core halves were covered by a thin polypropylene foil to minimise desiccation and alteration of the sediment surface during measurements.

5.4 Results

In the next sections, the five sediment gravity cores are described from north to south for their lithology, features visible in the CT or X-ray images, grain-size distributions, sortable silt properties, magnetic susceptibility trends, and Ca/(Ca+Fe) XRF data. The results of the AMS analysis performed on solitary corals extracted from the cores are given in Table 5.2.

All five sediment cores have a strong positive correlation of 0.80-0.99 between $\overline{SS}$ and SS% (Appendix H). The 9-point window running correlation between $\overline{SS}$ and SS% is > 0.5 for all core sections except the upper part of core 05_GC (34-40 cm) and the lower part of core 15_GC (369-419 cm) (Appendix H).

Table 5.2. Results of the AMS radiocarbon analysis, with calibrated ages reported as [median (95 % range minimum – 95 % range maximum)].

lab code	core	depth (cm)	material	¹⁴ C age (yr BP)	calibrated age (cal yr BP)
Beta-628716	CE21015_04_GC	204.5	Solitary coral	11580 ± 30	12900 (12753 – 13055)
Beta-628718	CE21015_04_GC	350	Solitary coral	12840 ± 40	14450 (14170 – 14781)
Beta-699041	B2306-GC01	270.5	Solitary coral	11670 ± 30	12990 (12828 – 13129)
Beta-699042	B2306-GC03	44	Solitary coral	9790 ± 30	10570 (10381 – 10737)

5.4.1 CE21015_04_GC

Core 04_GC consists of poorly sorted silts.

The lowermost part (449-407 cm) of the core consists of smooth, soft, olive brown mud (sandy fine silt) with several patches of sandy grains and pieces of IRD. No sedimentary or bioturbation structures are visible in the X-ray images. In this section, $\overline{SS}$ is constant and the lowest of the whole core at about 21 μm , while magnetic susceptibility values are stable and high (Figure 5.3, 5.4).

Above (407-323 cm), a coarse silt section ($\overline{SS}$ between 24 and 28 μm) is present that has a sharp lower boundary. It contains thin sandier sections with distinct but irregular boundaries at 369-372 cm, 363 cm, and 335-337 cm. No burrows are noticeable in the X-ray images, and magnetic susceptibility values remain stable at the highest of the core (Figure 5.3, 5.4).

Two sections of very coarse silts are present at 323-303 cm ($\overline{SS} \approx 28 \mu\text{m}$; deposit IV) and at 276-252 cm ($\overline{SS} \approx 26 \mu\text{m}$; deposit III), both with gradual upper and lower boundaries. Burrows are noticeable in the X-ray images of deposit IV. In between deposits IV and III (303-276 cm), coarse silts are present with $\overline{SS}$ values of 25 – 27 μm . The magnetic susceptibility trend decreases from the bottom of deposit IV (320 cm) to just above the top of deposit III (240 cm) (Figure 5.4).

A section of dark greyish brown sandy coarse silts (252-228 cm) where bioturbation is apparent in the X-ray images is bounded at the top by a sharp boundary at 228 cm (Figure 5.3). $\overline{SS}$ values here are quite constant at 25 μm .

Above (228-185 cm) is another coarser section generally consisting of sandy very coarse silt (deposit II). Whereas the $\overline{SS}$ values here decrease upwards from 29 μm to 25 μm , the magnetic susceptibility shows an upwards increasing trend. Within this unit, an irregular but sharp colour boundary at 198 cm separates olive grey from olive brown sediments above.

Between 185 cm and 165 cm, the dark greyish brown to olive brown sediments contain several grains of IRD and are muddier (medium silt) than the sections above and below. Within this section, $\overline{SS}$ values decrease upwards from 24 μm to 22 μm along with a slight decrease in magnetic susceptibility.

The section between 165 cm and 135 cm (deposit I) is sandier and olive brown in colour. Mottled bioturbation is subtle to the naked eye but clear from the X-ray images (Figure 5.3). The $\overline{SS}$ values decrease slightly upcore in this section from 27 μm to 26 μm . The magnetic susceptibility in deposit I decreases slightly upwards (Figure 5.4).

The upper 135 cm of core 04_GC contains bioturbated medium to coarse silt that is sandier at the top. $\overline{SS}$ at 135 cm depth is 24 μm and increases to about 26 μm at the core top (Figure 5.3). An

irregular colour boundary at 59 cm depth separates olive brown silts from greyish brown to light brownish grey silts above. The magnetic susceptibility first increases slightly upwards to about 75 cm core depth and then decreases towards the top of the core (Figure 5.4).

IRD was identified at several levels throughout the core: at 446.5 cm(?), 443.5 cm, 440 cm, 404 cm, 177.5 cm, and 170 cm (Figure 5.3).

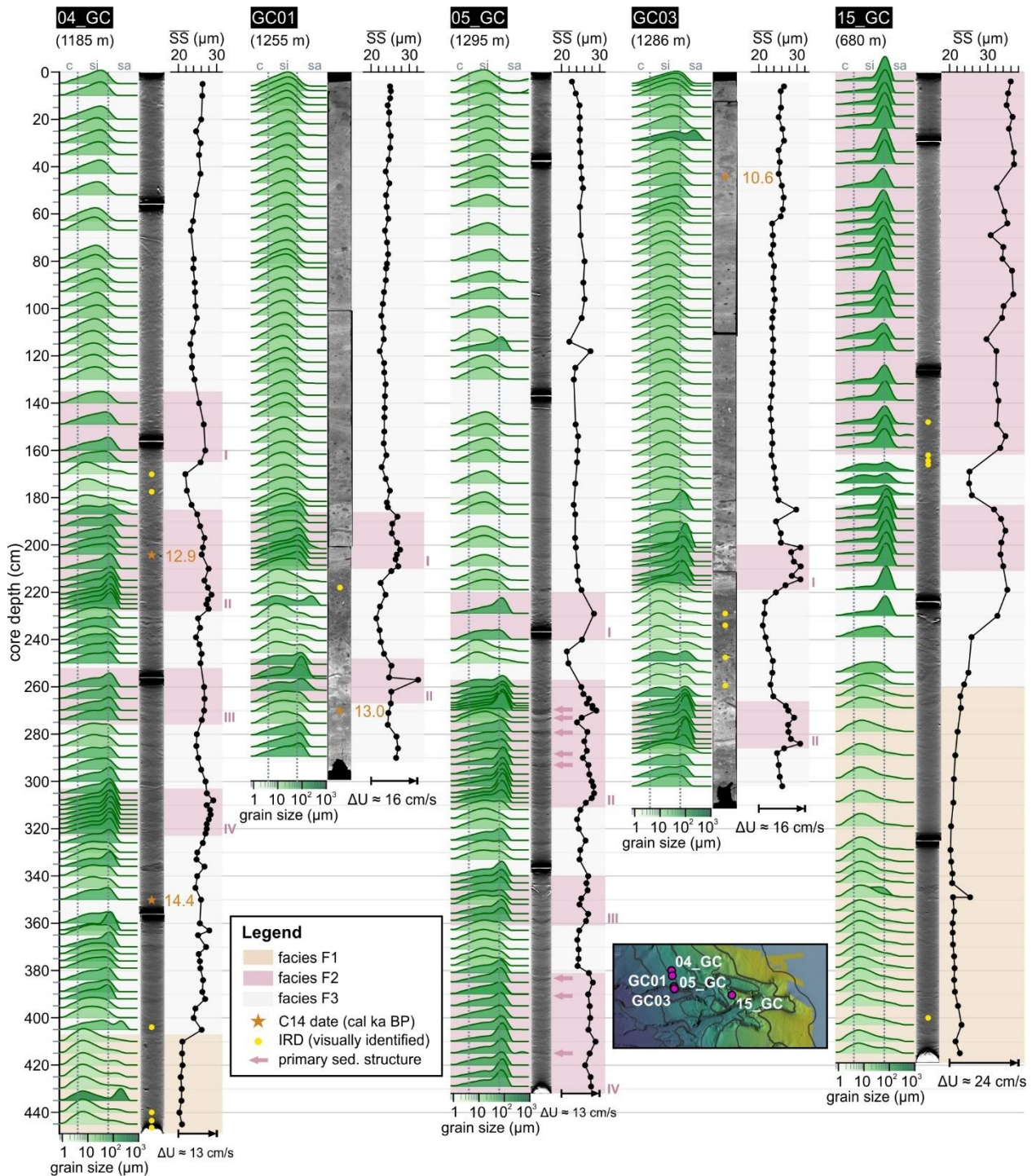


Figure 5.3. Grain-size distributions, X-ray (04_GC, 05_GC, 15_GC) or CT (GC01, GC03) data, and mean sortable silt graph. Higher intensities in the CT and X-ray data are represented by whiter colours. The flow speed change estimations (ΔU) are based on McCave and Andrews (2019). The background colours indicate the facies interpretations, with roman numerals indicating the facies F2 deposits as they are referred to in the text. The map shows the same area as in Figure 5.1B and has contour lines drawn every 250 m.

5.4.2 B2306-GC01

Core GC01 is bioturbated throughout and consists of poorly to very poorly sorted silts (Figure 5.3).

The lowest section of the core (292-267 cm) consists of coarse silts with $\overline{SS}$ values of 24-27 μm that are greyish brown in colour and gradually become lighter upwards in the section.

A slight colour change is present at 267 cm into less green sediments above. It is the bottom boundary of a coarser-grained section (coarse silt-very fine sand; $\overline{SS}$ of 25-32 μm) between 267 cm and 248 cm core depth (deposit II; Figure 5.3). Deposit II stands out in the CT images as having a higher intensity with a distinct lower boundary and gradual upper boundary (Figure 5.3). Magnetic susceptibility values here increase towards a maximum in the core at the top of deposit II (Figure 5.4).

Between 248 and 208 cm depth, the sediments are greyish brown medium silts with $\overline{SS}$ values of 21 μm – 24 μm and little variation in magnetic susceptibility. At 218 cm depth, a piece of IRD (0.8 cm diameter) is present.

At 208 cm, there is a transition to a sandier section (deposit I; $\overline{SS}$ of 26-28 μm) that extends up to 186 cm and is characterised by higher intensities in the CT images (Figure 5.3). The lower boundary is a distinct surface in the CT images and also identifiable as a colour boundary into the darker sediments of deposit I. The distinct upper boundary is visible in the CT images only. The magnetic susceptibility trend decreases upwards (Figure 5.4).

The upper section between 186 cm and 4 cm core depth consists of greyish brown medium to coarse silt that gradually becomes lighter towards the top of the core. $\overline{SS}$ values here are 23-25 μm , and magnetic susceptibility values show a gradual increase and subsequent decrease upcore (Figure 5.3, Figure 5.4).

5.4.3 CE21015_05_GC

Core 05_GC consists of poorly sorted silts without any visually identifiable IRD. From the bottom of the core to 311 cm core depth, the sediments gradually change colour from dark greyish brown to olive grey. Below 257 cm core depth, no bioturbation was identified either visually or in the X-ray images, and $\text{Ca}/(\text{Ca}+\text{Fe})$ values are constant.

An interval of very coarse silts (deposit IV; 432-381 cm) that contains indistinct layering in the X-ray images is bounded at the top by an irregular but distinct boundary at 381 cm, above which coarse silts are present. $\overline{SS}$ values in deposit IV are about 28 μm (Figure 5.3). Magnetic susceptibility values decrease slightly but steadily through deposit IV and the coarse silts above (Figure 5.4).

Another section of very coarse silts is identified from 361 cm to 340 cm with $\overline{SS}$ of 25-27 μm and a sharp lower boundary (deposit III; Figure 5.3). Above, sandy coarse silts (340-311 cm) are present with $\overline{SS}$ values of about 25 μm . The magnetic susceptibility trend remains decreasing upwards.

The section between 311 cm and 257 cm is another interval of coarser lithology (deposit II) within which coarser and finer-grained lithologies (sandy very coarse silt to silty very fine sand) alternate, with boundaries mostly gradual but sometimes sharp. The alternation of coarser and finer sediments is reflected in the variation in the $\overline{SS}$ values (24-30 μm) and layering in the X-ray images (Figure 5.3). Magnetic susceptibility values in deposit II first decrease to about 285 cm core depth and then start increasing towards the top of deposit II and the medium silt interval above (Figure 5.4).

The interval between 257 cm and 240 cm consists of medium silts with a dark greyish brown colour, mottled bioturbation, and $\overline{SS}$ values of 22 μm . Here, the $\text{Ca}/(\text{Ca}+\text{Fe})$ curve shows a decrease and subsequent increase in values.

Between 240 cm and 220 cm, another section of sandy very coarse silts is present (deposit I) and characterised by bioturbation, gradual upper and lower boundaries, and $\overline{SS}$ values of 26-28 μm (Figure 5.3). In deposit I, both magnetic susceptibility values and $\text{Ca}/(\text{Ca}+\text{Fe})$ values increase gradually upwards (Figure 5.4).

The upper 220 cm of core 05_GC is a bioturbated (sandy) coarse silt with $\overline{SS}$ values between 24 and 26 μm (Figure 5.3). The number of foraminifera increases towards the top of the core, with a high number visible in the upper 25 cm. The colour changes gradually upwards from dark greyish brown to light brownish grey into olive grey. This upper interval is interrupted by a sandier section with higher $\overline{SS}$ value (28 μm) and irregular boundaries from 120 cm to 116 cm. The magnetic susceptibility trend in the upper 220 cm of core 05_GC is quite variable. It increases slightly to 150 cm, then decreases to 125 cm, then increases to 60 cm and then decreases up to the top of the core (Figure 5.4). The $\text{Ca}/(\text{Ca}+\text{Fe})$ curve remains increasing towards 20 cm, from which it shows a slight decrease to the top of the core (Figure 5.4).

5.4.4 B2306-GC03

Core GC03 is bioturbated throughout and consists of poorly sorted silts.

The lowest section of the core (302-286 cm) consists of coarse silts ($\overline{SS} \approx 25 \mu\text{m}$), with a sharp colour boundary at 301 cm distinguishing between light brownish grey sediments below and greyish brown above. Both the magnetic susceptibility trend and the $\text{Ca}/(\text{Ca}+\text{Fe})$ curve show little variation here (Figure 5.4).

A coarser-grained interval consisting of dark grey very coarse silt to very fine sand is present between 286 and 266 cm (deposit II). It has a distinct lower boundary and diffuse upper boundary, $\overline{SS}$ values of 27-31 μm , and high intensities in the CT images (Figure 5.3). Magnetic susceptibility values increase upwards through deposit II, while $\text{Ca}/(\text{Ca}+\text{Fe})$ values remain constant (Figure 5.4).

Above, from 266 to 219 cm depth, grey to greyish brown medium silts are present with $\overline{SS}$ values of 21-24 μm . Pieces of IRD (0.5-1.3 cm diameter) were identified at 259.5 cm, 247.5 cm, 234 cm, and 229 cm core depth (Figure 5.3). Magnetic susceptibility values decrease here (Figure 5.4). $\text{Ca}/(\text{Ca}+\text{Fe})$ values first show a strong decrease and then strong increase towards the top of this section.

Between 219 and 200 cm depth, another interval (deposit I) of very coarse silts to very fine sands is characterised by a dark grey colour, higher intensities in the CT images, $\overline{SS}$ values of 30 μm , and diffuse upper and lower boundaries (Figure 5.3). Magnetic susceptibility values decrease very slightly from the bottom to the top of deposit I, while the $\text{Ca}/(\text{Ca}+\text{Fe})$ values increase slightly (Figure 5.4).

The upper section (200 cm – 4 cm core depth) consists of greyish brown coarse silts that lighten towards the top of the core. $\overline{SS}$ values are 26 μm at 200 cm core depth, decreasing to 24 μm from 180 cm to 60 cm, and increasing again to about 26 μm in the upper 60 cm (Figure 5.3). The magnetic susceptibility values increase and then decrease towards the top of the core (Figure 5.4). This coarse silt section is interrupted by a coarser-grained (very coarse silt) interval with higher $\overline{SS}$ value (30 μm) at 186-184 cm core depth that is grey in colour and has distinct upper and lower boundaries.

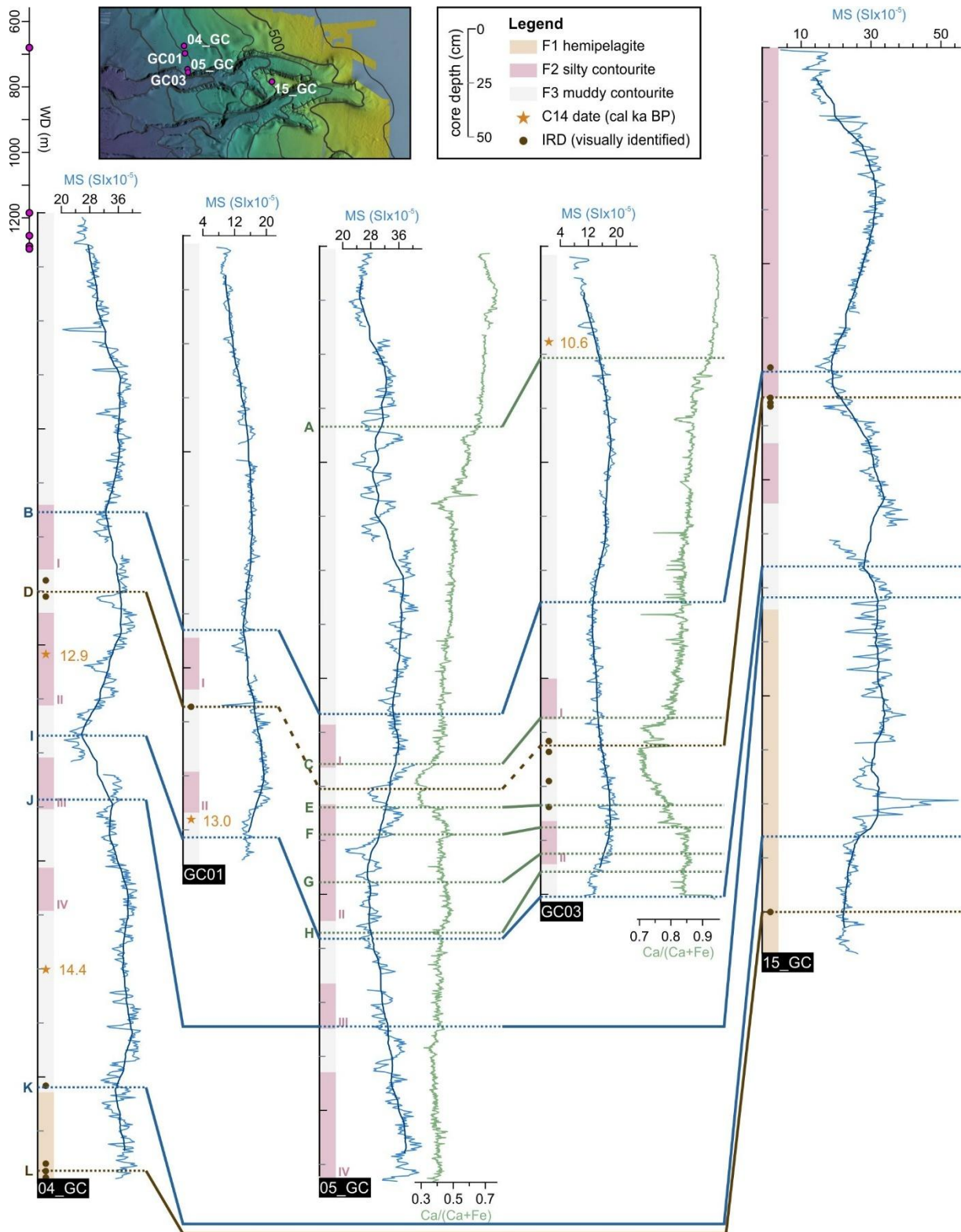


Figure 5.4. Magnetic susceptibility graph for each of the cores and $Ca/(Ca+Fe)$ graph for cores GC03 and 05_GC. Roman numerals indicate the facies F2 deposits as referred to in the text. The map shows the same area as in Figure 5.1B and has contour lines drawn every 250 m. Tie-points in core correlation are indicated by letters, which are further explained in the text. Blue correlation lines are based on changes in the magnetic susceptibility trend, green correlation lines are based on changes in the $Ca/(Ca+Fe)$ graph, and brown correlation lines are based on core intervals where IRD was identified.

5.4.5 CE21015_15_GC

Core 15_GC consists of poorly sorted silts. The X-ray images reveal that bioturbation appears at 350-345 cm core depth, then disappears up to 265 cm, from where the amount increases towards the top of the core. From 110 cm upwards, the bioturbation is visible to the naked eye (Figure 5.3).

