

Reconstructing uplift signals from sediment supply and tectonostratigraphy in a clastic shelf, Early Paleogene Shetland Platform

Lucas Albanese Valore

Thesis for the degree of Philosophiae Doctor (PhD)
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Scientific environment

This thesis for the degree of *philosophiae doctor* has been submitted to the Department of Earth Science at the University of Bergen, Norway. This research is part of a project entitled “S2S Future: signal propagation from source to sink for the future of earth resources and energy”, a Marie Skłodowska-Curie Innovative Training Network (MSCA ITN) funded by the European Union’s Horizon 2020 research and innovation programme, grant agreement No. 860383.

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This project involved several different partners and data providers. Access to seismic data at the University was provided by PGS ASA and CGG AS following an academic agreement with each company. TGS ASA provided access to the FMB database, which includes interpreted wells and maps of Northwest European geology. Access to seismic data owned by TGS was possible at VBER AS, who also supported this PhD project by hosting me for a research stay and by also providing access to interpreted seismic data. Equinor ASA also supported the project by hosting me for a month-long secondment at their office in Oslo.



Acknowledgements

Some say that by the end of your PhD journey, it feels like you know less than when you began. I would say that there is truth to this, as I am now more aware of my own vast ignorance. Yet, if I did learn anything, it was certainly because of the amazing people I had the luck to be surrounded by.

Chris, you are probably tired of hearing endless choirs of praise from your students, but I cannot refrain from adding my voice. Thank you for teaching me so much about life in and out of science. To anyone reading this thesis: the only reason this is not 100 pages longer is that Christian managed to transform me from an exceedingly wordy writer to a somewhat-less-wordy writer over the course of the last three years. Because he cannot perform miracles, of course, this is still not exactly brief (but we did our best). In this same vein, then: Christian, thank you for being the most pragmatic, reasonable, enthusiastic and accessible supervisor I could possibly ask for, and for continuously making this as much fun as it could be.

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Muito obrigado!

A handwritten signature in blue ink, appearing to read 'Lucas Alkmim', with a stylized flourish at the end.

Bergen, March 2024

Abstract

Sedimentary basins are vital archives of the planet's geodynamics and climate. The complex nature of this record implies that the extraction of tectonic and climatic fluctuations is a non-trivial endeavor, requiring one to consider a broad range of sedimentary phenomena capable of modulating these external input signals. Source-to-sink sedimentology aims to expand our understanding of these phenomena and consequently improve the analysis of original input signals altogether.

This thesis is an investigation of the external forcing parameters operating in the Shetland Platform and its adjacent basins during the Early Paleogene, when the area was subject to uplift and rapid climate change, resulting in exhumation and increased sedimentation. These processes have disputed causes, although they have been mostly linked to the emplacement of the North Atlantic Igneous Province, the Iceland Plume, and its associated fluctuations in dynamic topography (mantle-related topography).

In this thesis, the sedimentary record around the Shetland Platform is investigated using an extensive subsurface database with >200 000 km² of reflection seismic data and >1700 wells, leading to unparalleled constraints on the regional paleogeographic evolution and the characteristics of quantitative sedimentary signals preserved in the record. These results are organized in 3 papers. **Paper 1** focuses on the high-resolution stratigraphy and paleogeography of the East Shetland Platform. Seismic stratigraphic principles, seismic attribute analysis and seismic geomorphology are applied to assess process regimes, sediment transport directions, and the laterally varied stratal stacking patterns of the Paleocene – Early Eocene succession.

This stratigraphic framework is then expanded for a comparison with the Faroe-Shetland Basin and for an assessment of the regional sediment mass-budget in **Paper 2**. This paper documents distinct peaks in sediment supply during the Paleocene – Early Eocene, which are then used to investigate the nature of external input signals by applying a statistical-empirical model of river catchment erosion rates.

In addition to sediment supply signals, the stratigraphic record of the Shetland Platform can also be used to extract relative sea-level histories and, after removal of compaction

and eustasy: uplift. In **Paper 3**, clinoform trajectory analysis and backstripping are used to isolate the tectonic/dynamic topography component of vertical motions in the East Shetland Platform. When inserted into a broader paleogeographic context, these vertical motions can inform models of geodynamic evolution.

Results from this thesis can be used to calibrate models of dynamic, tectonic and climatic forcing on North Atlantic sediment routing systems and ultimately inform source-to-sink concepts in general. They can also be expanded in further case studies or used to improve models of along-strike sequence stratigraphy. The conclusions obtained here should be fruitful for further forward modelling and for continued discussions on North Atlantic geodynamics.

Sammendrag

Sedimentære bassenger er viktige arkiver for planetens geodynamikk og klima. Kompleksiteten til dette registeret innebærer dog at det er en ikke-triviell oppgave å ekstrahere tektoniske og klimatiske svingninger, noe som krever at man vurderer et bredt spekter av sedimentære fenomener som kan modulere disse eksterne signalene. *Source-to-sink*-sedimentologi har som mål å utvide vår forståelse av disse fenomenene og dermed forbedre analysen av de opprinnelige inngangssignalene.

Denne avhandlingen er en undersøkelse av de eksterne pådrivningsparameterne som opererte på Shetland-plattformen og dens tilstøtende bassenger under tidlig paleogen, når området var utsatt for landheving og rask klimaendring, noe som resulterte i eksponering og økt sedimentasjon. Disse prosessene har omdiskuterte årsaker, selv om de hovedsakelig har blitt knyttet til plasseringen av den nordatlantiske vulkanske provinsen, Island-plumen og dens assosierte svingninger i dynamisk topografi (mantelrelatert topografi).

Det sedimentære registeret rundt Shetland-plattformen har blitt undersøkt ved hjelp av en omfattende undergrunnsdatabase med seismiske data og brønninformasjon, som kan brukes til å trekke slutninger om den regionale paleogeografiske utviklingen og om de kvantitative sedimentære signalene bevart i registeret. Disse resultatene er organisert i 3 artikler. **Artikkel 1** fokuserer på høyoppløselig stratigrafi og paleogeografi på Øst-Shetland-plattformen. Artikkelen anvender seismiske stratigrafiske prinsipper, seismisk attributtanalyse og seismisk geomorfologi for å vurdere prosessregimer, retninger for sedimenttransport og lateralt varierende stratigrafiske stablingsmønstre for lagrekken fra paleocen til tidlig eocen.

Dette stratigrafiske rammeverket blir deretter utvidet for en sammenligning med Færøy-Shetland-bassenget og for en vurdering av den regionale sedimentmassens budsjett i **Artikkel 2**. Denne artikkelen dokumenterer distinkte topper i sedimenttilførsel i paleocen – tidlig eocen, som deretter brukes til å undersøke de eksterne inngangssignalene ved å anvende en statistisk-empirisk modell av erosjonsrater for elveerosjon.

I tillegg til signaler om sedimenttilførsel, kan det stratigrafiske registeret for Shetland-plattformen også brukes til å bestemme tidligere variasjoner i om relativt havnivå og til slutt landheving. I **Artikkel 3** brukes klinoforbane-analyse og «backstripping» for å isolere den tektoniske/dynamiske topografikomponenten av vertikale bevegelser i Øst-Shetland-plattformen. Når den settes inn i en bredere paleogeografisk sammenheng, kan disse vertikale bevegelsene informere modeller av geodynamisk utvikling.

Resultater fra denne avhandlingen kan brukes til å kalibrere modeller av dynamiske, tektoniske og klimatiske drivende faktorer på nordatlantisk sedimentdistribusjonssystemer og *source-to-sink*-sedimentologi generelt. De kan også utvides i ytterligere case-studier eller brukes til å informere konseptuelle modeller av langs-strøk sekvensstratigrafi. Konklusjonene oppnådd her bør være fruktbare for videre fremadrettet modellering og for fortsatte diskusjoner om nordatlantisk geodynamikk.