The lowermost section of core 15_GC (419-260 cm) consists of dark greyish brown fine silt at the bottom of the core that grades into soft and smooth (sandy) medium silt (Figure 5.3). It contains a dark greyish brown interval with a red hue between 400 cm and 393 cm, with a sharp and irregular upper boundary and a more gradual lower boundary. Two subtle, irregular colour changes at 357 cm and 345 cm are associated with a low amount of bioturbation, and the interval in between (357-345 cm) is a sandy very coarse silt with a $\overline{SS}$ value of 25 μm . Outside of the coarser interval, $\overline{SS}$ values in this lowermost section of the core (419-260 cm) are 20-23 μm . The magnetic susceptibility trend increases slightly from the bottom of the core to a peak at 357-345 cm that coincides with the very coarse silt interval at 357-345 cm. From 345 cm, the magnetic susceptibility trend is constant up to the top of the section at 260 cm (Figure 5.4).

The section between 260 cm and 211 cm consists of patches and intervals of muddy sand that are irregularly intercalated with patches of more clayey material, with a decreasing amount of mud and increasing amount of sandy patches upwards. $\overline{SS}$ values in this section increase from 24 μm to 35 μm . Magnetic susceptibility values display a small decrease and subsequent increase to the top of this section.

Above 211 cm, a silty very fine sand lithology of olive grey colour and $\overline{SS}$ values of 32-35 μm is present. It transitions gradually at 185-183 cm to dark greyish brown very poorly sorted coarse silts ($\overline{SS} \approx 26 \mu\text{m}$) above. The magnetic susceptibility trend is decreasing upwards in this interval.

At 162 cm, a sharp but irregular colour and lithology boundary is present as the lower boundary of a section of olive grey sandy very coarse silt to very fine sand. It is present up to 44 cm core depth and has $\overline{SS}$ values that vary between 30 μm and 36 μm (Figure 5.3). At 50-44 cm, there is an irregular boundary to foraminifera-rich sandy very coarse silt of slightly lighter olive grey colour that contains several shell fragments (1-3 mm to 1 cm in size) and has $\overline{SS}$ values of 36 μm . In this upper section of the core above 162 cm, there is a increase and subsequent decrease in magnetic susceptibility towards the top of the core.

In core 15_GC, pieces of IRD were found at 400 cm, 166-162 cm, and 148 cm (Figure 5.3).

5.4.6 Sedimentary facies

Three lithofacies are defined based on the grain size, structure and degree of bioturbation of the sediments.

Facies F1 consists of dark greyish to olive brown poorly to very poorly-sorted medium to fine silt situated at the bottom of cores 04_GC (449-407 cm) and 15_GC (419-260 cm) (Figure 5.3). Facies F1 contains IRD at the bottom of both these cores. Bioturbation is rare and primary sedimentary structures are absent. $\overline{SS}$ values are the lowest of all facies (20-25 μm , median 21 μm ; Appendix H).

Facies F2 represents the poorly sorted (sandy) very coarse silt to very fine sand sections with common bioturbation, and upper and lower boundaries that can be sharp or gradual. Facies F2 is (olive) brown to (olive) (dark) grey in colour and present in all cores (Figure 5.3). It stands out in the CT images as intervals of higher intensity. On the X-ray images, the facies is harder to distinguish from the background sediments. In core 05_GC, facies F2 contains primary sedimentary structures (Figure 5.3). $\overline{SS}$ values (24-37 μm , median 28 μm) are generally higher than facies F1 and F3, although there is some overlap with the range of $\overline{SS}$ values of facies F3 (Appendix

H). According to the calibration of McCave and Andrews (2019), bottom currents associated with this facies have a flow speed that is c. 9 cm/s above the flow speed where facies F1 was deposited (Appendix H).

The foraminifera-rich medium to (sandy) coarse silts of facies F3 are the main background sediments in the mid-slope cores (Figure 5.3). There is limited variation in the colour of the sediments, from olive and (dark) greyish brown to (light) brownish grey and olive grey. Bioturbation in this facies is common, primary sedimentary structures are absent, and IRD is present at distinct levels in cores 04_GC and 15_GC. $\overline{SS}$ values are mostly between 23 μm and 26 μm , although several sections (266-219 cm in core GC03, 257-240 cm in core 05_GC, 248-208 cm in core GC01, 185-165 cm in core 04_GC) are present where $\overline{SS}$ values are close to those of facies F1 (21-24 μm), and one section (260-211 cm in core 15_GC) where a large range of values is present that overlaps with those of facies F2 (24-35 μm). According to the calibration of (McCave and Andrews, 2019), the flow speed where F3 sediments were deposited was c. 5 cm/s higher than for facies F1, and c. 4 cm/s lower than for F2 (Appendix H).

5.4.7 Core correlation

The Ca/(Ca+Fe) records of cores 05_GC and GC03 show downcore increasing values at the very top of the cores that transition into a gradual decrease downcore. A first tie-point (A) is placed at this transition (Figure 5.4). In the mid-slope cores, the top of the cores is characterised by facies F3 sediments where the magnetic susceptibility trend first increases and subsequently decreases downcore. The F2 sediments at the top of core 15_GC have a similar magnetic susceptibility signal to the F3 sediments in the mid-slope cores and therefore thought to correlate. The dip in the magnetic susceptibility trend at the base of the upper section in all cores is used as second tie-point (B) (Figure 5.4). Cores 05_GC and GC03 are correlated at the base of deposit I by a change to a stronger decrease in Ca/(Ca+Fe) values (tie-point C). The F3 facies underneath contains IRD in all mid-slope cores except 05_GC and might correlate to the IRD-containing F3 sediments in core 15_GC (tie-point D). On the other hand, the magnetic susceptibility trend in this interval in core 15_GC is more alike deposit I in the mid-slope cores. Cores 05_GC and GC03 contain several tie-points (E-H) in deposit II from changes in the Ca/(Ca+Fe) curve (Figure 5.4). Below, F3 sediments in the mid-slope cores are characterised by a second dip in magnetic susceptibility that is thought to correlate to a magnetic susceptibility dip in core 15_GC as well (tie-point I). Deposits III and IV are only present in cores 04_GC and 05_GC with one tie-point (J) placed at the base of the third deposit, at a local maximum in the magnetic susceptibility trend curve. A similar local maximum is seen in core 15_GC and thought to correlate (Figure 5.4). The F1 sediments at the bottom of core 15_GC have a magnetic susceptibility trend that is quite constant, similar to the bottom (> 300 cm core depth) of core 04_GC, hence this F1 facies in 15_GC is correlated to the facies F3 and F1 deposits (along with deposit IV?) at the base of core 04_GC. The large peak in magnetic susceptibility at 357-345 cm in core 15_GC is not observed in core 04_GC, but the decrease just below may correlate to a slight decrease in values in core 04_GC (tie-point K). The lowermost tie-point (L) is placed where IRD is present in both cores 04_GC and 15_GC (Figure 5.4).

5.5 Discussion

The three lithofacies that were defined are here interpreted. The strong positive correlation (0.80-0.99) between $\overline{SS}$ and SS% for all five cores (Appendix H) indicates that their grain-size record can reliably be interpreted in terms of sorting by bottom current activity (McCave and Andrews, 2019) and is used to strengthen the interpretation of the lithofacies. However, the downcore running correlation between $\overline{SS}$ and SS% is < 0.5 in the upper part of core 05_GC (34-40 cm) and the lower part of core 15_GC (374-419 cm), which indicates these deposits are not strictly controlled by bottom currents (McCave and Andrews, 2019) and they are therefore omitted from palaeocurrent

reconstructions. A general tentative age model is built, after which the bottom current variability through time is interpreted for the MOW core flow (from the mid-slope cores) and the ENACW-MOW transition zone (from the upper-slope core).

5.5.1 Sedimentary facies interpretation

The fine-grained nature, massive structure, and poorly to very poorly-sorted texture of facies F1 suggests deposition from vertical settling of particles, i.e., hemipelagic deposition (Stow and Tabrez, 1998b; Stow and Smillie, 2020). Hemipelagites can be heavily bioturbated but only a low amount of mottling is present here, which may be reflective of low oxygenation of bottom waters in combination with the calm sedimentary environment (Stow and Smillie, 2020). Along with the presence of IRD, this further allows interpretation of a glaciogenic hemipelagite (cf. Stow and Smillie, 2020) at the bottom of facies F1 (Figure 5.3).

Facies F2 deposits are interpreted as silty contourites (cf. Stow and Faugères, 2008) based on their bi-gradational nature with gradual to distinct lower and upper boundaries along with their high grade of bioturbation (Stow and Faugères, 2008; Stow and Smillie, 2020). Primary sedimentary structures are potentially present as alternating coarser- and finer-grained intervals with irregular boundaries (core 05_GC; Figure 5.3) or a sharp contact in the CT image (core GC03; Figure 5.5B). Most of the silty contourite deposits identified here, however, are devoid of primary sedimentary structures, which may be due to destruction by the pervasive bioturbation. On average, they are characterised by higher $\overline{SS}$ values than both the hemipelagites of facies F1 and the background sediments of facies F3, which suggests deposition under increased bottom current velocities (c. 9 cm/s faster than F1 deposition and c. 4 cm/s faster than F3 deposition; Appendix H). $\overline{SS}$ variability within the F2 facies could reflect either changes in sediment supply (e.g., more/less terrestrial input) or fluctuations of bottom current strength (McCave et al., 2017; Stow and Smillie, 2020). Little variation in colour within the F2 sediments might suggest little variation in sediment source and hence higher likelihood of the latter option.

The abundance of planktonic foraminifera tests along with extensive bioturbation and absence of primary laminations in facies F3 suggest deposition in a low-energy environment. These deposits share many features with hemipelagites and they are very similar to other Holocene sediments described along the Irish margin and in the Rockall Trough (Howe, 1996; Knutz et al., 2002; Van Rooij et al., 2007b; Georgiopoulou et al., 2012; Tarlati et al., 2020), where their deposition is generally attributed to hemipelagic settling. However, the $\overline{SS}$ values of facies F3 are higher than those of hemipelagic facies F1, and the F3 grain-size distribution skews more towards coarse silt rather than fine silt (Figure 5.3). This suggests a (minimal) bottom current influence on the deposits of facies F3, with a flow speed that is c. 5 cm/s higher than during deposition of facies F1 (Appendix H). In the definition of sedimentary facies, a continuum exists from hemipelagites to muddy contourites and the boundary is arbitrary (Rebesco et al., 2014; de Castro et al., 2021), but the higher $\overline{SS}$ values of F3 sediments here justifies their interpretation as muddy contourites. This highlights the difficulty in interpreting contourites in open slope settings outside of drift systems (Stow and Faugères, 2008; Stow and Smillie, 2020). As facies F3 is present at the top of the mid-slope cores, and although there are no dates to constrain the core top ages, it is thought to represent the present-day sedimentation in the mid-slope area. It corresponds to the slow flow speeds (< 10 cm/s for 70 % of the time) measured by the current meter from mooring 112 (Pingree and Le Cann, 1990; White, 2007).

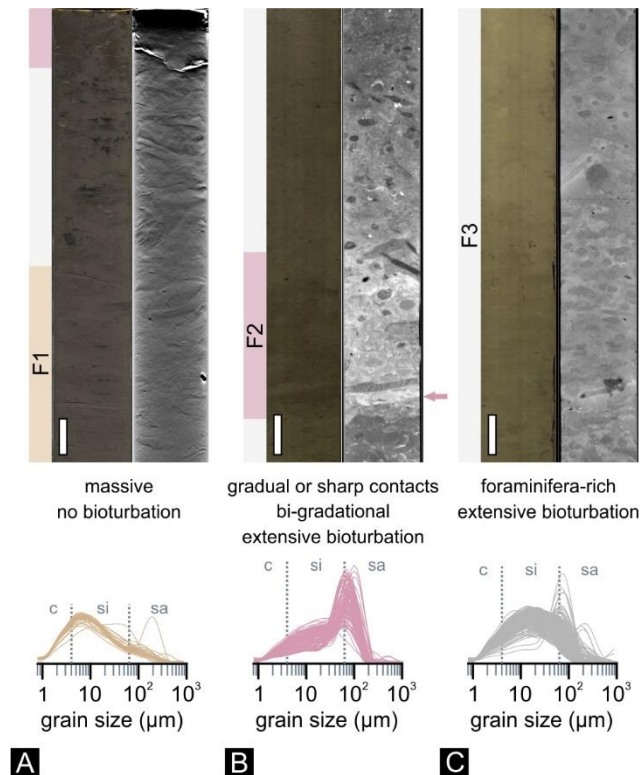


Figure 5.5. Photograph, X-ray or CT image, main characteristics and grain-size graph representative of sedimentary facies (A) F1 hemipelagite, (B) F2 silty contourite and (C) F3 muddy contourite. Images show (A) 224-287.5 cm of core 15_GC, (B) 235-292.5 cm of core GC03, and (C) 27.5-85 cm of core GC01. The white vertical scale bar represents 5 cm. The grain-size graph for each of the facies includes all grain-size distributions in the five cores classified as that facies (see Figure 5.3). The pink arrow in (B) indicates a potential primary sedimentary structure destroyed by bioturbation.

5.5.2 Tentative age model

Limitations of radiocarbon dating on solitary corals include the likelihood of their lateral transport by bottom currents, and uncertainty about the reservoir age of the water mass from which they extracted carbon, as this value is subject to spatial and temporal changes (Ascough et al., 2005; Clark et al., 2022b). Lateral transport of the coral would result in its final deposition in surrounding sediments that are younger than the coral. Since corals are benthic organisms, it is likely the bottom waters from which they extracted carbon had been disconnected from the atmospheric carbon reservoir for some (unknown) time, making the reported coral ages higher than they are. Hence, the radiocarbon ages reported here are treated as maximum ages of the sediments in which they sit (Figure 5.4). Radiocarbon ages are combined with levels where IRD was identified visually to build a tentative age model.

On the Irish margin, IRD was deposited during the Younger Dryas (12.9-11.7 kyr BP) and at the end of the last deglaciation during the early phase of Heinrich Stadial 1 (18.2-16.7 kyr BP) (Howe, 1995; Peck et al., 2007a; Scourse et al., 2009a; Tarlati et al., 2020; O'Reilly et al., 2022). This leads to interpreting the upper core interval where IRD is identified as deposited during the Younger Dryas, and the IRD at the bottom of cores 04_GC and 15_GC as deposited during the end of the deglaciation. This interpretation fits with the radiocarbon dates (Figure 5.4), and a more detailed IRD analysis (concentration, flux, provenance) would help to verify. From this tentative age model, the four silty contourites (deposits I-IV) deposited during periods of increased bottom current intensity north of Bilbo Channel are placed between 14.5 kyr BP and the earliest Holocene (< 11.7 kyr BP), i.e., during the Bølling-Allerød and Younger Dryas periods (Rasmussen et al., 2006). From the correlation between the mid-slope cores north of Bilbo Channel and upper-slope core 15_GC, upper-slope deposition of hemipelagic facies F1 seems to have continued longer

after the deglaciation. The presence of the silty contourite below the IRD-containing deposits in core 15_GC suggests that one period of intensified bottom currents occurred on the upper slope before the Younger Dryas. The upper contouritic section in core 15_GC is of Holocene age (Figure 5.4).

5.5.3 Sedimentary record of the slopes surrounding the Gollum Channel System

From the tentative age model and sediment characteristics of the five sediment cores, it is possible to reconstruct the variability of both MOW core flow at 1200-1300 m water depth and bottom current action at the ENACW-MOW interface since the last deglaciation.

5.5.3.1 Last deglaciation

During the final part of the last deglaciation, deposition on the interfluvies of the Gollum Channel System at 700 m and 1200-1300 m water depth was characterised by hemipelagic brown fine silts with low δS values. This indicates that bottom currents at this time were low-intensity across the slope (Figure 5.6), and corresponds to flow speed minima recorded on the upper slope (1000 m) of the northern Bay of Biscay between 19 and 16.5 kyr BP (Toucanne et al., 2021), plumite deposition 18-15.5 kyr BP in a calm environment on the Donegal-Barra Fan (Tarlatti et al., 2020), and muddy hemipelagic deposition during late glacial times on the upper slope (c. 1000 m) of Porcupine Bank (Øvrebø et al., 2006). Sluggish bottom current conditions along this margin during the early part of Heinrich Stadial 1 have been ascribed to a weakened Glacial Eastern Boundary Current at this time (Toucanne et al., 2021).

During late Quaternary cold periods, many proximal records suggest a strengthening of the lower core of MOW flow, which flows westwards into the Atlantic from the Gulf of Cádiz (Schönfeld and Zahn, 2000; Rogerson et al., 2012a). MOW upper core flow was much reduced, and so was the total outflow volume at the Strait of Gibraltar (Schönfeld and Zahn, 2000; Rogerson et al., 2012b), preventing MOW from reaching the Porcupine Seabight. Instead, at the end of the last deglaciation, the core locations were more likely bathed by a water mass sourced from the north (Peck et al., 2007b) that was transported cyclonically along the West-Atlantic margin and then east to be incorporated into the Glacial Eastern Boundary Current (Toucanne et al., 2021).

5.5.3.2 Bølling-Allerød to Younger Dryas

At 1200-1300 m water depth, bottom current strength increased during several time intervals between 15 and 11 kyr BP (Figure 5.3, 5.6). This invigoration corresponds to a strengthening of the European Slope Current after 17-16 kyr BP, peaking at 13.6-12.0 kyr BP interpreted by Depuydt et al. (2024) from a benthic foraminifera record in the southern Bay of Biscay. Similarly, Toucanne et al. (2021) interpreted a rapid increase in the strength of this slope current 16.5-14.7 kyr BP at c. 1000 m water depth from flow speed records and isotopic data in the northern Bay of Biscay. Hanebuth et al. (2015) also noted a strengthening from 17 kyr BP and particularly 16-8 kyr BP on the Iberian margin. This strengthening is part of a reinvigoration of the upper branch of the Atlantic Meridional Overturning Circulation that was potentially caused by a rise in atmospheric CO₂ (Zhang et al., 2017; Ng et al., 2018) and resulted in northwards heat transport and increased melting of high-latitude ice sheets at the start of the Bølling-Allerød (Ng et al., 2018; Brendryen et al., 2020; Toucanne et al., 2021; Depuydt et al., 2024).

At 700 m water depth, deposition remained hemipelagic for a longer time after the deglaciation and one interval of silty contourites deposited before the Younger Dryas suggests only one period of stronger bottom currents (Figure 5.3, 5.6). The gradual change in sedimentology from hemipelagite to silty contourite implies a gradual change in the bottom current environment. The absence of a signal of increased flow strength during the Bølling-Allerød here may be due to a deeper settling depth of the MOW before the Holocene (Rogerson et al., 2012b; Sierro et al., 2020). Vertical shifts in the settling and flow depth of the MOW during deglaciations have been

explained by variability in freshwater input and sea level in the Mediterranean Sea changing the salinity (and therefore density) of the water flowing out at the Strait of Gibraltar (Bahr et al., 2015; Kaboth et al., 2016; Sierro et al., 2020), and by variability of the vertical density gradient in the Atlantic due to changes in freshwater input into the North Atlantic Ocean (Rogerson et al., 2012a; Hanebuth et al., 2015; Sierro et al., 2020). In this latter process, increased freshwater input reduces the density of intermediate and deep water masses created at high latitudes, and MOW is thought to be forced to greater depths (Sierro et al., 2020).

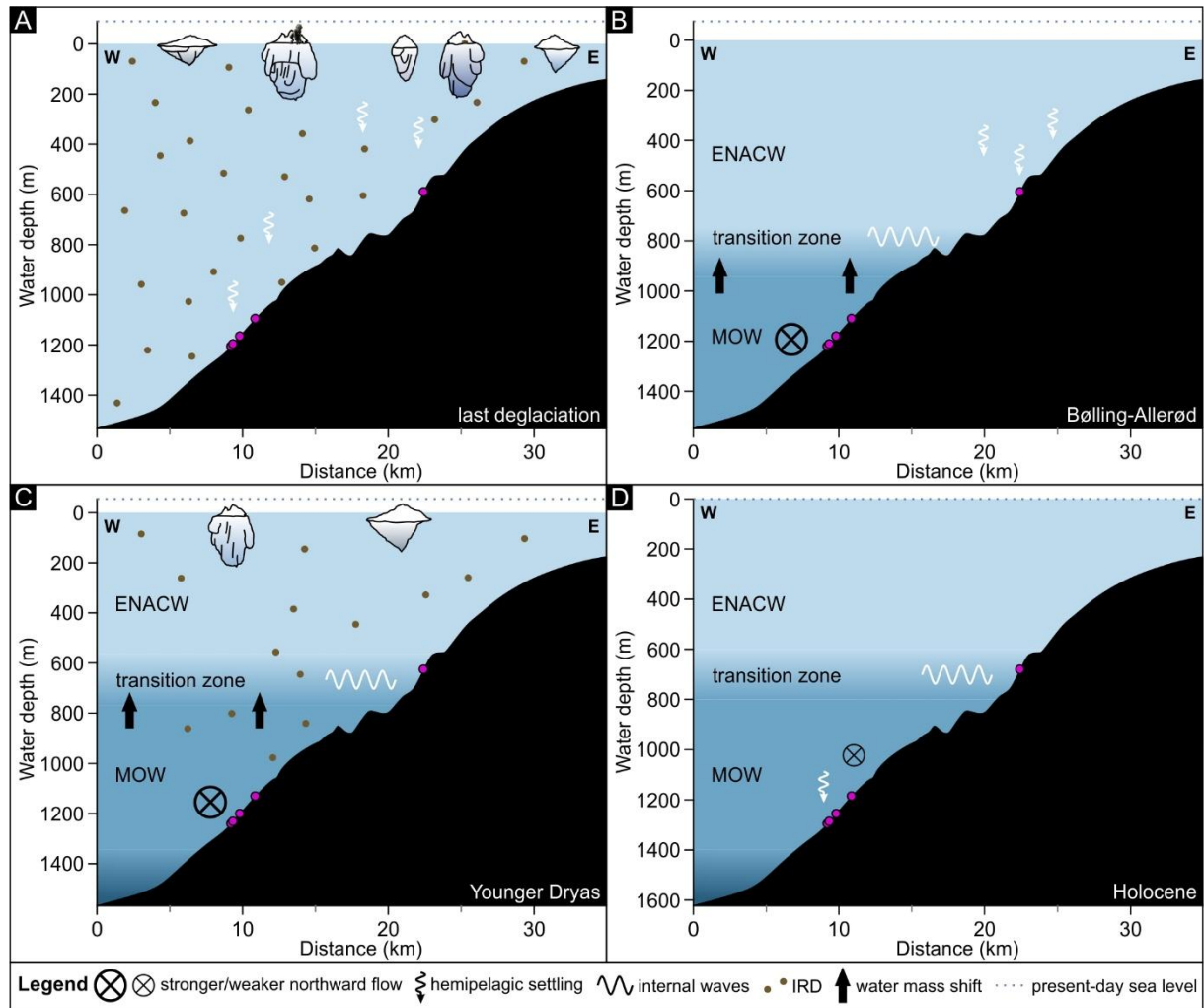


Figure 5.6. Schematic drawings showing the evolution of sedimentary processes and oceanographic changes over the upper continental slope in the Gollum Channel System area since the last glacial period. Relative sea level changes are from PALTIDE (Bradley et al., 2011; Ward et al., 2016; Scourse et al., 2024). (A) End of the last deglaciation: hemipelagic settling with IRD dominates across the slope. (B) Bølling-Allerød: presence of the Mediterranean Outflow Water (MOW) at a deeper settling depth than today creates silty contourites at 1200-1300 m water depth. Deposition at 700 m is hemipelagic. (C) Younger Dryas: MOW has moved upwards and the ENACW-MOW interface starts influencing deposition (silty contourites) at 700 m water depth. Bottom currents at 1200-1300 m water depth remain strong (silty contourite). (D) The situation in the early Holocene is similar to today: bottom currents from MOW core flow are less strong at 1200-1300 m water depth and produce muddy contourites, while the strongest currents occur at the ENACW-MOW interface (silty contourites).

5.5.3.3 Holocene

Increased bottom current velocities at the upper slope from the onset of the Holocene onwards could in turn be explained by a gradual shoaling of the MOW, potentially in combination with a vertical expansion (Schönfeld and Zahn, 2000). This is similar to the upward shift of the core of the slope current that Øvrebo et al. (2006) interpreted from core transects on the western

Porcupine Bank. Additionally, a 250-300 m upward shift of the MOW during the deglaciation and early Holocene was interpreted on the Iberian margin from sediment drift evolution at the lower boundary of the water mass, where it interacts with the Labrador Sea Water (Hanebuth et al., 2015), and from age differences between a shallower and a deeper cold-water coral chain in the Belgica Mound Province (Wienberg et al., 2020).

Bottom current velocities around 700 m water depth have remained high since the beginning of the Holocene, which is likely linked to the presence of internal waves at the MOW-ENACW interface and their capability of enhancing bottom currents (White, 2007; White and Dorschel, 2010). At 1200-1300 m water depth, on the other hand, one last period of intensified bottom currents occurred at the beginning of the Holocene, after which bottom current intensity decreased to present-day values (Figure 5.3, 5.6). This is interpreted to also be related to the upward shift of the MOW water mass and the associated MOW core flow to shallower water depths than the core locations. Overall, Holocene deposition shows little variability in both mid- and upper-slope cores and suggests present-day oceanography was established c. 10 ka and has remained stable since. This is in agreement with Holocene stability in flow speeds at intermediate depths in the northern Bay of Biscay (Toucanne et al., 2021), but differs from diminishing bottom current energies throughout the middle-late Holocene inferred from sediment cores in the southern Bay of Biscay (Depuydt et al., 2024). The latter are thought to be related to a contraction of the sub-polar gyre along with long-term weakening of the meridional density gradient by the absence of freshwater input from melting ice sheets at high latitudes combined with decreasing insolation no longer warming waters at subtropical latitudes from c. 8 ka onwards (Depuydt et al., 2024).

5.5.4 Implications for the Irish margin placed in their palaeoceanographic context

In the lower part of core 04_GC, the distinct boundary between the brown mud at the top of the hemipelagic facies and the greyish coarse silt above suggests a sudden change in oceanography rather than a gradual depth shift of the governing water mass. From core top ages at 750-1000 m water depth in the Belgica Mound Province, Wienberg et al. (2020) suggest re-entrance of the MOW into the Porcupine Seabight after the Last Glacial Maximum from 18 kyr BP at the earliest. In the Bay of Biscay, Depuydt et al. (2024) found evidence for the presence of MOW from 15.5 kyr BP onwards in benthic foraminifera assemblages at 1200 m water depth. In the study area, the muddy and silty contourites in core 04_GC suggest MOW presence at 1200-1300 m water depth at the end of the last deglaciation (c. 16 kyr BP) at the earliest and definitely from 14.4 kyr BP onwards (maximum age of deposit IV). Since the MOW is here interpreted to have settled deeper when it first entered the Porcupine Seabight and then shallowed during the Bølling-Allerød to early Holocene, it is possible that the early MOW signature is present here rather than at the shallower depths of the Belgica Mound Province. Seeing as bottom currents in the Belgica Mound Province are enhanced compared to the Gollum Channel System area, core top ages there may be affected by deglaciation/Holocene erosion and the 18 kyr BP MOW entrance implied from them (Wienberg et al., 2020) may be interpreted as a maximum age.

In general, the findings of this study fit into the ubiquity of late Quaternary contourites along the Irish margin from Goban Spur to the Donegal-Barra Fan (Øvrebø et al., 2006; Van Rooij et al., 2007b; Tarlati et al., 2020; Fabian et al., 2023). This highlights the influence of the European Slope Current on margin-wide deposition, even on the slopes of a canyon-channel system where downslope processes might traditionally be considered dominant. Moreover, the MOW that is carried by the European Slope Current creates internal waves and tides at its upper and lower interfaces and is thus responsible for enhanced bottom current energy at 600-800 m water depth and 1500 m water depth in the Belgica Mound Province and on Goban Spur (White, 2007; Delivet et al., 2016). This influence is here shown to have created Holocene silty contourites in the Gollum Channel System as well.

Placing the results of this study in their palaeoceanographic context, they fit within the Glacial Eastern Boundary Current (i.e., the glacial equivalent of the European Slope Current; Toucanne et al., 2021) weakening/strengthening model of Toucanne et al. (2021) during interstadials/stadials of the last glacial period. In this model, weakening of the Glacial Eastern Boundary Current (and, by extension, the Atlantic Meridional Overturning Circulation) preceded Heinrich events, which in turn provided a positive feedback mechanism, weakening the circulation system further through large-scale freshwater input (Toucanne et al., 2021). Because of the weaker circulation, intermediate waters in the North Atlantic were cooled less efficiently (Mignot et al., 2007). These warmer waters were carried north by the Glacial Eastern Boundary Current, where they caused the Atlantic Meridional Overturning Circulation to reinvigorate and the climate system to switch back to an interstadial period (Toucanne et al., 2021). If the mid-slope contourite record of the Bølling-Allerød and Younger Dryas presented here indeed predominantly reflects the strength of the European Slope Current (during a period of overall shallowing of the MOW), dating of each individual contourite interval may provide a very detailed archive of strengthening/weakening phases of this current at the transition between an intense cold period and a stable interglacial period. This may illuminate further on the intricate interactions between ocean circulation and climate. To reconstruct the variability in the Holocene environment, this study should be extended to cores closer to the water depths of the present-day core of the alongslope current (c. 1000 m), as its inferred upward movement before the Holocene has placed the mid-slope cores presented here outside of its influence. Since the MOW and the related density gradient with the overlying ENACW is essential for the development of deep-sea habitats through enhanced bottom currents (Chapter 4; White and Dorschel, 2010; Wienberg et al., 2020), understanding the variability in its settling depth is relevant to understanding how these habitats may change in light of past and future climate changes.

5.6 Conclusion

Five sediment cores situated between the Bilbo and Frodo channels of the Gollum Channel System in the Porcupine Seabight were analysed to reconstruct variability in bottom current strength related to the flow of the Mediterranean Outflow Water at a distal site. Three lithofacies representing different bottom current states were identified: hemipelagites, muddy contourites, and silty contourites. The combination of upper-slope (700 m water depth) and mid-slope (1200-1300 m water depth) cores allows bottom current reconstruction since the last deglaciation. Even though the age model proposed here is tentative and can be improved, this study highlights the knowledge that can be gained from just five strategically positioned sediment cores.

At the end of the last deglaciation (c. 16-17 kyr BP), bottom currents were slow and deposition mostly hemipelagic, with the presence of IRD highlighting the glacial influence on the margin. During the Bølling-Allerød and Younger Dryas, silty contourites indicate several periods of intensified bottom currents at 1200-1300 m water depth, while hemipelagic deposition at 700 m water depth suggests absence of current-related deposition. This is interpreted as deposition from MOW core flow at 1200-1300 m and absence of this water mass at 700 m. The MOW is thought to have shoaled during this period (15-11 kyr BP) and settled at its present-day water depth (core at 800-1000 m) in the early Holocene. This is visible as a gradual change from hemipelagic over muddy contouritic into silty contouritic deposition on the upper slope. Since the early Holocene, deposition has remained stable, producing muddy contourites under the influence of a slow northward flow on the mid-slope and silty contourites on the upper slope. Here, the transition zone between ENACW and MOW is associated with a permanent pycnocline where internal waves enhance bottom currents. This ties into the margin-wide dominance of the European Slope Current (and especially the MOW that it carries) on sedimentation, and illustrates the importance of the internal waves at the MOW upper boundary outside the Belgica Mound Province, and likely along the entire eastern Porcupine slope.

It is worth noting that contourites related to MOW flow on the Porcupine slope are produced in an open slope setting in a canyon-channel system, which differs from the large contourite drift bodies where these records are usually sought. This study illustrates that next to their record of downslope processes from the channel floors, canyon-channel systems can also provide a record of alongslope processes and when combined, these may provide a valuable archive for climate variability.

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*“An understanding of the natural world and what’s in it
is a source of not only great curiosity but great fulfilment.”*

Sir David Attenborough

CHAPTER 6 Conclusions

6.1 Answers to research questions

The data and analyses presented in this thesis have significantly improved the understanding of sedimentary and oceanographic processes in the Gollum and Kings channel systems that cross-cut the continental slope in the Porcupine Seabight basin offshore Ireland. High-resolution multibeam bathymetry mapping (15 m here versus 50 m before; Beyer et al., 2003) revealed features that had not been recognised before. Some of these were investigated in more detail using very high-resolution AUV-based geophysical data. Findings from these mapping efforts were ground-truthed with sediment gravity cores and current meter data. The Gollum Channel System shares features that have been observed in many other submarine canyon-channel systems, such as topographically steered tidal bottom currents, but also contains unique aspects such as cold-water coral mini-mounds. Additionally, the stratigraphy and subbottom structures underneath the two channel systems were described in detail for the first time. Combining these different types and resolutions of datasets has enabled the interpretation of slope processes and their evolution since the Cretaceous, with a focus on the late Neogene – Quaternary and a detailed description of post-Last Glacial Maximum processes.

When were the Kings and Gollum channel systems formed and which processes were involved in their development?

Single-channel seismic data (Chapter 2) suggest that incision of the Kings Channel System by erosive downslope gravity flows likely occurred in the late Pliocene – early Quaternary at the onset of northern hemisphere glacial oscillations. The seismic reflector at the bases of the channels is the RD1 reflector that is recognised across the eastern margin of the Porcupine Seabight and has been dated in the Belgica Mound Province, where it represents a hiatus between upper Miocene and upper Pliocene sediments. It was created by erosion from high-energy bottom currents associated with the introduction of the Mediterranean Outflow Water in the basin 3.5 – 3.3 Ma during a general phase of strengthening of the slope current that brings this water mass north from the Gulf of Cádiz (Van Rooij et al., 2003; Khélifi et al., 2009). The Kings channel floors being the same RD1 reflector indicates this phase of erosion/non-deposition across the eastern Porcupine margin locally included a downslope component where gravity flows were likely sourced from early Quaternary glacial processes. After the initial incision, gravity flows through the Kings Channel System were mostly depositional and formed internal levees, although some phases of flank erosion and channel widening or lateral displacement of the thalweg are recognised as well.

The origin of the Gollum Channel System was unknown, but erosive gravity flows resulting from glacial sediment input into the Porcupine Seabight were always thought influential in the formation of the system (Wheeler et al., 2003). The multi-channel seismic data presented in Chapter 3 indicate that the system originated before the onset of Quaternary glacial cycles. The initial incision is here tentatively attributed to a phase of North Atlantic margin tilting in the early Pliocene that has been recognised in seismic data further north along the margin (Praeg et al., 2005; Stoker et al., 2005b). The channels are also shown to have formed asynchronously, and while some gravity flows from glacial sediment input in the Quaternary contributed to the infill of the oldest channels, others were erosive and created new branches in the system.

An indication for the morphological differences between the two channel systems may be found in the orientation of their branches. The Kings channels are oriented towards Ireland, while the Gollum channels are oriented more east-west and potentially received sediment input from the shelf rather than direct terrestrial input from Ireland. These different sediment sources may have

produced different amounts and types of sediment, which may have led to gravity flows of different magnitudes being brought to the two channel systems.

Do the downslope processes in the Gollum Channel System interact with the alongslope currents over the eastern slope of the Porcupine Seabight?

An evolution in interaction between downslope flows in and alongslope currents over the Gollum Channel System is suggested by the (a)symmetry of the levees seen in the seismic data of the oldest channels. Underneath the RD1 reflector (in early-middle Pliocene sediments), levees are symmetrical and without any indication for alongslope currents influencing downslope transport (Figure 3.6). Above the RD1 reflector, northern levees are built out more than the southern flanks, which is interpreted as an expression of a northerly alongslope current thieving a portion of fine-grained sediments from the suspension clouds of gravity flows running through the channels. This asymmetry disappears gradually towards the present-day seafloor (i.e., upwards in the seismic record), where asymmetry is still visible but less pronounced as smoother, more rounded southern flanks and slightly elevated northern edges. This is interpreted as a diminishing interaction of the alongslope current with downslope processes as the channels become deeper and gravity flows completely contained within them.

The recent importance of alongslope currents on deposition in the Gollum Channel System is also shown in the sediment cores of Chapter 5. Silty contourites created by internal waves at the upper boundary of the Mediterranean Outflow Water are the dominant facies at 700 m water depth. They are also an important component of the cores at 1200-1300 m water depth, where their presence between the Bølling-Allerød and early Holocene and subsequent replacement by muddy contourites suggests core flow of the Mediterranean Outflow Water and subsequent shoaling to its present-day depth (c. 800-1000 m water depth). Besides the silty contourites that suggest increased bottom current energies, the sediment cores consist of hemipelagic to muddy contouritic deposits without any indication of recent overspill from gravity flows.

Is downslope activity in the Gollum Channel System linked to Quaternary glacial-interglacial cycles?

Downslope activity in the Gollum Channel System is inferred between 23 and 9 kyr BP from turbidites in the channel floor sediment cores presented in Chapter 4. In the period since 9 kyr BP, no downslope activity is recorded and the channel floor environment is dominated by tidal bottom current activity. This does suggest a cyclicity in downslope activity linked to climate variability, although activity seems to persist for several millennia after a deglaciation and into a period of elevated sea levels where the Gollum Channel System is detached from terrestrial input.

Which processes are active in the Gollum Channel System presently?

Inside the channels at < 1000 m water depth, the present-day environment is dominated by tidal bottom currents oriented along the channel axes that create sandy megaripples (Chapter 4). This persistent up- and downslope movement combined with the absence of downslope activity since 9 kyr BP has allowed the development of cold-water coral mound habitats in the upper channels (700 – 1000 m water depth). The deeper reaches of the channels are dominated by hemipelagic settling. Although there are some indications for downslope activity in bedforms in the channel head of one of the channels, this activity is minimal in volume and spatial extent. Thus, the Gollum Channel System is considered generally inactive in the present-day highstand conditions.

The parallel character of the Quaternary seismic reflections on the slopes surrounding the Gollum channels (Chapter 2) suggests a dominance of hemipelagic settling on present-day deposition. Sediment cores, however, present a higher-resolution dataset. Muddy to silty contourites deposited by Mediterranean Outflow Water core flow and internal waves at its upper

interface are found to make up a significant part of the deposition during the deglaciation and interglacial following the Last Glacial Maximum (Chapter 5).

6.2 Wider implications

6.2.1 The Gollum Channel System as a conduit for sediment input into the Porcupine Seabight

6.2.1.1 Pre-Quaternary

The initial incision of the Gollum Channel System is here tentatively linked to a phase of early Pliocene margin tilting (Chapter 3). Significant build-up of levees underneath the RD1 reflector (i.e., before the Quaternary) suggests downslope activity remained high after this first erosive phase and the system may have started extending towards the Porcupine Abyssal Plain. Activity was purely downslope and carried particles west into the basin with limited overspill that formed outer levees on northern and southern flanks in close proximity to the channels.

6.2.1.2 Early on in the Quaternary

When interaction between downslope and alongslope currents became possible, particles brought into the basin from the Celtic Sea Shelf were also transported to the north. This combination of terrestrial input and alongslope dispersal likely provided both nutrients and sediments essential for cold-water coral growth and mound build-up in the cold-water coral mound provinces in the northern Porcupine Seabight. The time interval where downslope/alongslope flow interaction is inferred to have occurred over the Gollum Channel System (i.e., the late Pliocene and early Pleistocene) coincides with the period when mound growth was fastest and continuous in the Belgica Mound Province (Huvenne et al., 2009b; Thierens et al., 2013). Current velocity measurements in the Gollum Channel System area (mooring 112 of Pingree and Le Cann (1990)) and seafloor sediment characteristics (Chapter 5) do indicate that the present-day alongslope flow is low in energy and not capable of transporting sediment larger than coarse silt. On the other hand, the silty contourites in the sediment cores between Bilbo and Frodo channels (Chapter 5) suggest that bottom current energy fluctuates both temporally and spatially and has been capable of suspending sandy grains over the eastern slope of the basin in the past, perhaps over longer distances. In the Belgica Mound Province, the sediment record indicates that the early Quaternary was a period of energetic bottom currents that restricted sediment deposition to the framework of growing cold-water coral mounds; sedimentation on the open slopes adjacent to the mounds was limited to absent (Huvenne et al., 2009b; Thierens et al., 2010; Matossian and Van Rooij, 2024). Although local topographic steering by the mounds (Matossian and Van Rooij, 2024) and current enhancement by trapped internal tides (White, 2007; White et al., 2007) were important to create such a high-energy environment, it is likely that early Quaternary bottom currents over the Gollum Channel System also were stronger than today and capable of thieving the suspended sediment from downslope-running gravity currents.

Temporal extrapolation of the results of Chapters 4 and 5 would suggest channel activity was highest during glacial periods and subsequent deglaciations throughout the Quaternary, while alongslope currents were weak during glaciations, increased during deglaciations and strongest during interglacial periods. Therefore, interaction between downslope and alongslope flows in the Gollum Channel System likely occurred mostly during (early) Quaternary deglaciations. On the other hand, the variability in both downslope and alongslope activity may have been smaller at the start of the Quaternary, when glacial-interglacial cycles were shorter (41 kyr) and produced less extreme environmental changes than during the 100 kyr-cycles from the Middle Pleistocene onwards (Lisiecki and Raymo, 2005). So perhaps, a regular interaction with little dependence on glacial/interglacial conditions was possible in the Early Pleistocene.

6.2.1.3 Later in the Quaternary

Downslope activity through the Gollum channels was likely most intense in the cold periods of the late Quaternary when extreme sea-level lowstands with extensive marine-terminating ice sheets provided efficient sediment supply (Chapter 4). By this time, though, turbidity currents were likely mostly confined within the Gollum channels as the channel flanks had been built up (Chapter 3). Thieving of sediments from their suspension clouds by the alongslope current was hereby likely restricted and sediment dispersal towards the north reduced.