List of Publications

Publications included in this thesis:

Paper 1: Valore, L., Sømme, T., Patruno, S., Robin, C., Guillocheau, F., Eide, C. H.. Paleogeography and 3D variability of a dynamically uplifted shelf: observations from seismic stratigraphy of the Paleocene East Shetland Platform.

(accepted pending revisions in *Basin Research*)

Paper 2: Valore, L., Sømme, T., Eide, C. H. High-resolution sediment mass-budget of the Shetland Platform during the Paleocene - Early Eocene – constraints on external forcing by uplift and climate.

(manuscript in review in the *Journal of the Geological Society*)

Paper 3: Valore, L., Sømme, T., Schofield, N., Eide, C. H. Short-wavelength uplift variations in the Paleocene East Shetland Platform reveal lithosphere imprint on plume-related support.

(manuscript prepared for submission to *Geology*)

Appendix A: Sømme, T. O., Huwe, S. I., Martinsen, O. J., Sandbakken, P. T., Skogseid, J., Valore, L. A. 2023. Stratigraphic expression of the Paleocene-Eocene Thermal Maximum climate event during long-lived transient uplift—An example from a shallow to deep-marine clastic system in the Norwegian Sea. *Frontiers in Earth Science*, 11, 1082203.

The article in Appendix A was published open access and is reproduced in this thesis under the terms of the Creative Commons Attribution License (CC BY). All other articles will be open access once published.

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1. Introduction

The source-to-sink approach

Source-to-sink sedimentology is a field of study concerned with establishing quantitative constraints for the consecutive generation, transfer and deposition of sediments in a given **sediment routing system**. Its primary objective is to determine how these processes relate to the system's **boundary conditions** - the physical parameters that govern its dynamics (Figure 1.1). This is possible because sediment generation and deposition respond to and record **signals** associated with these boundary conditions and their fluctuations (often termed **forcing parameters**). In turn, achieving a holistic understanding of signals and their forcing parameters is important because it enables first-order predictions and extrapolations for specific case studies, something that is relevant for all applications of sedimentary geology (Castelltort et al., 2023). This is akin to the philosophy behind sequence stratigraphy, upon which source-to-sink sedimentology aims to build.

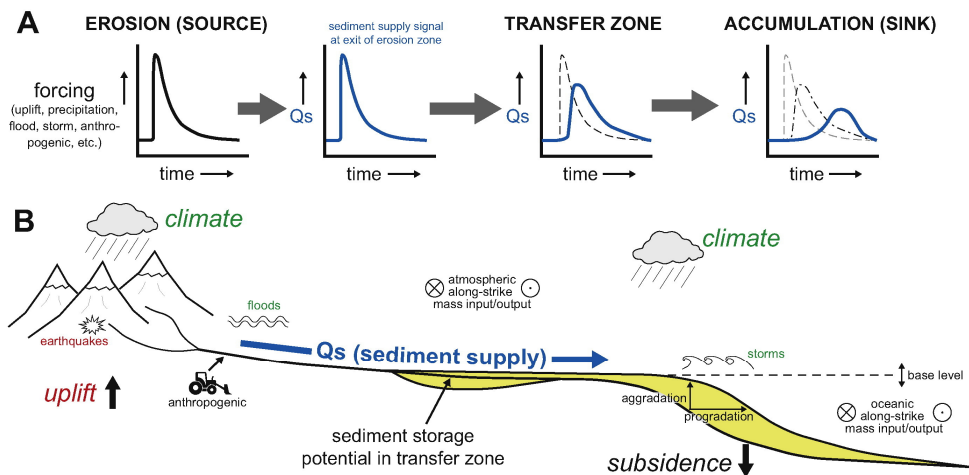


Figure 1.1 – Generalized concepts of source-to-sink sedimentology from Romans et al. (2016). A) The diagrams show a simplified representation of a signal generated by a change in boundary conditions, which could be a number of different processes shown in B. This signal is observed as change in sediment supply (Q_s), which is then transferred and recorded in the sink after a time delay and loss in initial amplitude. B) Main components of the sediment routing system.