Analysis of sand-sized sediment grains deposited within the coral framework in the Belgica Mound Province indicates that most grains are sub-angular to sub-rounded (Thierens et al., 2010). A larger share of sub-rounded grains at the start of the Quaternary is interpreted as a stronger influence of bottom currents on deposition then compared to more sub-angular grains indicating a dominance of glacial processes on deposition in the late Quaternary (Thierens et al., 2010). This fits with an alongslope current thieving grains off downslope currents over the Gollum and Kings channels in the early Quaternary and a reduction of this interaction and sediment dispersal to the north when gravity flows became confined to the channels later in the Quaternary. At the same time, contour current intensity in the basin likely reduced towards the late Quaternary (Matossian and Van Rooij, 2024), which could have reduced the likelihood of downslope/alongslope current interaction further. Other potential pathways for sandy particles to the Belgica Mound Province (cf. Figure 1.1) are direct input from glacial processes (i.e., not through the channels); resuspension of outer shelf sediments by, e.g., storms, and local transport onto the slope; and primary productivity.

For muddy sediments, on the other hand, as the different datasets presented in this thesis show a dominance of contourites and hemipelagites in slope deposits, transport in the basin is likely mainly from alongslope currents and vertical settling of terrigenous and biogenic particles. This is also indicated by the similarity in characteristics of muddy background sediments between cores inside Bilbo channel (Chapter 4), those on the slope just north of the channel (Chapter 5), and even across the Irish margin (Øvrebø et al., 2006; Tarlati et al., 2020). This suggests a similar source of sediments along the entire margin, which makes deposition from a large-scale process such as an alongslope current seem more likely. In this context, it may be worth keeping in mind that gravity flows can contribute particles to intermediate nepheloid layers and perhaps in the later part of the Quaternary, terrestrial input through the channels was temporarily stored in nepheloid layers before being captured by the alongslope current. This process is thought to have created a contourite drift next to the São Vicente canyon on the Iberian margin (Mencaroni et al., 2021), and may allow gravity currents in the Gollum channels to still provide input to the contour current even when direct interaction is restricted.

The periods of intense downslope activity in the late Quaternary were also periods of enhanced particle delivery to the west of the basin, and may be responsible for further extending the system towards the abyssal plain. The abundant thin turbidites identified in the channel floor sediment cores at 1900 m water depth (Chapter 4), however, suggest that gravity flows were low in volume and most flows did not reach much further downslope and contributed only to upper slope sedimentation. Only some turbidites are thicker (and more sandy) and likely correspond to flows that continued to greater depths. Some of these (Figure 4.8) were deposited when the British-Irish Ice Sheet had retreated from the shelf completely and sea levels were increasing (Clark et al., 2022b; Scourse et al., 2024), leaving the channels disconnected from their sediment source. Finally, the hiatus in core B2306-GC04 (Figure 4.9) may be related to a larger flushing event that caused erosion in the upper slope area and perhaps continued to the end of the system. The data presented here are not capable of explaining these differences in gravity flow intensity, which highlights the uncertainty that is still associated with the sources and processes responsible for downslope activity in the Gollum Channel System.

6.2.2 Dominance of the alongslope current on deposition across the margin

The discussion above (section 6.2.1) highlights how the European Slope Current has been instrumental to the sedimentary processes across the Porcupine Seabight – both as a source of sediment and as a manner of sediment dispersal. More specifically, the presence of the Mediterranean Outflow Water – carried by the European Slope Current – has been essential in Pliocene-Quaternary slope construction. This is evidenced by the seismic and sedimentary data presented here, from the presence of contourite facies on the slope in the single-channel seismic data of Chapter 2, the evolution from symmetric to asymmetric levees next to several of the Gollum channels in the multi-channel seismic data of Chapter 3, the megaripples created by tidal bottom currents shown in Chapter 4, and the contourite facies on the slope present in the gravity cores of Chapter 5. The introduction of the Mediterranean Outflow Water into the Porcupine Seabight also created the large-scale RD1 erosional surface whose topography formed the basis for subsequent Quaternary deposition. The contourite facies are linked to the core flow of the alongslope current as well as internal waves and tides created at the interface between the Mediterranean Outflow Water and the overlying Eastern North Atlantic Central Water.

The influence of the European Slope Current on sedimentation over the Gollum and Kings channel systems fits into a margin-wide dominance of alongslope processes on Quaternary sedimentation. A pattern of alongslope current strengthening/weakening during interglacial/glacial periods similar to what has been described here is identified across the Irish margin on Goban Spur (de Graciansky et al., 1984; Delivet et al., 2016), in the Belgica Mound Province (Van Rooij et al., 2007b; Huvenne et al., 2009b), in the Hovland Mound Province (Dorschel et al., 2005), on west Porcupine Bank (Øvrebø et al., 2006; O'Reilly et al., 2022), in the eastern Rockall Trough (Øvrebø et al., 2005), and on the Donegal-Barra Fan (Howe, 1996; Knutz et al., 2002; Tarlati et al., 2020): widespread contourite deposition with a higher carbonate component from primary productivity alternates with IRD-containing hemipelagites to make up the Quaternary sediment stack from Goban Spur to the northeastern Rockall Trough. Locally, bottom currents have been strong enough to interrupt deposition or cause erosion (Øvrebø et al., 2006; O'Reilly et al., 2022; Matossian and Van Rooij, 2024). Lastly, an alongslope moat was identified along the northern flank of Porcupine Bank by Sacchetti et al. (2011) at water depths 600-850 m, i.e., at the interface between the Mediterranean Outflow Water and the overlying Eastern North Atlantic Central Water.

In specific settings, the dominance of alongslope processes on deposition is interrupted by gravitational processes: turbidites are found in Porcupine Bank Canyon (Øvrebø et al., 2005), gullies and canyons are present along the eastern margin of the Rockall Trough (Sacchetti et al., 2011), and slope instabilities have led to the formation of slide scars and deposition of debrites at the bases of mounds or slopes (Øvrebø et al., 2005; Sacchetti et al., 2011; Tarlati et al., 2020). As in the Gollum and Kings channel systems, downslope activity further north along the Irish margin was dominant when ice was close to the shelf edge during periods of lowered sea levels (Øvrebø et al., 2005; Tarlati et al., 2020). From the discussion above (section 6.2.1), however, it is clear that the alongslope current is important for sediment dispersal across the margin both when downslope processes are active and when they are not.

6.3 Outlook

The findings presented in this thesis demonstrate that a submarine canyon-channel system that does not actively host large-scale gravity flows because it is disconnected from terrestrial sediment sources in the present-day high sea-level environment can serve as a valuable palaeoclimate and palaeoceanography archive. Both aspects could be explored further and serve as complementary records to those built from the Belgica Mound Province, especially for time intervals when intense bottom currents in the Belgica Mound Province made it a dominantly non-depositional or erosional environment. Further research might also focus on the main

outstanding uncertainties within the sedimentary environment of the Gollum and Kings channel systems, i.e., the temporal framework of the downslope activity, and the processes that trigger or allow downslope activity to occur.

6.3.1 Gollum Channel System as an archive for palaeoclimate and ice dynamics

6.3.1.1 Processes leading to downslope activity

There seems to be a missing link between glacial processes on the inner/middle shelf south of Ireland (Giglio et al., 2021, 2022) and sediment delivery to the shelf edge of the Gollum and Kings channels. The seafloor on the outer shelf looks smoothed in bathymetry data, and very rare and few discontinuous reflectors are present below a high-amplitude seafloor reflection in seismic data (unpublished data). Any subsurface structures that could connect the two areas were likely either eroded by glacial processes themselves or subsequent reworking by shelf processes, and otherwise obscured by the limited penetration potential of high-resolution seismic data.

The Gollum channel floor deposits might help advance the investigation of the dynamics of the southwestern edge of the British-Irish Ice Sheet. Further research on the composition and provenance of the turbidites in the Bilbo channel floor sediment cores (Chapter 4) could elucidate on their source and perhaps on ice flow directions and meltwater pathways from the hinterland to the shelf edge. A similar investigation in one of the Kings channels would allow comparison between the two systems and might lead to a more robust explanation of why they have such different geometries: are their differences indeed linked to different sediment sources, or are there other factors to consider? Furthermore, provenance studies on post-Last Glacial Maximum turbidites within the channels may clarify on the process responsible for initiation of gravity flows after ice had retreated from the shelf edge: do they result from slope failure and subsequent flow transformation within the channels or channel heads, or was sediment delivery from the inner to outer shelf still efficient?

Further investigation of the channel head bedforms (Chapter 4) might also be interesting in this investigation, as it could reveal which processes create downslope flows at the shelf edge during highstand conditions of low sediment supply. Moored current measurements in combination with a sediment trap could provide an estimate of the frequency, volume, and composition of present-day activity in the channel head area.

6.3.1.2 Timing of downslope activity

Large-scale timing of activity in the channel systems is still unclear. This is partly due to a degree of uncertainty associated with the age of the sediments at the top of the seismic unit underneath the RD1 reflector that contains the first indications of channel presence (Chapter 3). Better age control here would help verify the hypothesis of early Pliocene margin tilting as the process that started the incision of the system.

Additionally, there is uncertainty about the length of the RD1 hiatus outside of the Belgica Mound Province, and, therefore, the timing of Quaternary levee build-up (as an indicator for downslope activity) in the channel systems. Interesting to consider in this regard is the difference in climatic cyclicity between the Early Pleistocene (c. 40 kyr cycles of smaller amplitude) and the Middle-Late Pleistocene (c. 100 kyr cycles of larger amplitude) (Lisiecki and Raymo, 2005). The less extreme sea-level decreases in the early Quaternary (Lisiecki and Raymo, 2005) are unlikely to have exposed the Celtic Sea Shelf and provided direct terrestrial input to the channels. Still, gravity flows are thought to have been active and contributed to the erosion that created the RD1 reflector (most obvious in the Kings Channel System; Chapter 2), so what is the process behind early Quaternary activity? Earliest Quaternary IRD has been identified in the Belgica Mound Province and was attributed a British Isles source (Thierens et al., 2012). This suggests that marine-terminating ice sheets with calving margins were present near the Porcupine Seabight. Although it is unlikely that these early Quaternary ice sheets extended all the way to the

southwestern shelf edge for a direct connection with the channel systems (Thierens et al., 2012), they may have contributed to sediment build-up on the shelves surrounding Ireland, which in turn may have triggered downslope activity in the channels.

In this hypothesis, the connection between early Quaternary terrestrial and/or marine-terminating ice sheets in southern Ireland and the Kings Channel System (closer to southwest Ireland) may have been more efficient than that to the Gollum Channel System. If the Gollum Channel System received input mostly from the Celtic Sea Shelf (here inferred from the orientation of its branches), this connection may have been missing as the shelf was largely drowned and not covered by an ice sheet (i.e., similar to the early Holocene situation). Hence, an alternative – or additional – explanation for the differences between the two channel systems discussed in Chapter 2 may be a difference in timing of the main phase of downslope activity in the two systems. Potentially, both channel systems were restricted to the upper slope in the early Quaternary and large-scale downslope flows produced by the large catchment area of the Celtic Sea Shelf during the more extreme late Quaternary glaciations allowed the Gollum Channel System to build out further towards the Porcupine Abyssal Plain, as well as to the south with additional branches formed there.

This hypothesis could be tested by investigating the deeper reaches of the Gollum Channel System in seismic data and verify whether they are as old as the upslope sections. Sediment cores in the thalwegs could also be helpful in this regard. Since only the largest gravity flows would have reached the deepest parts of the system, the record can be expected to be condensed and even short cores (several m) may contain a long time interval of downslope activity. Lastly, more detailed analysis of the timing of incision of the southern branches of the Gollum Channel System could help establish when this part of the system became active and whether they were eroded by extensive late Quaternary flows, or if a connection to the shelf existed earlier.

6.3.2 Gollum Channel System as an archive for palaeoceanography

The Quaternary palaeoceanography in the basin and along the Irish margin is dominated by the European Slope Current carrying Mediterranean Outflow Water. Variability in contourite characteristics in sediment cores suggest this water mass has varied both in core flow strength and in vertical position within the water column. From the data presented here as well as earlier research along the flow path, there is a link with glacial-interglacial variability (see references in section 5.5). The water depth where the Mediterranean Outflow Water is present, is tied to the relative densities of the intermediate water masses in the North Atlantic (Rogerson et al., 2012b). As the density of the Mediterranean Outflow Water itself is dependent on evaporation and freshwater input into the Mediterranean Sea (Bahr et al., 2015), its settling depth in the water column is not only dependent on climate but also influences climate through the role it plays in the Atlantic Meridional Overturning Circulation (Toucanne et al., 2021). Understanding how this water mass shifted in the past may thus allow a better understanding of (future) climate patterns in the North Atlantic and their effects on oceanic processes. The temporal variability in flow strength as well as the changes in water depth of the Mediterranean Outflow Water in the Porcupine Seabight interpreted from the sediment cores presented in Chapter 5 could be verified, improved, and explored over longer timescales. The age model of the existing cores can be improved with additional radiocarbon dates (on foraminifera) and a more robust correlation from additional XRF data. A more in-depth IRD analysis with accurate flux calculations and provenance study would be beneficial in this regard as well. Additionally, this could improve the understanding on the source of iceberg input to the area throughout the last deglaciation and elucidate on surface ocean currents. To gain insight into glacial-interglacial variability of the water mass dynamics through the (late) Quaternary, longer sediment cores could be correlated to high-resolution seismic or subbottom profiler data. This type of correlation was not possible with the cores presented in Chapter 5, as they encompassed only two seismic reflections in subbottom

profiler data. To better estimate the water depth interval over which the Mediterranean Outflow Water shifted, additional cores at water depths between the upper-slope and mid-slope cores presented here should be included in the analysis. Furthermore, inclusion of additional cores at greater water depths may constrain the lower boundary of the shift. This could also increase the understanding of the processes acting at the lower boundary of the Mediterranean Outflow Water, at the interface of the Labrador Sea Water. As a density gradient is present there as well, internal waves are expected to form (e.g., Hanebuth et al., 2015) and influence sedimentation but this has not been investigated in the Porcupine Seabight.

6.3.3 Further questions arising from this work

The Gollum Channel System is another example of a mixed system where alongslope and downslope processes have interacted in the past (e.g., Gong et al., 2018; Fuhrmann et al., 2020; Georgiopoulou et al., 2021; Mencaroni et al., 2021). It has evolved from a downslope-dominant system when the channels were incised, into a system where both processes interact, and finally, one where the alongslope current is a dominant source of sediments to build the eastern Porcupine Seabight slope sequence (Chapter 3, Chapter 5). This highlights the need for incorporation of the distribution of sediments by alongslope currents in source-to-sink studies and models. Additionally, the alongslope current in the Gollum Channel System is crucial for the development of the cold-water coral mini-mounds inside the channels, both by supplying the material to build the mounds, and by enabling the enhanced bottom currents by bringing two water masses with contrasting densities into the basin (Chapter 4). These mini-mounds are quite uncommon in the location that they have formed. Cold-water coral habitats in active canyon-channel systems are usually found on steep rocky outcrops (e.g., Mortensen and Buhl-Mortensen, 2005; Orejas et al., 2009; Huvenne et al., 2011; Brooke and Ross, 2014). Could these mound structures or other deep-water habitats be expected to develop in similar systems where downslope activity is limited but alongslope processes promote enhanced bottom currents on the channel floors? And what would be the required balance of enough food supply but not too much sediment that it would wipe out the habitat?

The Pliocene age of the Gollum Channel System also brings up questions about the controls on the formation of canyon-channel systems. If the Gollum Channel System was indeed formed during a phase of margin tilting along the Northeast Atlantic margin from Norway to Ireland, as is found here (Chapter 3), then how come no other large-scale canyon-channel systems were formed further north along the margin at this time? Offshore Shetland and Scotland, this tilting episode resulted in the progradation of sediment wedges instead (Dahlgren et al., 2005; Stoker et al., 2005b). This may be related to the palaeoslope angle (Piper and Normark, 2009; Bernhardt and Schwanghart, 2025), the process of sediment delivery at the coast (channelised on a fluvial-dominated coast versus not channelised on a wave-dominated coast; Paumard et al., 2020), the rate of sediment delivery (e.g., Ó Cofaigh et al., 2004), or the water-sediment ratio and thus the process by which sediment is deposited on the slope (e.g., slides and debris flows on prograding wedges versus turbidity currents in canyon-channel systems, Weaver et al., 2000; Sejrup et al., 2005). During the Quaternary, deposition in glaciogenic sediment wedges continued north of the Porcupine Seabight, though in a more aggradational manner than during the Pliocene, while contouritic deposition with a share of downslope input prevailed in the Porcupine Seabight (Stoker et al., 2005b; Matossian and Van Rooij, 2024). This has been suggested to be due to a larger influence of meltwater sorting sediment before bringing it to the shelf edge at the southern edge of the ice sheet, versus more direct input from glacial processes (proximal ice margin) to the shelf edge further north (Sejrup et al., 2005). A further implication of the Pliocene age of the Gollum Channel System (i.e., before the onset of major glaciations in the northern hemisphere – let alone the mid-latitudes of the Irish margin) is that the particle input from glacial systems is not the main control for the distribution of canyons on this margin. Moreover, most large-scale canyon-channel systems are present south of the formerly glaciated margin (Weaver et al., 2000).

This is contrary to the Scotian margin in the Northwest Atlantic, where numerous incisions resulted from lowered sea levels in the late Pliocene and sediment delivery through glaciofluvial systems (Piper and Normark, 1989). In general, it seems not only a high amount of sediment delivery but specifically in a concentrated/channelised way may be a main control on canyon-channel system initiation. Many canyon-channel systems host enhanced bottom currents in the present-day highstand that may be crucial in maintaining a dynamic environment at their floors preventing infill and burial after the initial formation of a system. This hints at another potential link with alongslope currents and the presence of specific water masses at the continental slope.

In general, very little is still known about the deeper reaches of the Gollum Channel System (Kenyon et al., 1998). From bathymetry and backscatter data, a deep-sea fan seems to be present but is very limited in extent (Akhmetzhanov et al., 2003). Given the large scale of the system as well as the long period over which it has existed (Chapter 3), this is unexpected. Even systems built out by Quaternary glaciomarine processes have generally more extensive fans, and activity in the Gollum Channel System started about 1 Myr before the onset of northern hemisphere glaciations (e.g., Bourillet et al., 2003; Ó Cofaigh et al., 2004). The possibility exists that the limited extent of the fan is due to the action of bottom currents on the Porcupine Abyssal Plain redistributing sediment further north (Rice et al., 1991). At the time of channel formation (early Pliocene; Chapter 3) when erosional gravity flows were presumably bringing large amounts of material into the deep sea, strengthened North Atlantic Deep Water formation caused erosion in deep marine settings (Stoker et al., 2005a). On the other hand, the thin turbidites in the upper slope channel sections (Chapter 4) suggest the run-out distances of any turbidity currents coming through the system were not very large. Although the Irish-Sea Ice Stream drained a large sector of the British-Irish Ice Sheet and directed material towards the Celtic margin (e.g., Bourillet et al., 2003; Van Landeghem et al., 2020), the east-west orientation of the valleys described by Giglio et al. (2021, 2022) on the Celtic Sea Shelf imply that material must have been brought to the Porcupine margin as well. Does the absence of large turbidites in the Gollum Channel System imply that very little sediment was supplied to this section of the margin, or were other processes on the shelf responsible for preventing transport from the ice margin towards the channel heads? If sediment was diverted northwards, could this be responsible for the formation of the Kings Channel System just to the north of an existing downslope pathway that was the Gollum Channel System – although a fan was not identified at the foot of this system either? Is all of this just an effect of a rapid retreat of the ice sheet to the inner shelf (Clark et al., 2022) with the sediment mostly stored on the shelf rather than transported to the deep sea? Or was the system most active before the period covered by the sediment cores presented in Chapter 4 and/or in other branches than studied here, and is a more extensive sampling survey with longer cores and/or in additional branches necessary to fully assess the sediment supply through the system?

6.4 Final thoughts

Across the research chapters presented in this thesis, it has become clear that small-scale features (e.g., ripples, grain-size variability in sediment cores) that can only be recognised in high-resolution datasets can provide detailed insight into large-scale processes (e.g., tidal bottom currents, alongslope current variability). It is, however, important to keep in mind that these small-scale features are highly susceptible to local effects such as topography (e.g., a current meter placed next to a cold-water coral mini-mound is likely to record a different current regime from one placed in a setting with a flat seafloor). They are also more likely to change over shorter timescales than large-scale features identified in lower-resolution datasets. Therefore, it is worth remembering that such small-scale features might represent a snapshot rather than the average sedimentary and oceanographic conditions of their environment. This emphasises the need for an integrated approach to (marine) research where strategically placed high-resolution point measurements can provide ground-truthing to lower-resolution large-scale datasets. Vice versa, the large-scale datasets are necessary to put high-resolution data in their wider context. As is

illustrated here, such a combination of methods allows for an attempt at unravelling the interplay between oceanic processes acting on continental slopes and gaining some understanding of their complexity. To sum up: it's complicated.

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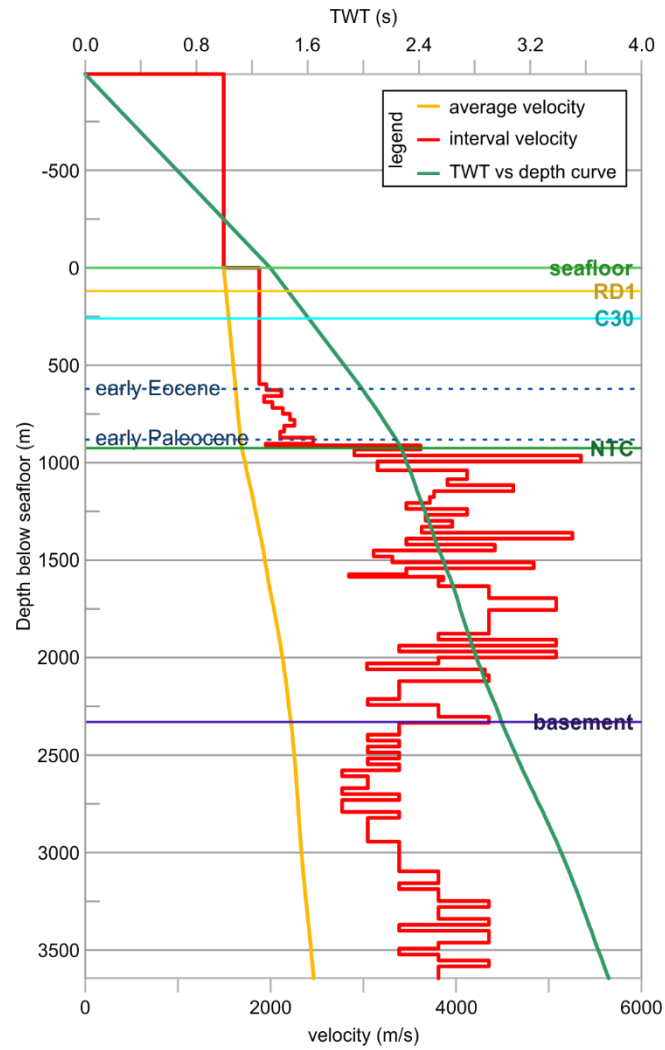
Appendices

Appendix A

Appendix A. A brief overview of the processing steps used for each seismic dataset used in Chapter 3. More details can be found on the website of the Irish Department of Communications, Energy and Natural Resources (Department of Communications, Energy and Natural Resources, 2010).

survey	company	low- / high-cut frequency (Hz)	processing steps
1997/01	Nopec Geoventures	3 / 218	bandpass filter multiple attenuation deconvolution before stacking dip moveout stacking deconvolution after stacking migration time-variant filter
1997/02	Geco-Prakla	3 / 102	FK-filter deconvolution before stacking dip moveout stacking deconvolution after stacking time-variant filter migration FK-filter
1997/03	Fugro-Geoteam	none / 180	signature deconvolution resampling dip moveout stacking migration FK-filter bandpass filter
1997/06	Spectrum	3 / 125	resampling dip moveout multiple attenuation designature deconvolution before stacking stacking

Appendix B



Appendix B. Velocity model used for the time-depth conversion at the location of well 62/7-1. The depths of boundary reflectors seafloor, RD1, C30, NTC and basement are also shown, as well as the depth of the uppermost early Eocene and early Paleocene samples.

Appendix C

Details on Bayesian age-depth modeling for core B2306-GC02

- input data

lab ID	Age	Error	Depth	cc
Beta-699043	2810	30	78	2
Beta-699044	8370	30	237	2

- R code

```
library(rbacon)
```

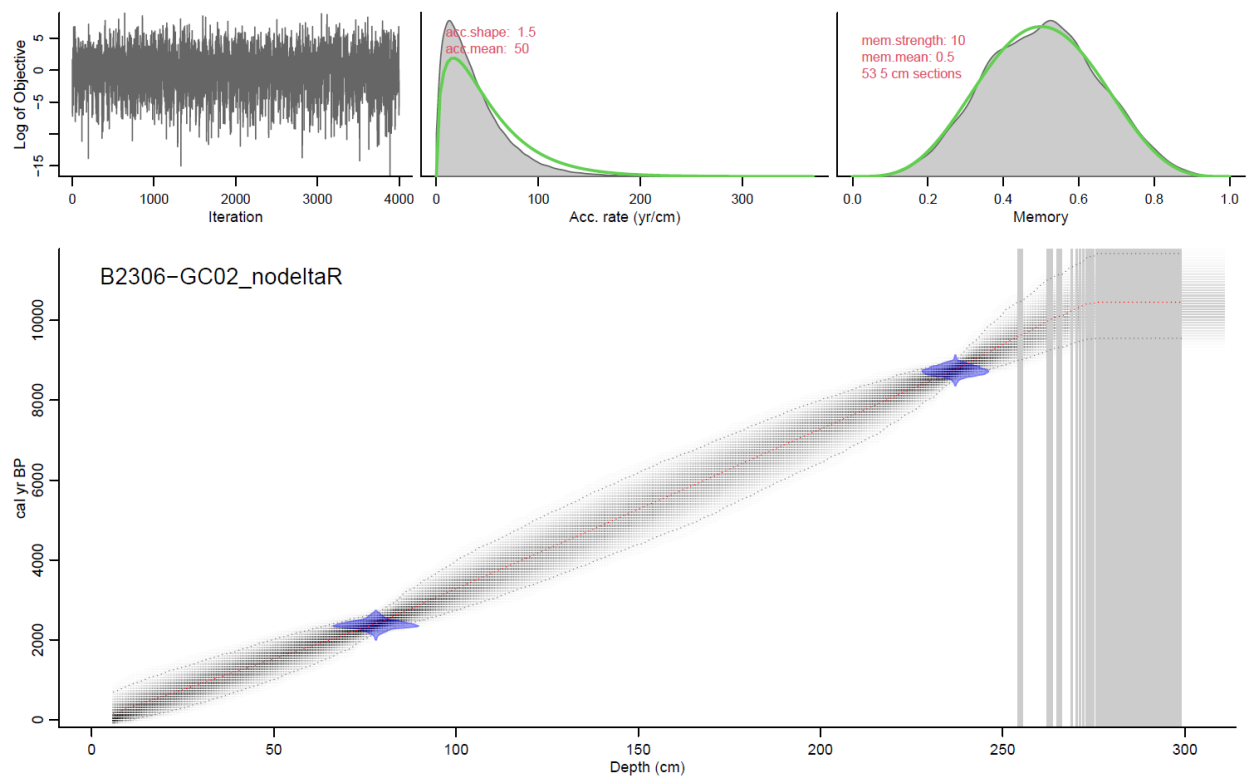
```
slumpsGC02
c(299,275,275.6,274.04,273.9,273.5,273.44,273.24,273.2,272.94,272.14,272,271.26,271,270.
4,270.24,268.96,268.8,266.08,264.84,263.58,263.44,262.78,262.64,262.28,262.2,255.42,255.
3,255.2,255.1,255,254.84,254.78,254.76,254.48,254.3)
```

```
Bacon("B2306-GC02_nodeltaR", d.min=6, d.max=299, slump=slumpsGC02)
```

```
Bacon.hist(254)
```

```
Bacon.hist(274)
```

- results



Details on Bayesian age-depth modeling for core B2306-GC04

- input data

lab ID	Age	Error	Depth	cc
Beta-699045	5720	30	104	2
Beta-699046	14540	40	168	2
Beta-699047	17300	50	232	2

- R code

```
library(rbacon)
```

```
slumpsGC04 <- c(290,263,261.8,259.56,176.9,176.24,161.5,152.3)
```

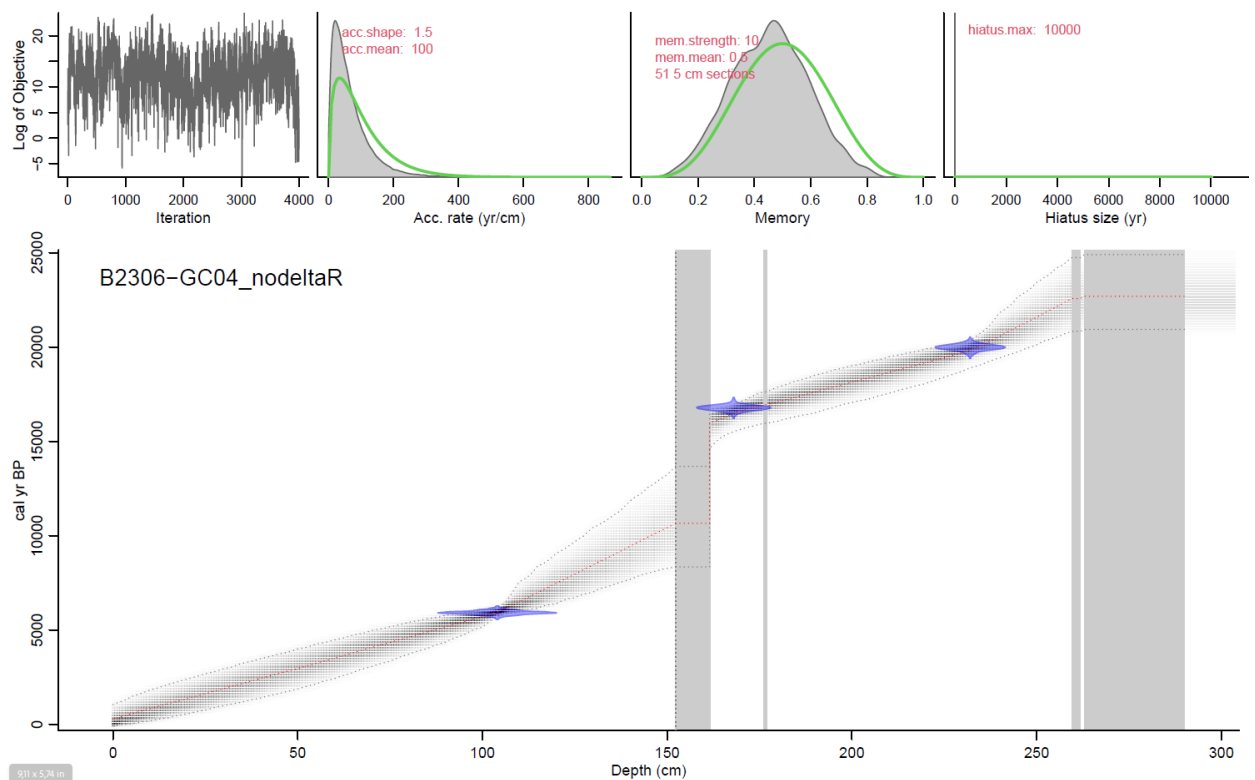
```
Bacon("B2306-GC04_nodeltaR",      hiatus.depths=152.3,      d.min=0,      d.max=290,
slump=slumpsGC04)
```

```
Bacon.hist(152.3)
```

```
Bacon.hist(161.50002)
```

```
Bacon.hist(263)
```

- results



Appendix D

Methods for sediment core XRF scanning and high-resolution grain-size prediction

Cores B2306-GC02 and B2306-GC04 were scanned for their geochemical properties using an ITRAX machine at the Austrian Core Scanning Facility of the University of Innsbruck. The Cr X-ray tube was set to a voltage of 30 kV and a current of 55 mA, while the exposure time was set to 8 s. All core sections were scanned at 0.2 mm resolution. The core halves were covered by a thin polypropylene foil to minimise desiccation and alteration of the sediment surface during measurements. Selected core intervals were measured thrice to test robustness and reproducibility, and only elements where the Pearson correlation coefficient (r) for the repeated measurements was 0.75 or higher in both cores were included in the further analysis: Si, K, Ca, Ti, Fe, Rb, Sr, Zr and Ba. Although Cl also had an r value above 0.75, it is strongly related to pore water chemistry rather than sediment composition (Bertrand et al., 2024), and therefore not included in further analyses. For each core, the element intensities (in counts per seconds; cps) were then centre-log transformed to reduce the unwanted effects of physical properties and matrix effects on the data (Bertrand et al., 2024).

A partial least squares regression was performed to predict mean grain size at the resolution of the XRF measurements (0.2 mm) across the full depths of both cores. In this procedure, the method described in Liu et al. (2019) is implemented in R. The prediction was based on a subset of elements (Si, K, Ti, Fe, Rb and Zr), since the remaining elements (Ca, Sr and Ba) are strongly affected by biogenic processes rather than lithogenic grain size (Jaccard et al., 2010; Liu et al., 2019). The selected element counts were transformed into centre-log ratios and analysed by principal component analysis (PCA). All data points of the two cores plot within a single cluster (Appendix D1), indicating that one predictive model combining XRF data from both cores is acceptable. The mean grain size was calculated for each of the grain-size measurements on both cores. To establish the training dataset, XRF measurements at the depths where grain-size measurements were taken, were averaged over the thickness of the grain-size samples (c. 4 mm). The predictive grain-size model was then built by performing a partial least squares regression on the training dataset, with the mean grain-size data as dependent variables and the XRF centre-log ratios as explanatory variables. Finally, based on this model, the mean grain size was predicted over the full depths of both cores. The Pearson correlation coefficient between the predicted and measured mean grain sizes is 0.72 (Appendix D2). The predicted mean grain sizes are presented in the main text as moving average graphs with bin size 20. All of these steps were coded in R using the `clr`-function from composition package (v2.0-8) (van den Boogaart et al., 2024), the `prcomp`-function from the stats package (v4.4.1; base R) the `plsr`-function from the pls package (v2.8-5) (Mevik and Wehrens, 2007), and the `predict`-function from the stats package (v4.4.1; base R).

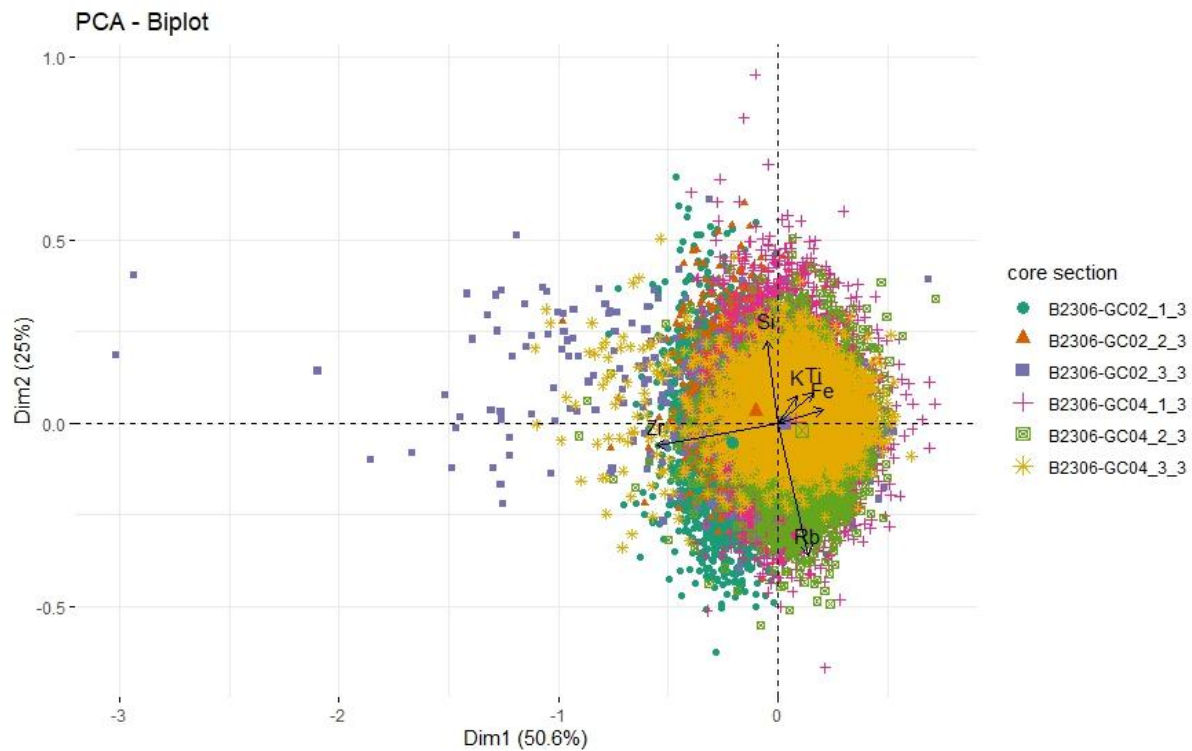
Bertrand, S., Tjallingii, R., Kylander, M. E., Wilhelm, B., Roberts, S. J., Arnaud, F., et al. (2024). Inorganic geochemistry of lake sediments: A review of analytical techniques and guidelines for data interpretation. *Earth Sci Rev* 249, 104639. doi: 10.1016/j.earscirev.2023.104639

Jaccard, S. L., Galbraith, E. D., Sigman, D. M., and Haug, G. H. (2010). A pervasive link between Antarctic ice core and subarctic Pacific sediment records over the past 800 kyrs. *Quat Sci Rev* 29, 206–212. doi: 10.1016/j.quascirev.2009.10.007

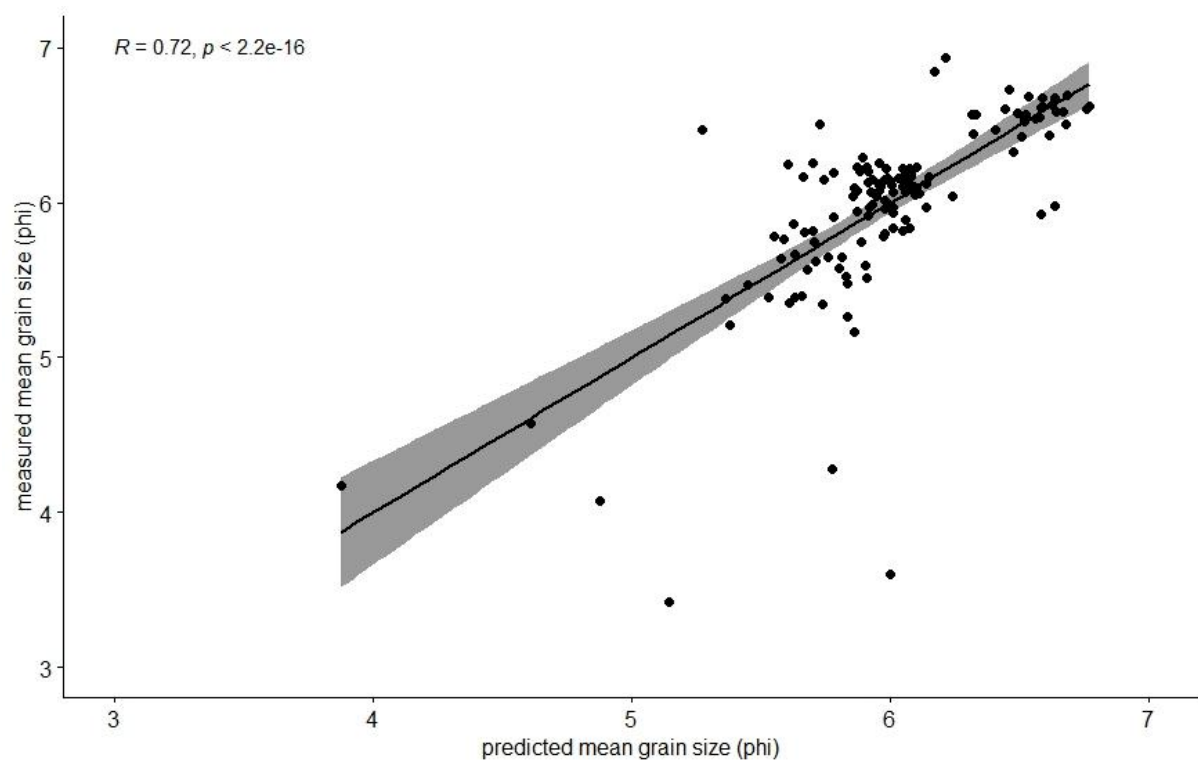
Liu, D., Bertrand, S., and Weltje, G. J. (2019). An Empirical Method to Predict Sediment Grain Size From Inorganic Geochemical Measurements. *Geochemistry, Geophysics, Geosystems* 20, 3690–3704. doi: 10.1029/2018GC008154

Mevik, B.-H., and Wehrens, R. (2007). The pls Package: Principal Component and Partial Least Squares Regression in R. *J Stat Softw* 18, 1–23. Available at: <http://www.jstatsoft.org/>

van den Boogaart, K. G., Tolosana-Delgado, R., and Bren, M. (2024). Homepage of “compositions”, an R package for compositional data analysis. Available at: <http://www.stat.boogaart.de/compositions/> (Accessed November 19, 2024).