Typical examples of forcing parameters are tectonic uplift and climate change, which are able to modify the intensity of erosion in the sediment source-area and the magnitude of sediment deposition in the basin, resulting in what can be termed an **external** or **allogenic** signal. For example, an increase in uplift rates or precipitation results in a traceable increase in sediment fluxes and sediment deposition rates across the basin – this is often called a Q_s (sediment flux) signal, a typical measurement in source-to-sink studies (Allen et al., 2013; Yuan et al., 2019). At the same time, the intrinsic variability and dynamics of sediment transport also result in **internal** or **autogenic** signals, which are fluctuations in the sedimentary record that do not require any external forcing. A common example of an autogenic change would be the avulsion of a delta lobe due to its natural compensational stacking (Figure 1.2). As observed in outcrops or wells, the stratigraphic result of lobe avulsion might be a change in facies stacking patterns (*e.g.* retrogradation and progradation), which is itself a non-unique response, since it could also be produced by changes in the sediment supply or sea-level forced by an allogenic phenomenon. Importantly, the distinction between allogenic and autogenic is often a matter of spatial and temporal scales. When these processes happen at the same scale, such differentiation often becomes more complicated (Bhattacharya et al., 2019; Burgess and Duller, 2024).

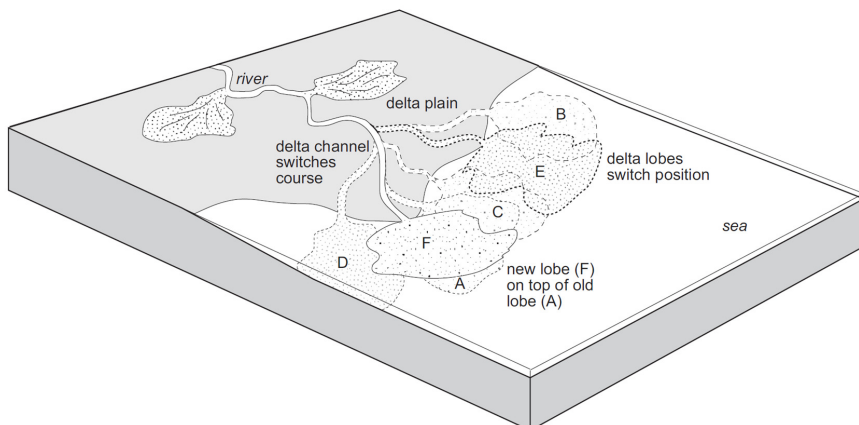


Figure 1.2 – Schematic representation of delta avulsion and overlapping lobes (Nichols, 2009). Avulsion results in a series of coarsening-up lobe successions.

While the underlying concept of source-to-sink sedimentology is certainly not novel (*e.g.* Schumm, 1977), the techniques and workflows required to obtain such quantitative constraints have become more prominent in the last decades, made available after advances in our understanding of quantitative geomorphology, landscape evolution and sediment transport dynamics (see Castelltort et al., 2023). Just as these techniques began to offer novel answers into the nature of source-to-sink sedimentology, they also opened several new lanes for further investigation of sedimentary signals, often challenging the more intuitive concepts derived from qualitative sequence stratigraphy and its usual assumptions on underlying geological controls (*e.g.* Castelltort and Van Den Driessche, 2003; Armitage et al., 2013; Burgess and Prince, 2015; Romans et al., 2016; Wolf et al., 2022).