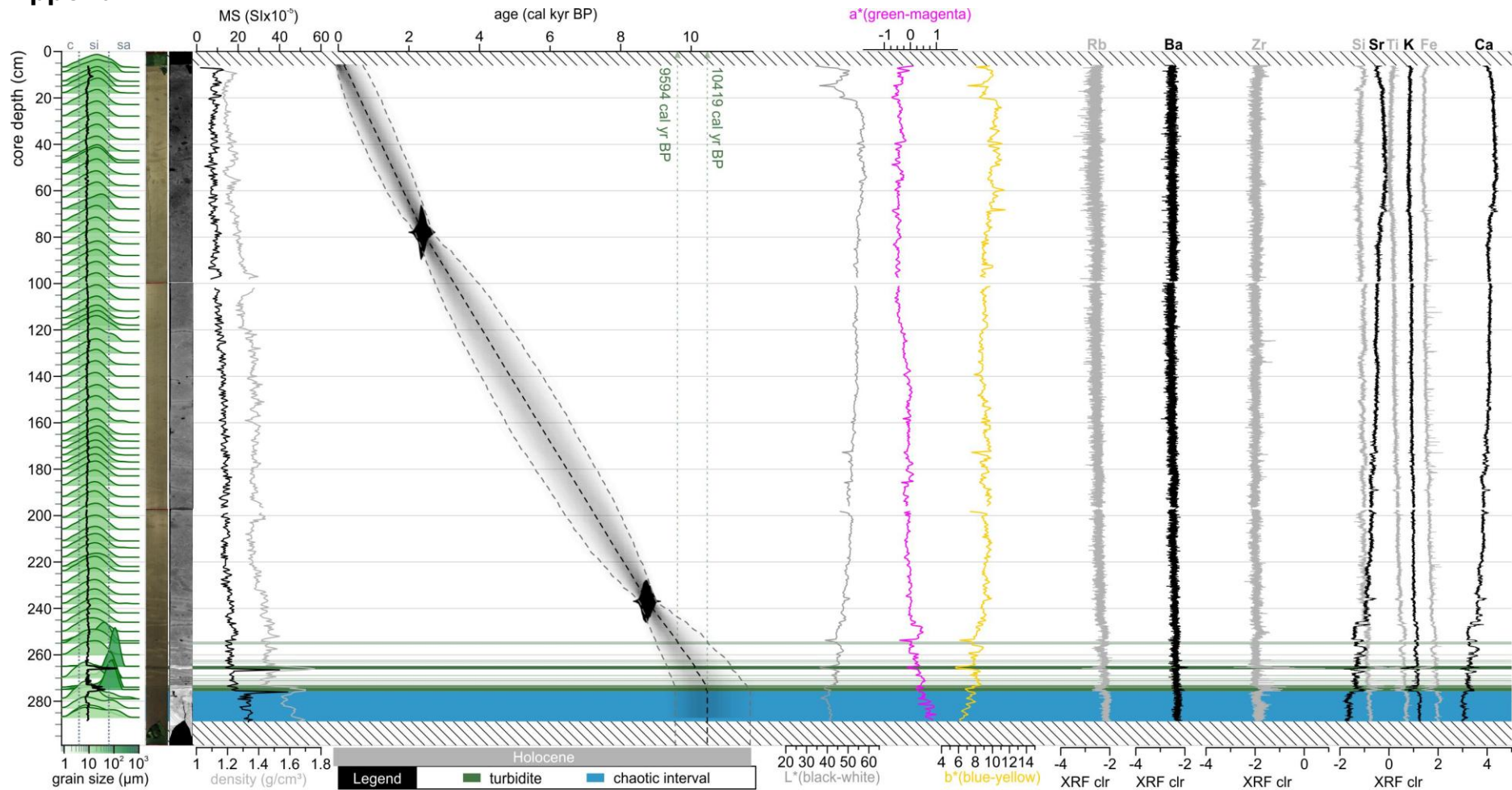


Appendix D1. Biplot of the principal component analysis performed on centre-log ratios of a subset of elements (Si, K, Ti, Fe, Rb, Zr) from XRF measurements on cores B2306-GC02 and B2306-GC04.



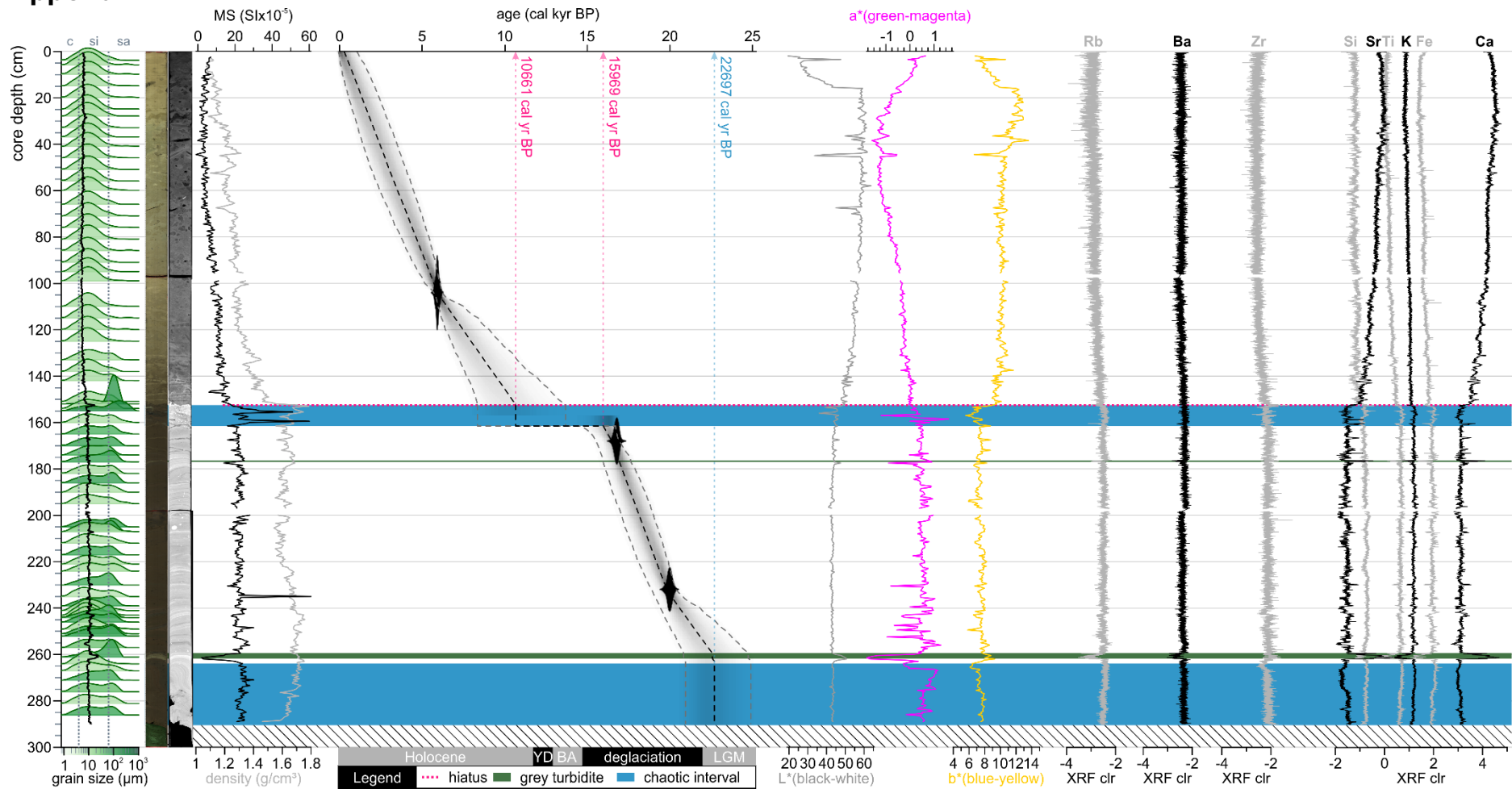
Appendix D2. Correlation plot between the measured mean grain sizes and the predicted mean grain sizes for cores B2306-GC02 and G2306-GC04. The grey area indicates the 95 % confidence interval.

Appendix E



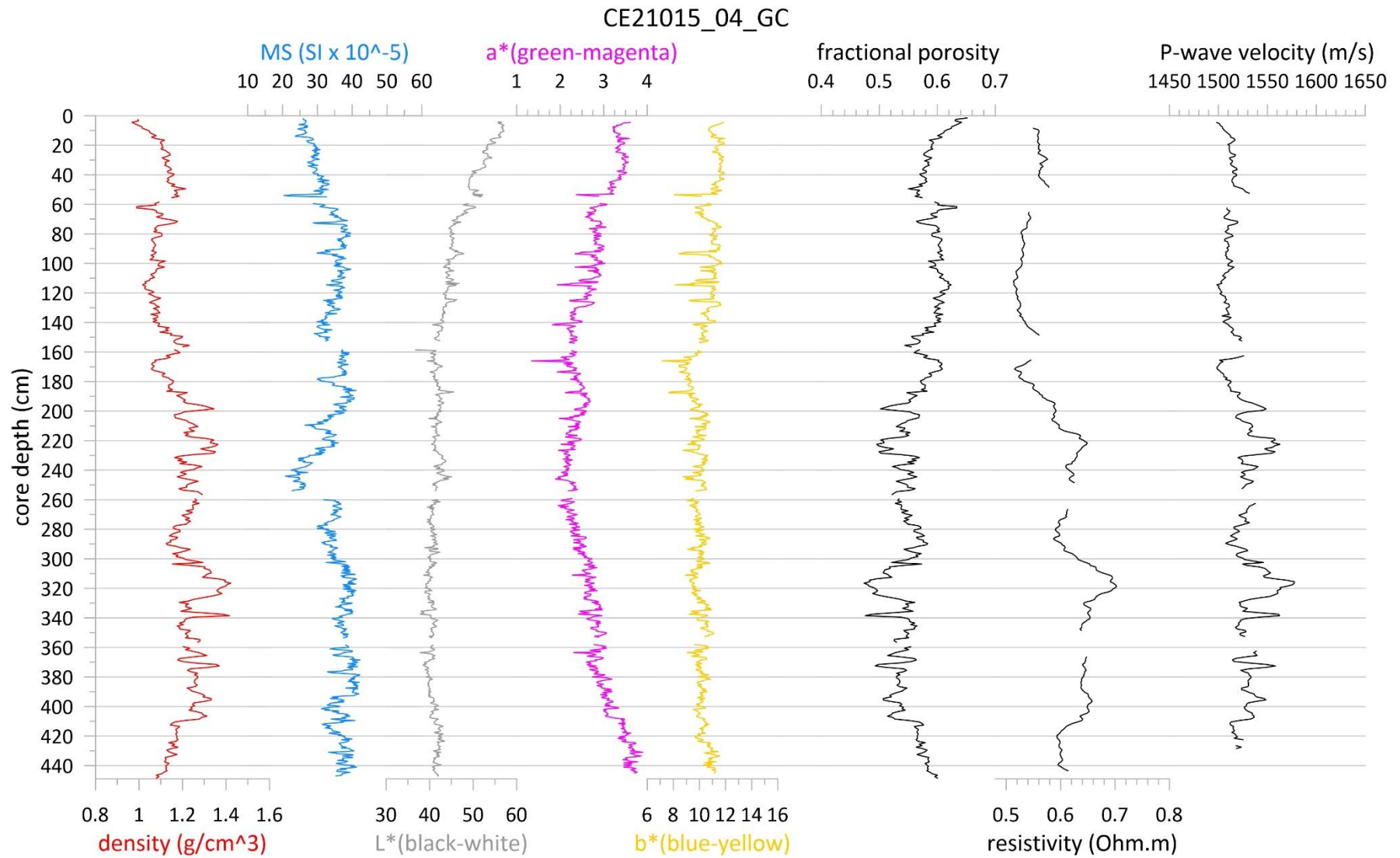
Appendix E. Core B2306-GC02. From left to right: grain-size distribution plots coloured according to the mode (with darker green representing a coarser mode) and mean grain size predicted from XRF data plotted in black; core photograph; CT scan; magnetic susceptibility; gamma-ray density; age-depth model; spectrometry data ($L^*a^*b^*$); XRF centre-log ratios for selected elements (Rb, Ba, Zr, Si, Sr, Ti, K, Fe, Ca).

Appendix F

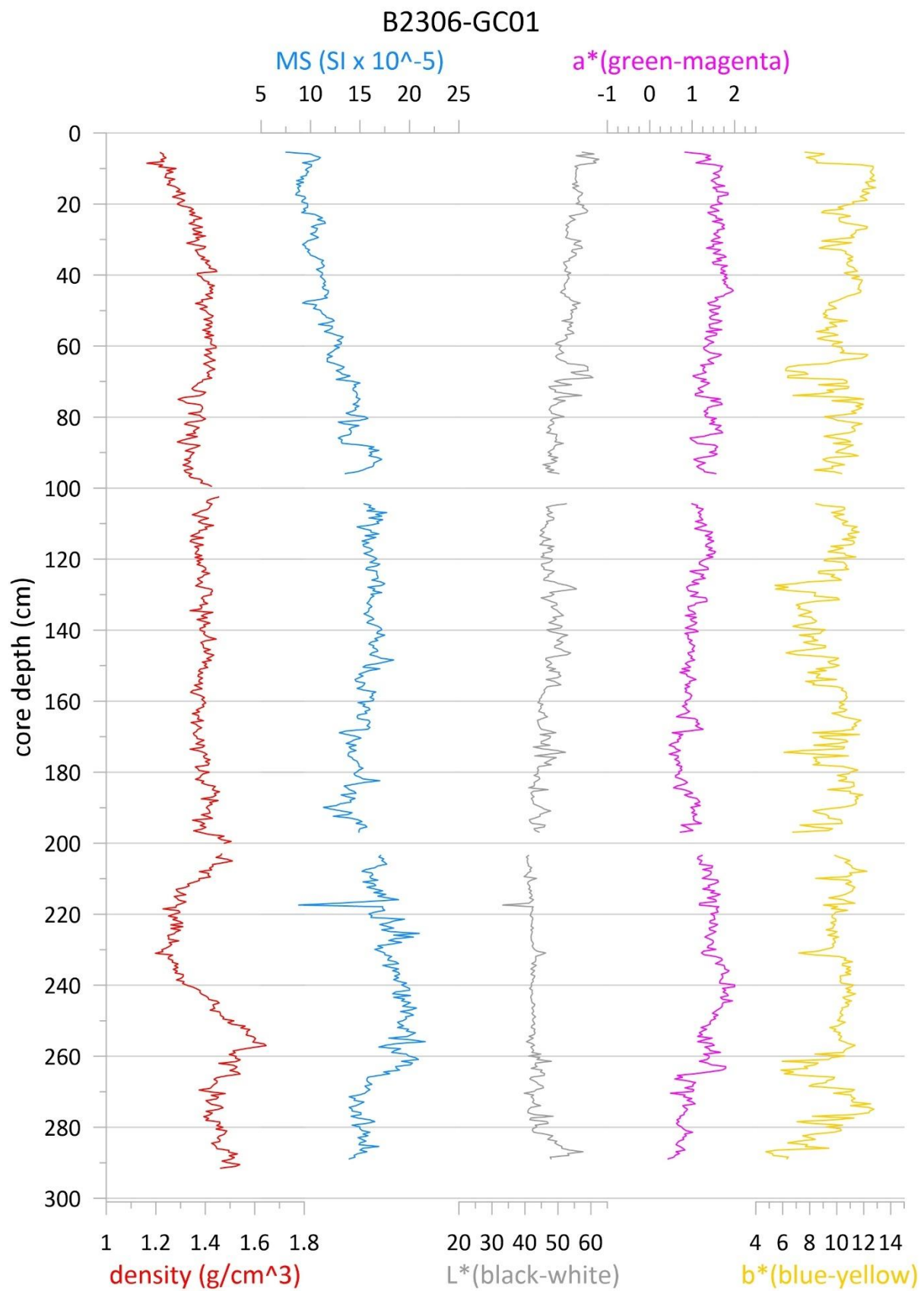


Appendix F. Core B2306-GC04. From left to right: grain-size distribution plots coloured according to the mode (with darker green representing a coarser mode) and mean grain size predicted from XRF data plotted in black; core photograph; CT scan; magnetic susceptibility; gamma-ray density; age-depth model; spectrometry data ($L^*a^*b^*$); XRF centre-log ratios for selected elements (Rb, Ba, Zr, Si, Sr, Ti, K, Fe, Ca).

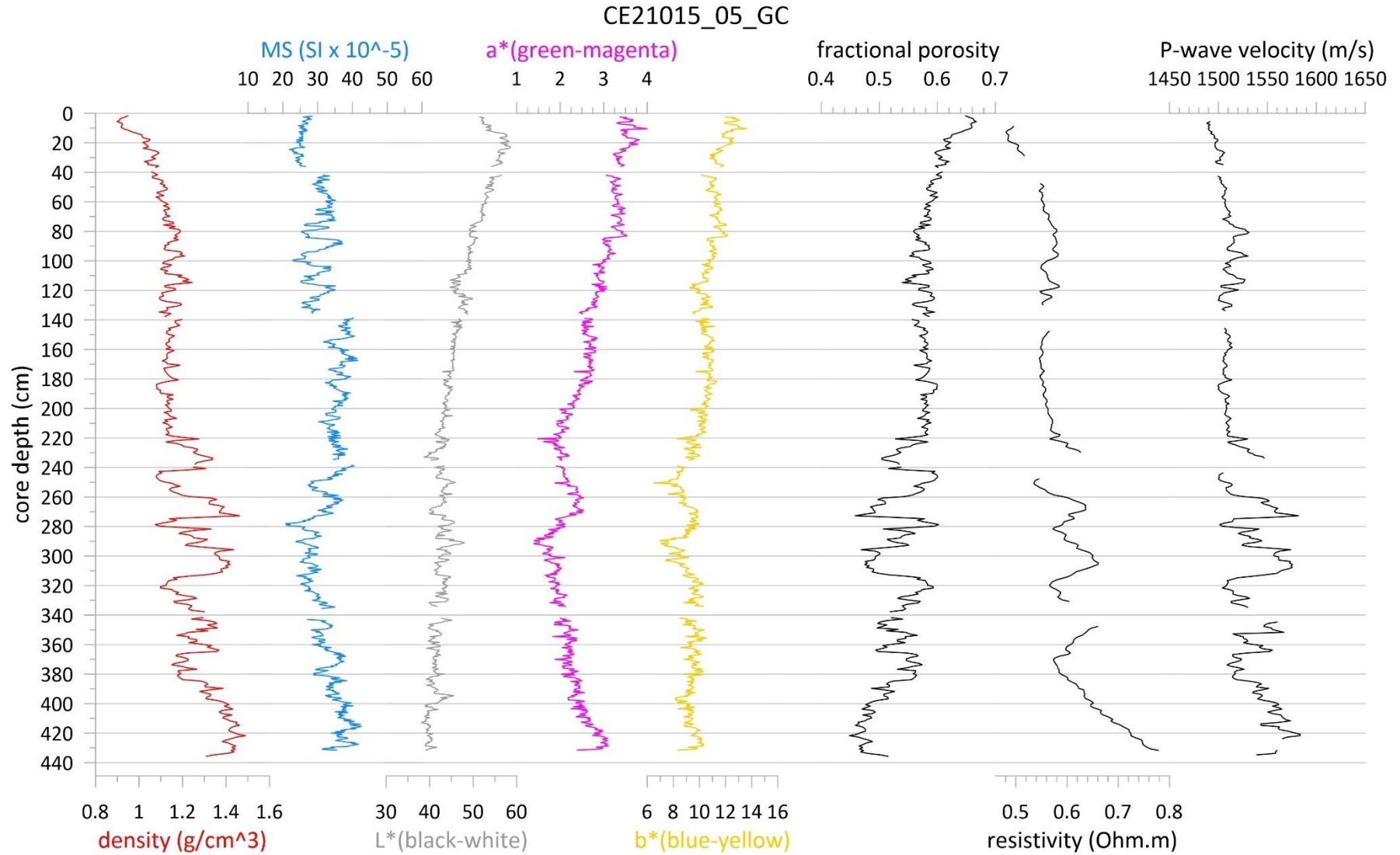
Appendix G



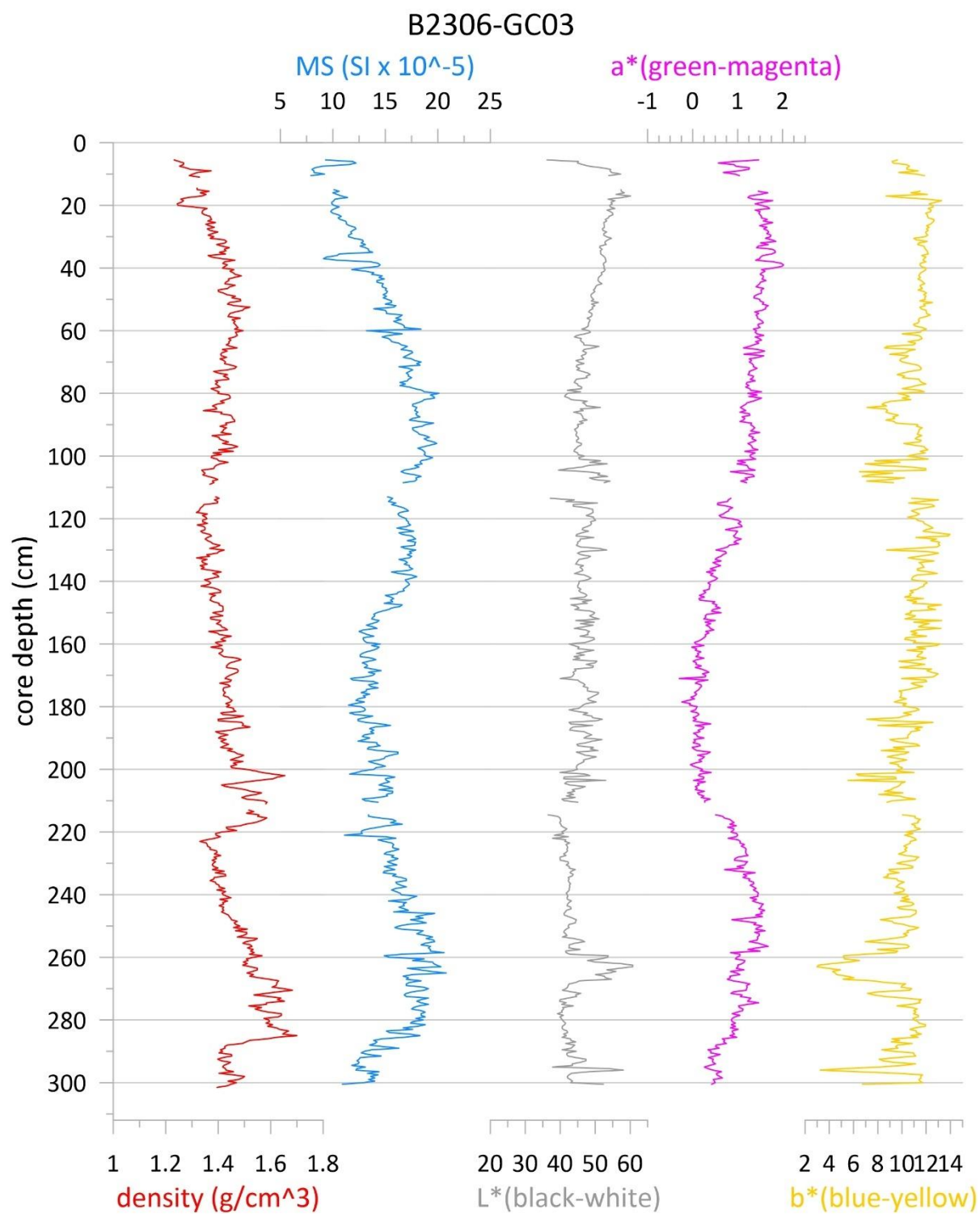
Appendix G1. Core CE21015_04_GC. From left to right: gamma-ray density; magnetic susceptibility; spectrometry data (L*a*b*); fractional porosity; resistivity; P-wave velocity.