A primary reason for this is that the dynamics of sediment transport and the consequent downstream transfer of sedimentary signals (or simply **signal propagation**) are considerably complex (Romans et al., 2016; Tofelde et al., 2021). Much like our climate, sediment routing systems are dynamic processes governed by multiple, sometimes feedbacking parameters operating at various scales. It is possible to analyze this by looking at the relationship between catchment erosion and downstream sediment flux generation. At catchment scale, the dynamics of bedrock incision can be described in a power law:

$$\frac{dz}{dt} = U - K A^m S^n$$

where the rate of topographic change $\frac{dz}{dt}$ is a result of uplift (U) minus an erosion term, here considered to be proportional to area (A , which acts as a proxy for discharge) and channel slope (S) multiplied by an erodibility constant K , dependent on lithology, vegetation and precipitation rates (Whipple and Tucker, 1999). The area (m) and slope (n) exponents are positive numbers, with m/n often assumed to be near 0.5 ($0.35 < m/n < 0.6$, Howard and Kerby, 1983; Whipple and Tucker, 1999). An important derivation of this model is that at $\frac{dz}{dt} = 0$, uplift and erosion are completely balanced. This is often described as equilibrium or **steady-state** conditions (Figure 1.3).

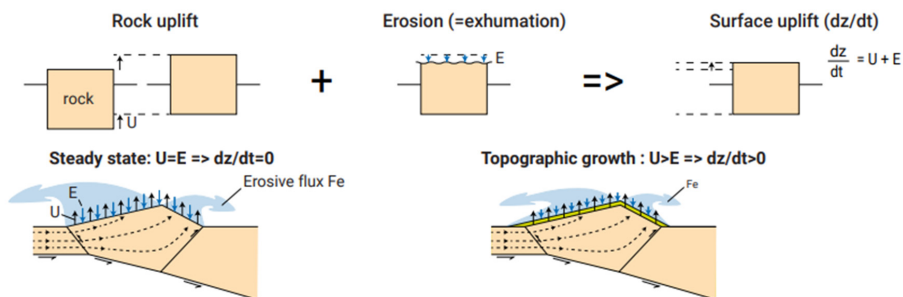


Figure 1.3 - Steady-state topography, where U (rock uplift) = E (erosion or exhumation), leading to no surface uplift, $dz/dt = 0$. From Castelltort et al. (2023).

This stream power law formulation can be used to describe the behavior of river channels when changes in boundary conditions are introduced. For instance, rainfall and discharge have a first-order control on the erodibility of a river stream. For bedrock incising rivers, when discharge increases due to a change in climate (Figure 1.4.A1), it also augments the erosive power of the stream, leading to channel degradation (Figure 1.4.B1-C1) and increasing the erosion rates and downstream sediment flux (Q_{Sout} in Figure 1.4.D1). Over time, however, degradation will gradually cause river slopes to flatten, which consequently decreases erosion rates, effectively returning Q_{Sout} to the previous equilibrium conditions. Critically, this means that **discharge-induced** signals are essentially self-limited (Figure 1.4.D1). On the other hand, uplift has a first-order control on the slopes of the stream itself (Figure 1.4.A2-B2). For conditions of fixed erodibility and discharge (e.g. fixed climate and bedrock lithology), an increase in uplift will eventually lead to higher catchment topographies and steeper slopes, which in turn will augment erosion rates (Figure 1.4.B2-D2). This situation will be maintained for as long as the uplift rates are kept the same ($t3$ in Figure 1.4.D2 represents a steady-state sediment flux for the new uplift rate).