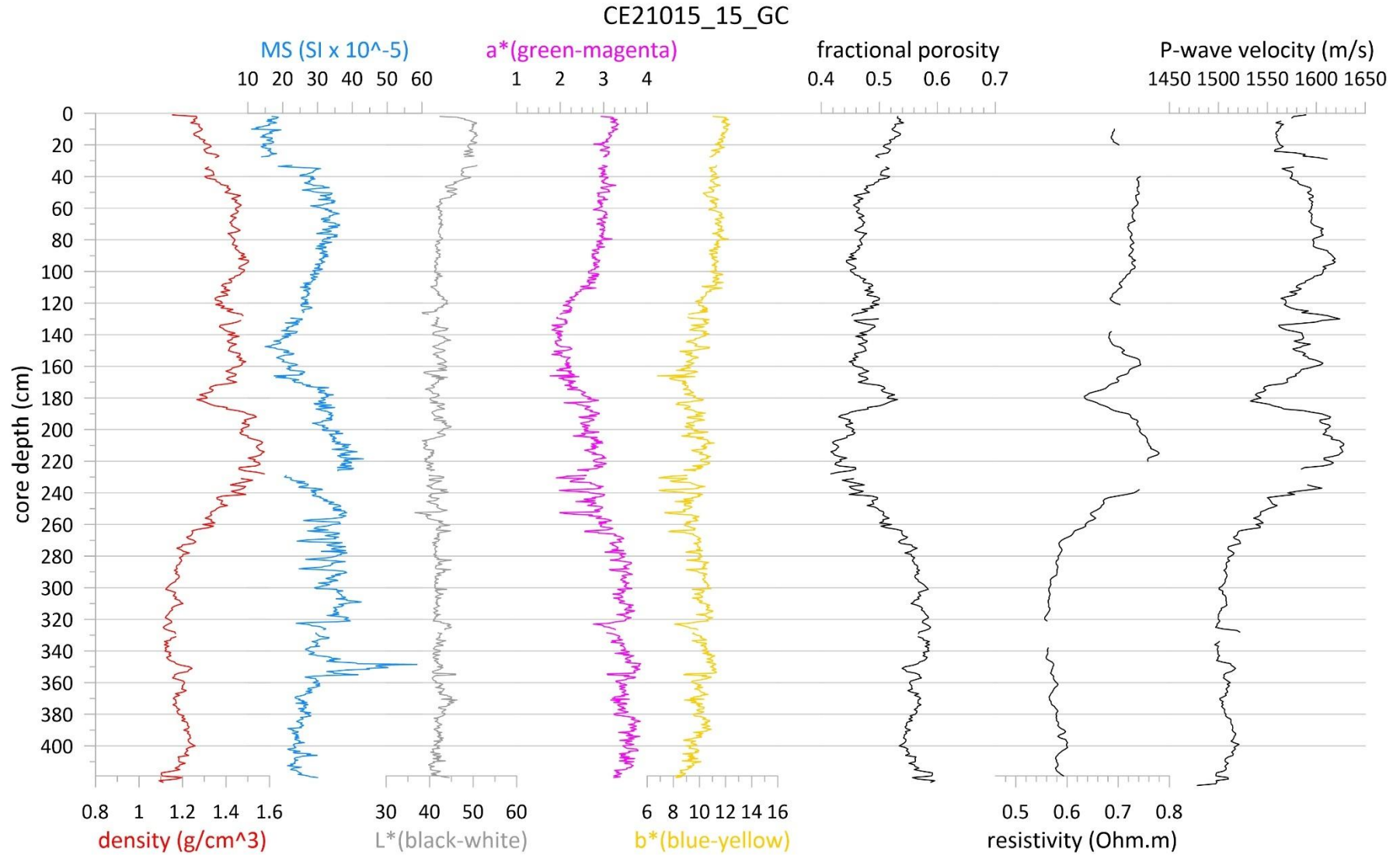
Appendix G2. Core B2306-GC01. From left to right: gamma-ray density; magnetic susceptibility; spectrometry data (L*a*b*).



Appendix G3. Core CE21015_05_GC. From left to right: gamma-ray density; magnetic susceptibility; spectrometry data ($L^*a^*b^*$); fractional porosity; resistivity; P-wave velocity.



Appendix G4. Core B2306-GC03. From left to right: gamma-ray density; magnetic susceptibility; spectrometry data (L*a*b*).



Appendix G5. Core CE21015_15_GC. From left to right: gamma-ray density; magnetic susceptibility; spectrometry data ($L^*a^*b^*$); fractional porosity; resistivity; P-wave velocity.

Appendix H

Appendix H1. Sortable silt calculations for core CE21015_04_GC. SS% = percentage sortable silt; SSmean = mean grain size of sortable silt fraction; rrunx = downcore running correlation value between SS% and SSmean for window length x.

sample depth (cm)	% < 63 µm	% 10-63 µm	SS% (µm)	SSmean (µm)	rrun5	rrun7	rrun9	facies
5	86.94	57.63	66.29	26.34				3
10	87.39	58.52	66.96	26.23				3
20	86.73	57.37	66.16	25.97	0.93			3
25	90.76	58.03	63.94	24.64	0.94	0.86		3
30	87.13	56.76	65.14	25.80	0.83	0.92	0.95	3
35	88.25	57.49	65.15	25.40	0.86	0.95	0.97	3
43	88.41	59.03	66.77	25.76	0.96	0.97	0.97	3
52	91.07	58.00	63.69	24.70	1.00	0.98	0.95	3
63	93.16	56.39	60.53	23.82	1.00	0.97	0.94	3
67	93.60	55.65	59.46	23.29	0.90	0.95	0.96	3
79	93.00	57.45	61.77	24.05	0.89	0.90	0.95	3
83	92.70	58.18	62.76	23.90	0.92	0.91	0.92	3
89	92.08	58.52	63.55	24.28	0.66	0.92	0.88	3
93	91.89	57.85	62.96	24.36	0.86	0.79	0.92	3
99	92.06	58.47	63.51	24.48	0.90	0.94	0.92	3
104	89.38	57.15	63.94	24.81	0.97	0.95	0.94	3
110	91.69	57.28	62.47	23.80	0.96	0.96	0.94	3
115	92.77	55.69	60.03	23.18	0.95	0.95	0.94	3
120	90.51	54.93	60.69	23.60	0.85	0.94	0.97	3
125	91.59	55.99	61.14	23.54	0.99	0.98	0.98	3
130	89.58	55.68	62.16	24.22	0.99	0.99	0.95	3
140	84.20	54.04	64.19	25.45	0.99	0.96	0.90	2
149	77.82	52.07	66.91	26.62	0.91	0.93	0.93	2
160	71.85	47.78	66.50	27.01	0.99	0.97	0.93	2
165	79.80	50.48	63.26	25.81	1.00	0.99	0.94	2
170	84.33	42.46	50.34	21.85	0.99	0.97	0.95	3
177	85.23	44.49	52.20	22.24	0.95	0.97	0.97	3
183	73.70	39.55	53.67	23.44	0.98	0.96	0.97	3
187	68.91	38.99	56.58	24.93	0.96	0.97	0.97	2
192	68.82	41.22	59.89	25.68	0.97	0.96	0.98	2
197	68.94	45.27	65.66	26.82	0.99	0.98	0.97	2
201	73.81	47.60	64.48	26.41	0.95	0.96	0.96	2
204	75.98	48.15	63.37	26.09	0.93	0.88	0.89	2
210	65.28	44.50	68.16	27.88	0.85	0.80	0.82	2
215	66.60	44.57	66.93	26.81	0.76	0.78	0.77	2
218	63.24	41.86	66.20	27.71	0.22	0.73	0.79	2
221	56.39	37.94	67.27	28.65	0.28	0.31	0.88	2
223	64.27	42.29	65.80	27.72	0.76	0.87	0.89	2
225	63.46	41.98	66.15	27.32	0.97	0.97	0.94	2
227	61.05	40.97	67.12	27.90	0.98	0.98	0.98	2
231	80.99	50.38	62.21	25.14	0.99	0.99	0.97	3

Appendix H

235	74.80	46.65	62.36	25.66	0.99	0.99	0.97	3
239	83.56	50.66	60.63	24.61	0.94	0.98	0.98	3
242	78.40	49.03	62.54	25.52	0.96	0.96	0.98	3
246	75.81	48.83	64.41	26.07	0.97	0.97	0.97	3
250	76.40	48.53	63.51	25.76	0.94	0.98	0.97	3
260	69.08	44.78	64.82	26.72	0.95	0.95	0.97	2
265	70.59	45.93	65.07	26.84	0.98	0.98	0.97	2
270	70.71	46.07	65.16	26.71	0.99	0.99	0.98	2
274	74.66	47.76	63.97	26.11	0.99	0.99	0.98	2
280	81.29	50.45	62.06	24.82	0.98	0.98	0.97	3
285	82.83	51.63	62.33	24.70	0.95	0.98	0.98	3
290	78.83	49.24	62.46	25.20	0.99	0.99	0.99	3
295	79.37	51.11	64.40	25.97	0.99	1.00	0.99	3
300	68.22	45.34	66.46	27.09	1.00	1.00	0.99	3
305	68.35	45.76	66.94	27.44	1.00	0.99	0.99	2
308	58.59	41.01	70.00	29.12	0.98	0.99	0.99	2
310	65.90	43.89	66.60	27.39	0.98	0.98	0.99	2
312	54.73	37.12	67.83	28.27	0.99	0.99	0.94	2
314	56.75	38.91	68.57	28.41	0.98	0.96	0.94	2
316	60.33	40.79	67.61	27.93	0.93	0.89	0.95	2
318	64.74	43.08	66.54	27.41	0.89	0.90	0.96	2
320	61.69	40.36	65.42	27.40	0.77	0.96	0.97	2
322	67.47	44.73	66.29	27.09	0.93	0.96	0.94	2
326	69.65	45.51	65.33	26.39	0.94	0.90	0.93	3
330	81.59	51.15	62.68	24.95	0.87	0.90	0.93	3
333	82.49	50.60	61.34	24.74	0.83	0.91	0.91	3
336	67.13	42.76	63.70	26.83	0.90	0.86	0.90	3
340	82.74	51.30	62.00	24.92	0.88	0.85	0.92	3
345	81.68	50.26	61.54	24.46	0.87	0.96	0.95	3
350	77.89	49.96	64.15	25.92	0.98	0.95	0.95	3
360	78.57	49.27	62.71	25.62	0.98	0.98	0.94	3
363	57.05	37.95	66.52	28.07	0.97	0.97	0.96	3
365	81.74	50.63	61.94	25.14	0.97	0.94	0.95	3
370	67.46	44.67	66.22	27.25	0.95	0.93	0.93	3
373	82.64	52.15	63.11	25.30	0.94	0.94	0.92	3
376	81.94	52.47	64.03	25.61	0.94	0.87	0.92	3
379	80.87	51.88	64.16	25.71	0.44	0.87	0.92	3
384	77.70	49.95	64.28	26.44	0.67	0.88	0.95	3
389	79.43	50.40	63.45	26.25	0.94	0.95	0.93	3
392	77.73	50.96	65.55	26.97	1.00	0.97	0.97	3
396	86.94	51.68	59.44	24.45	1.00	0.98	0.98	3
400	89.38	52.19	58.39	23.99	0.98	0.98	0.99	3
405	65.12	41.27	63.38	26.07	0.99	0.99	0.99	3
410	88.23	38.90	44.09	21.06	0.99	0.99	0.99	1
415	88.26	37.77	42.79	20.99	1.00	0.99	0.99	1
420	84.39	36.13	42.82	21.08	0.87	1.00	0.99	1

Appendix H

425	85.80	35.31	41.16	20.61	0.98	0.92	1.00	1
430	83.80	35.32	42.15	20.88	0.99	0.96		1
435	64.02	26.92	42.05	20.82	0.97			1
440	87.09	35.21	40.43	20.27				1
445	86.17	36.18	41.98	20.96				1
correlation =	0.902							

facies	SSmean_avg (μm)	SSmean_min (μm)	SSmean_max (μm)	SSmean_median (μm)
F1	20.8	20.3	21.1	20.9
F2	27.1	24.9	29.1	27.1
F3	25.1	21.8	28.1	25.1

Appendix H

Appendix H2. Sortable silt calculations for core B2306-GC01. SS% = percentage sortable silt; SSmean = mean grain size of sortable silt fraction; rrunx = downcore running correlation value between SS% and SSmean for window length x.

sample depth (cm)	% < 63 µm	% 10-63 µm	SS% (µm)	SSmean (µm)	rrun5	rrun7	rrun9	facies
6	89.22	57.81	64.80	24.99				3
8	87.70	57.48	65.54	25.22				3
11	89.23	58.67	65.75	25.02	0.76			3
14	90.93	58.70	64.56	24.39	0.93	0.75		3
17	90.70	59.01	65.06	24.78	0.95	0.86	0.73	3
22	89.86	58.35	64.94	24.68	0.95	0.93	0.94	3
27	88.05	58.62	66.58	25.16	0.91	0.98	0.97	3
32	89.46	58.85	65.78	25.02	0.99	0.98	0.98	3
37	90.67	59.20	65.28	24.63	0.99	0.99	0.99	3
42	91.74	57.66	62.85	23.87	0.98	0.99	0.99	3
47	90.01	58.78	65.31	24.82	0.98	0.98	0.99	3
52	91.01	56.98	62.62	23.95	0.99	0.99	0.98	3
57	90.50	57.42	63.45	24.13	0.99	0.98	0.94	3
62	90.16	58.22	64.58	24.53	0.96	0.95	0.94	3
67	92.85	57.73	62.17	23.68	0.90	0.91	0.90	3
72	91.78	57.96	63.15	24.22	0.91	0.82	0.81	3
77	90.28	57.33	63.50	24.48	0.77	0.84	0.88	3
81	92.07	58.38	63.41	24.20	0.90	0.84	0.93	3
83	92.20	56.80	61.60	23.98	0.90	0.94	0.93	3
88	91.49	55.95	61.16	23.90	0.90	0.94	0.94	3
93	93.32	56.74	60.80	23.43	0.97	0.90	0.92	3
98	93.44	55.85	59.77	23.13	0.86	0.85	0.93	3
103	94.51	56.15	59.41	22.71	0.85	0.91	0.92	3
108	93.63	57.03	60.91	23.21	0.95	0.95	0.91	3
113	93.61	57.38	61.29	23.37	1.00	0.94	0.96	3
118	95.29	55.42	58.17	22.31	0.96	0.97	0.96	3
123	93.57	57.35	61.29	23.45	0.98	0.97	0.96	3
127	92.82	56.35	60.71	23.48	0.98	0.96	0.95	3
132	91.97	57.45	62.46	23.85	0.86	0.96	0.96	3
137	93.12	57.68	61.95	23.67	0.85	0.82	0.96	3
142	93.27	56.90	61.00	23.60	0.82	0.78	0.67	3
146	92.43	57.02	61.69	23.54	0.48	0.64	0.89	3
152	91.00	55.60	61.09	23.51	0.62	0.88	0.89	3
157	91.77	56.15	61.19	23.35	0.89	0.89	0.92	3
162	89.03	55.00	61.78	24.02	0.92	0.93	0.94	3
167	92.16	53.85	58.43	22.82	0.96	0.95	0.94	3
172	88.73	54.82	61.79	23.88	0.99	0.96	0.97	3
177	84.99	54.22	63.80	24.94	0.99	0.98	0.97	3
182	88.01	54.97	62.46	24.16	1.00	0.97	0.97	3
184	86.89	54.91	63.20	24.42	0.98	0.97	0.97	3
188	75.52	50.60	67.01	26.92	0.96	0.97	0.98	2
191	82.39	52.82	64.11	25.59	0.96	0.98	0.98	2

Appendix H

195	81.46	51.86	63.66	25.45	0.98	0.97	0.97	2
199	73.49	49.10	66.81	26.81	0.98	0.96	0.96	2
202	62.52	42.22	67.53	27.61	0.95	0.96	0.97	2
204	73.96	49.32	66.69	26.75	0.91	0.96	0.98	2
206	76.08	50.58	66.48	26.42	0.95	0.99	0.98	2
209	72.18	48.75	67.53	27.12	1.00	0.99	0.99	2
211	82.15	51.69	62.92	25.33	0.99	1.00	0.99	3
216	87.43	47.00	53.76	22.54	0.99	0.99	0.99	3
221	80.28	45.46	56.62	23.82	0.99	0.99	0.99	3
226	65.51	34.16	52.15	22.37	0.95	0.97	0.98	3
231	89.13	45.19	50.71	21.43	0.91	0.86	0.95	3
236	79.09	40.60	51.33	22.26	0.89	0.96	0.96	3
241	80.91	41.51	51.31	22.61	0.97	0.97	0.99	3
246	78.10	41.82	53.55	23.43	0.99	1.00	1.00	3
251	70.64	41.96	59.41	25.35	1.00	1.00	0.98	2
256	71.24	41.02	57.59	24.74	1.00	0.98	0.97	2
257	42.03	31.31	74.49	32.30	0.98	0.97	0.96	2
262	63.45	37.50	59.09	25.32	0.97	0.97	0.95	2
267	78.56	48.80	62.12	25.20	0.98	0.96	0.94	3
271	81.57	48.83	59.87	24.56	0.85	0.95	0.94	3
276	83.74	50.41	60.20	24.41	1.00	0.93		3
281	71.71	47.44	66.16	26.67	1.00			3
286	69.60	46.97	67.49	27.08				3
290	71.02	46.74	65.81	26.58				3
correlation =	0.806							

facies	SSmean_avg (μm)	SSmean_min (μm)	SSmean_max (μm)	SSmean_median (μm)
F1	NA	NA	NA	NA
F2	26.7	24.7	32.3	26.6
F3	24.1	21.4	27.1	24.0

Appendix H

Appendix H3. Sortable silt calculations for core CE21015_05_GC. SS% = percentage sortable silt; SSmean = mean grain size of sortable silt fraction; rrunx = downcore running correlation value between SS% and SSmean for window length x.

sample depth (cm)	% < 63 µm	% 10-63 µm	SS% (µm)	SSmean (µm)	rrun5	rrun7	rrun9	facies
4	86.74	49.38	56.93	22.78				3
9	87.10	52.02	59.72	23.77				3
14	89.75	55.32	61.64	24.74	0.96			3
19	89.17	55.20	61.90	24.79	0.87	0.96		3
24	89.41	57.10	63.86	24.89	0.83	0.78	0.94	3
29	89.53	56.65	63.27	24.79	-0.11	0.53	0.79	3
34	89.40	57.04	63.81	24.97	0.16	0.34	0.49	3
40	89.29	55.37	62.01	25.09	0.24	0.26	0.37	3
45	88.16	56.69	64.30	25.31	0.44	0.31	0.57	3
49	89.62	56.94	63.53	25.64	0.42	0.63	0.51	3
57	88.99	55.57	62.44	24.85	0.72	0.59	0.51	3
69	88.75	56.95	64.17	25.05	0.60	0.50	0.55	3
80	85.79	55.84	65.08	26.06	0.59	0.56	0.95	3
89	82.87	52.32	63.14	25.74	0.44	0.96	0.93	3
96	85.80	54.86	63.94	26.07	1.00	0.94	0.93	3
104	89.20	55.12	61.79	25.31	0.99	0.96	0.94	3
114	95.61	50.11	52.42	22.09	0.96	0.97	0.96	3
118	63.40	41.92	66.12	27.59	0.96	0.96	0.94	3
125	93.52	56.05	59.94	23.67	0.95	0.93	0.94	3
130	93.83	53.71	57.24	23.28	0.95	0.93	0.93	3
149	93.78	55.90	59.60	23.74	0.95	0.95	0.92	3
154	92.55	57.21	61.81	24.35	0.96	0.91	0.95	3
160	92.71	56.37	60.80	24.23	0.86	0.88	0.85	3
165	92.98	55.57	59.77	24.09	0.88	0.87	0.86	3
174	93.82	55.89	59.58	23.65	0.86	0.88	0.87	3
183	94.14	55.52	58.98	23.23	0.76	0.85	0.86	3
187	93.03	55.74	59.92	23.64	0.95	0.81	0.65	3
197	89.34	53.20	59.55	23.59	0.95	0.59	0.87	3
201	89.84	54.03	60.15	23.80	0.08	0.88	0.98	3
209	88.97	53.64	60.29	24.11	0.85	0.98	0.94	3
215	87.69	52.26	59.60	24.33	0.98	0.95	0.93	3
219	83.17	51.28	61.65	25.25	0.97	0.95	0.95	3
229	57.31	40.27	70.26	28.51	0.99	0.98	0.93	2
240	78.55	47.36	60.30	25.41	0.99	0.97	0.95	2
245	88.47	42.04	47.52	21.46	0.99	0.98	0.96	3
250	83.05	41.92	50.48	21.85	0.98	0.98	0.97	3
260	65.93	37.47	56.84	25.20	0.99	0.98	0.98	2
263	63.91	38.10	59.62	25.75	0.99	0.99	0.99	2
265	59.43	37.27	62.72	27.22	0.99	0.99	0.99	2
267	66.43	40.69	61.24	26.66	0.99	0.99	0.98	2
268	52.19	34.14	65.42	28.11	0.99	0.96	0.96	2
269	48.61	32.18	66.19	28.15	0.96	0.97	0.97	2

Appendix H

270	49.28	33.42	67.81	29.14	1.00	0.98	0.97	2
273	77.01	46.81	60.78	25.31	1.00	0.99	0.98	2
275	88.26	50.76	57.52	24.11	1.00	0.99	0.98	2
279	70.03	44.50	63.55	26.79	0.99	0.98	0.99	2
283	74.54	45.84	61.50	26.00	0.97	0.98	0.98	2
287	68.89	43.75	63.50	26.66	0.89	0.97	0.97	2
290	76.75	47.78	62.25	25.63	0.91	0.93	0.98	2
293	50.36	32.48	64.49	27.06	0.98	0.95	0.96	2
297	53.80	34.61	64.33	27.30	0.94	0.96	0.96	2
300	49.46	32.20	65.10	27.92	0.91	0.95	0.95	2
302	53.48	36.08	67.46	28.35	0.92	0.95	0.97	2
305	45.33	30.33	66.91	28.54	0.96	0.99	0.99	2
307	53.94	35.26	65.36	27.93	0.99	0.99	0.99	2
309	66.61	41.66	62.54	26.40	1.00	0.99	0.98	2
312	77.92	46.45	59.61	24.96	1.00	0.98	0.98	3
316	87.62	51.04	58.25	24.12	0.96	0.97	0.97	3
320	83.04	48.93	58.92	24.58	0.99	0.95	0.95	3
325	70.21	45.13	64.27	26.32	0.97	0.92	0.94	3
329	79.24	47.65	60.13	24.94	0.91	0.94	0.94	3
333	78.85	46.09	58.45	24.75	0.89	0.92	0.94	3
340	61.32	38.66	63.04	27.01	0.96	0.91	0.93	2
343	64.39	40.73	63.26	26.55	0.98	0.97	0.94	2
346	65.37	41.34	63.24	26.86	0.98	0.97	0.96	2
349.5	75.89	45.04	59.35	25.20	0.98	0.96	0.95	2
352	78.07	45.77	58.62	24.75	0.98	0.95	0.95	2
356	62.94	40.85	64.90	27.01	0.95	0.96	0.94	2
359	64.69	40.81	63.08	26.45	0.97	0.93	0.81	2
363	82.52	49.73	60.27	24.69	0.98	0.82	0.81	3
367	83.66	48.77	58.30	24.18	0.71	0.87	0.86	3
371	85.15	51.32	60.27	24.50	0.63	0.83	0.93	3
375	82.51	51.60	62.54	24.49	0.78	0.92	0.93	3
378	82.72	48.81	59.00	24.24	0.91	0.92	0.94	3
381	68.63	43.76	63.76	27.18	0.90	0.93	0.94	2
385	64.82	43.73	67.46	28.21	0.98	0.91	0.93	2
390	71.53	46.28	64.69	27.00	0.86	0.97	0.91	2
395	65.73	42.98	65.39	27.40	0.93	0.78	0.95	2
400	68.04	45.42	66.76	27.51	0.81	0.82	0.87	2
405	62.22	40.39	64.92	27.07	0.77	0.88	0.89	2
410	49.37	33.05	66.95	28.90	0.87	0.86	0.86	2
415	57.42	37.53	65.37	27.39	0.92	0.85		2
420	65.06	40.68	62.52	26.20	0.92			2
425	51.62	34.26	66.38	27.61				2
429	57.36	37.50	65.37	27.76				2
correlation =	0.880							

facies	SSmean_avg (μm)	SSmean_min (μm)	SSmean_max (μm)	SSmean_median (μm)
F1	NA	NA	NA	NA
F2	27.0	24.1	29.1	27.0
F3	24.4	21.5	27.6	24.5

Appendix H

Appendix H4. Sortable silt calculations for core B2306-GC03. SS% = percentage sortable silt; SSmean = mean grain size of sortable silt fraction; rrunx = downcore running correlation value between SS% and SSmean for window length x.

sample depth (cm)	% < 63 µm	% 10-63 µm	SS% (µm)	SSmean (µm)	rrun5	rrun7	rrun9	facies
6	79.96	55.18	69.01	26.57				3
8	82.31	55.43	67.34	25.72				3
15	82.92	56.28	67.87	25.79	0.98			3
19	84.55	55.83	66.03	25.17	0.98	0.96		3
24	81.90	56.20	68.62	26.16	0.96	0.94	0.95	3
29	56.97	39.39	69.14	26.57	0.94	0.94	0.94	3
33	85.04	55.34	65.07	25.16	0.94	0.95	0.92	3
38	84.85	57.16	67.37	25.46	0.93	0.91	0.92	3
43	83.50	55.07	65.96	25.19	0.87	0.91	0.90	3
48	82.55	55.87	67.68	25.99	0.89	0.86	0.97	3
53	81.32	55.40	68.12	26.47	0.93	0.97	0.98	3
58	82.91	56.18	67.77	26.23	0.99	0.99	0.98	3
61	86.17	58.23	67.58	25.71	0.99	0.99	0.99	3
64	91.90	56.91	61.93	23.50	0.99	0.99	0.99	3
69	93.05	58.07	62.40	23.65	1.00	0.99	0.99	3
73	92.06	57.64	62.61	23.76	0.97	1.00	0.99	3
77	92.30	56.75	61.49	23.23	0.97	0.97	1.00	3
82	91.65	58.34	63.66	23.98	0.97	0.98	0.98	3
87	91.63	57.47	62.71	23.77	0.99	0.98	0.98	3
92	91.66	57.87	63.14	23.85	0.95	0.98	0.98	3
96	91.41	58.30	63.78	24.17	0.99	0.98	0.99	3
101	92.55	58.01	62.68	23.74	0.99	0.99	0.97	3
104	92.83	57.63	62.09	23.54	1.00	0.97	0.95	3
108	93.39	57.83	61.92	23.45	0.95	0.95	0.91	3
113	91.63	57.41	62.65	23.70	0.88	0.77	0.88	3
118	92.23	58.14	63.03	23.74	0.75	0.76	0.70	3
123	91.91	57.09	62.12	23.64	0.75	0.66	0.88	3
127	90.97	57.14	62.81	23.60	0.59	0.88	0.90	3
132	90.13	55.54	61.62	23.54	0.82	0.89	0.90	3
137	91.87	56.92	61.95	23.26	0.86	0.85	0.88	3
142	92.20	55.57	60.27	23.02	0.78	0.84	0.82	3
147	90.37	54.78	60.62	23.17	0.83	0.76	0.88	3
152	90.32	55.70	61.66	23.43	0.90	0.90	0.90	3
157	89.11	55.10	61.83	23.55	0.94	0.94	0.93	3
162	90.85	56.18	61.84	23.31	0.93	0.94	0.95	3
167	88.58	55.92	63.13	23.93	0.95	0.96	0.99	3
172	84.57	53.38	63.12	24.23	0.97	1.00	1.00	3
176	84.31	53.55	63.52	24.42	1.00	1.00	1.00	3
181	80.95	51.96	64.19	25.11	1.00	1.00	1.00	3
185	59.71	42.73	71.55	29.73	1.00	1.00	1.00	3
190	83.60	52.76	63.11	24.48	1.00	1.00	0.99	3
195	79.17	51.94	65.60	25.72	1.00	0.98	0.98	3

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199	79.89	52.39	65.59	25.78	0.99	0.98	0.98	3
201	37.50	27.78	74.09	30.86	0.98	0.99	0.98	2
203	62.54	44.90	71.80	28.44	0.97	0.98	0.98	2
207	65.71	47.97	72.99	29.13	0.94	0.96	0.98	2
209	54.12	40.84	75.46	30.91	0.91	0.94	0.96	2
213	67.31	48.09	71.44	28.61	0.95	0.97	0.98	2
214.5	55.12	40.67	73.78	30.86	0.98	0.99	0.99	2
217	71.85	48.45	67.43	26.95	0.99	0.99	0.99	2
220	82.19	49.05	59.68	24.47	0.99	0.99	0.99	3
224	87.48	45.34	51.83	21.64	1.00	0.99	0.99	3
229	88.84	46.00	51.78	21.46	0.99	0.99	0.99	3
234	82.93	40.80	49.20	21.02	0.75	0.94	0.98	3
239	80.45	41.13	51.12	21.77	0.94	0.94	0.94	3
244	72.42	37.70	52.06	22.47	0.99	0.95	0.95	3
249	69.84	38.16	54.64	23.66	0.98	0.98	0.98	3
254	71.27	39.35	55.21	23.53	0.93	0.99	0.99	3
259	76.72	41.05	53.50	23.05	0.99	1.00	1.00	3
264	69.42	37.85	54.53	23.77	1.00	1.00	1.00	3
268	56.34	36.61	64.99	27.06	1.00	1.00	1.00	2
270	54.58	36.67	67.18	27.69	1.00	1.00	1.00	2
273	41.82	29.93	71.56	29.20	0.97	0.99	0.98	2
276	64.51	43.97	68.15	27.65	0.93	0.94	0.96	2
279	57.31	38.90	67.88	27.73	0.94	0.96	0.97	2
282	49.08	33.53	68.31	28.34	0.99	0.98	0.98	2
284	35.87	26.01	72.52	30.77	1.00	0.99	0.99	2
286	67.06	43.91	65.48	26.59	1.00	1.00	0.95	2
288	75.09	47.00	62.60	24.75	1.00	0.95		3
293	75.10	47.56	63.33	25.17	0.80			3
298	74.54	47.48	63.70	25.47				3
302	81.08	54.90	67.71	26.11				3
correlation =	0.889							

facies	SSmean_avg (μm)	SSmean_min (μm)	SSmean_max (μm)	SSmean_median (μm)
F1	NA	NA	NA	NA
F2	28.7	26.6	30.9	28.4
F3	24.3	21.0	29.7	23.9

Appendix H

Appendix H5. Sortable silt calculations for core CE21015_15_GC. SS% = percentage sortable silt; SSmean = mean grain size of sortable silt fraction; rrunx = downcore running correlation value between SS% and SSmean for window length x.

sample depth	% < 63 µm	% 10-63 µm	SS% (µm)	SSmean (µm)	rrun5	rrun7	rrun9	facies
4	50.72	39.53	77.93	36.02				2
9	51.62	39.92	77.33	35.31				2
14	53.79	41.47	77.10	34.94	0.89			2
19	51.30	41.06	80.05	36.50	0.99	0.95		2
24	53.14	41.62	78.32	35.54	0.99	0.98	0.92	2
34	50.75	40.92	80.63	36.83	1.00	0.96	0.91	2
39	50.18	40.28	80.27	36.89	0.97	0.94	0.96	2
49	67.76	50.51	74.54	32.37	0.95	0.96	0.96	2
59	61.59	48.19	78.25	34.38	0.96	0.96	0.95	2
64	59.29	47.27	79.73	35.14	0.99	0.95	0.96	2
69	67.75	47.09	69.51	30.75	1.00	0.99	0.97	2
74	60.74	46.65	76.80	34.07	0.99	0.99	0.99	2
79	60.55	46.81	77.31	33.86	1.00	0.99	0.98	2
84	55.26	45.11	81.64	36.35	0.99	0.98	0.98	2
94	55.20	45.50	82.41	36.69	0.99	0.99	0.98	2
99	61.38	47.79	77.85	34.14	1.00	1.00	0.97	2
104	61.15	47.78	78.14	33.76	1.00	0.98	0.97	2
113	69.55	48.77	70.12	29.67	0.94	0.96	0.96	2
118	59.30	44.32	74.74	32.25	0.87	0.90	0.94	2
132	58.99	42.41	71.90	32.11	0.82	0.88	0.89	2
139	55.60	40.78	73.36	32.79	0.84	0.89	0.97	2
149	56.07	41.07	73.25	32.41	0.88	0.98	0.98	2
154	47.37	36.36	76.76	34.64	0.99	0.99	0.99	2
159	56.78	43.52	76.64	33.29	0.99	0.99	0.99	2
169	66.84	38.37	57.41	25.20	0.99	0.99	0.99	3
174	71.66	40.45	56.45	25.11	0.99	0.99	0.99	3
179	67.63	40.40	59.74	25.81	1.00	0.99	0.99	3
185	46.50	32.87	70.68	31.65	1.00	1.00	0.99	2
189	43.74	32.43	74.16	33.48	1.00	1.00	1.00	2
194	42.25	32.73	77.46	34.62	0.98	1.00	1.00	2
199	47.09	34.77	73.83	33.54	0.98	0.97	1.00	2
204	46.18	34.05	73.73	33.34	0.92	0.95	0.99	2
209	44.67	33.58	75.18	34.04	0.99	0.99	0.99	2
219	44.63	34.26	76.76	35.12	1.00	1.00	1.00	3
230	54.96	39.85	72.51	32.52	1.00	1.00	0.98	3
239	84.19	53.04	63.00	25.73	1.00	0.98	0.98	3
254	86.60	52.59	60.72	24.90	0.95	0.97	0.97	3
259	89.19	52.06	58.37	23.80	0.97	0.95	0.97	3
264	87.34	45.38	51.96	22.81	0.98	0.99	0.95	1
269	87.30	46.13	52.84	22.95	0.98	0.99	0.99	1
279	88.65	43.68	49.27	22.08	1.00	0.99	0.99	1
289	89.86	42.49	47.29	21.49	1.00	0.98	0.98	1

Appendix H

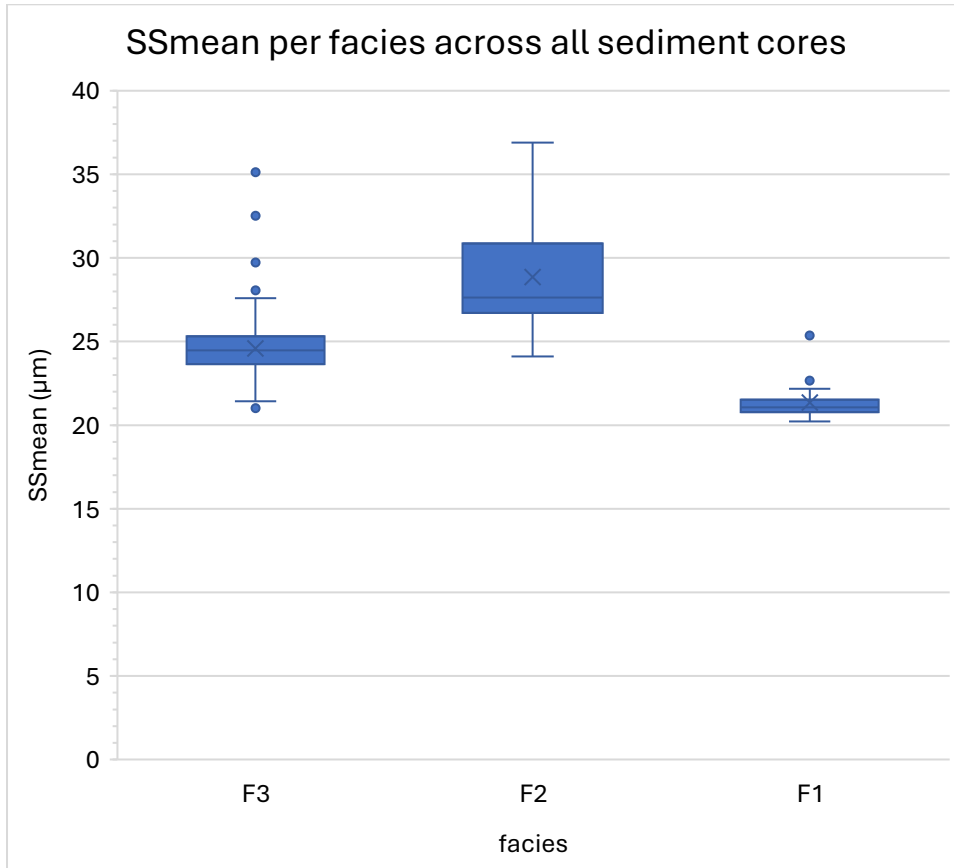
299	89.47	41.06	45.89	21.11	0.96	0.98	0.99	1
309	90.54	41.32	45.63	20.98	0.96	0.98	0.98	1
319	91.86	41.32	44.98	20.36	0.94	0.96	0.97	1
329	92.53	40.75	44.04	20.22	0.91	0.90	0.99	1
334	93.29	41.65	44.65	20.39	0.75	1.00	0.99	1
339	90.74	41.26	45.48	20.74	1.00	1.00	0.99	1
343	90.63	40.47	44.65	20.65	1.00	0.99	0.98	1
349_A	83.33	51.23	61.47	25.36	0.99	0.99	0.99	1
349_B	91.33	41.38	45.31	20.88	1.00	0.99	0.99	1
355	85.74	38.38	44.77	21.16	1.00	0.99	0.99	1
360	90.03	39.67	44.07	20.78	0.53	0.99	0.99	1
364	90.97	39.74	43.69	20.78	0.82	0.57	0.99	1
369	89.90	39.76	44.23	20.87	0.87	0.86	0.27	1
374	90.86	40.27	44.32	21.17	0.86	0.54	-0.60	1
379	88.24	39.89	45.21	21.29	0.20	-0.66	-0.07	1
384	89.81	39.91	44.43	21.16	-0.91	-0.16	-0.08	1
389	88.65	39.01	44.01	21.37	-0.14	-0.07	-0.06	1
395	86.83	36.52	42.06	22.18	0.18	0.04		1
403	85.48	38.32	44.82	23.10	0.55			1
410	87.24	36.69	42.05	21.55				1
415	81.01	35.79	44.18	22.66				1
correlation =	0.985							

facies	SSmean_avg (μm)	SSmean_min (μm)	SSmean_max (μm)	SSmean_median (μm)
F1	21.5	20.2	25.4	21.2
F2	34.0	29.7	36.9	34.1
F3	27.3	23.8	35.1	25.5

Appendix H

Appendix H6. Average, minimum, maximum, and median $\overline{SS}$ values for the three identified facies over all five sediment cores.

facies	SSmean_avg (μm)	SSmean_min (μm)	SSmean_max (μm)	SSmean_median (μm)
F1	21.4	20.2	25.4	21.1
F2	28.9	24.1	36.9	27.5
F3	24.6	21.0	35.1	24.5



Appendix H7. Flow speed differences (ΔU) between three different lithofacies calculated from the median sortable silt value according to the calibration of McCave and Andrews (2019), i.e., $1.33 \text{ cm s}^{-1}/\mu\text{m}$.

facies	deltaU_F1 (m/s)	deltaU_F2 (m/s)	deltaU_F3 (m/s)
F1	0.0	-8.5	-4.6
F2	8.5	0.0	3.9
F3	4.6	-3.9	0.0