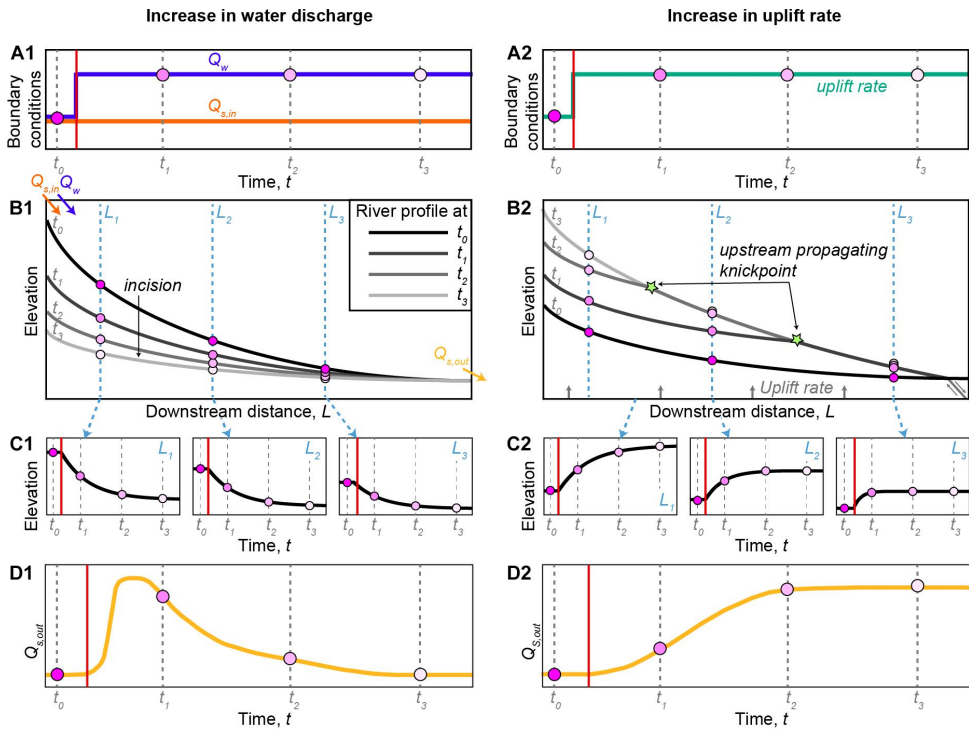


Figure 1.4 – Conceptual diagrams showing the response of river profiles and sediment flux to forcing by discharge and uplift (Tofelde et al., 2021). Colored circles represent four moments in time during landscape transience. In the left-hand side, an increase in water discharge (Q_w) with respect to upstream supply or sediment discharge ($Q_{s,in}$) is shown to cause incision (degradation) of a theoretical river profile (B1). Degradation leads to a momentary increase in $Q_{s,out}$ (t_1 in D1), which returns to the initial value after final landscape adjustment (t_3 in D1) due to a decrease in the river slopes (t_3 in B1). In the right-hand side, an increase in uplift is shown to cause steeper river slopes through the upstream migration of a knickpoint (B2). This increase in slope leads to higher $Q_{s,out}$, sustained after the landscape reaches a new steady-state (t_3 in D2).

From this results an important conclusion. It is known that erosion rates and **sediment fluxes** in a river catchment are in essence **proportional to average uplift rates** (Figures 1.3, 1.4). However, as described above, this behavior will be modulated by the erodibility efficiency (K) of river channels, which is also governed by climate. In effect, this means that the rate of landscape change (or **transience**) and the **landscape response time** to a perturbation is determined by the erodibility and transport efficiency and the size of the landscape itself, with typical response times in

the order of 10^4 to 10^6 years (e.g. Allen, 2008; Armitage et al., 2013; Tofelde et al., 2021). A simple consideration in source-to-sink studies is that the relationship between **perturbation timescales** and landscape response times has a strong bearing on the propagation of Q_s signals (Figure 1.5).

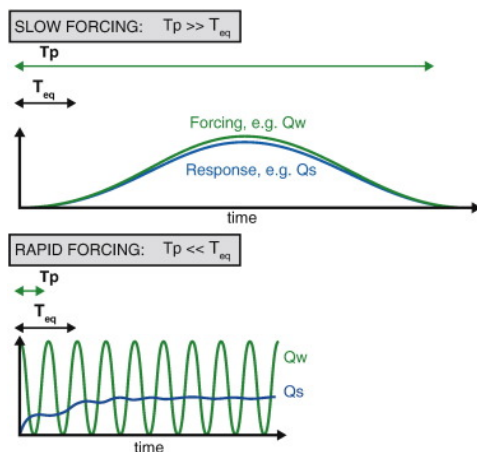


Figure 1.5 – Conceptual figure from Romans et al. (2016) showing the difference between slow and rapid forcing (T_p) with respect to landscape response or equilibrium times (T_{eq}). With $T_p \ll T_{eq}$, the landscape has no time to fully adjust to the forcing, which leads to a buffer Q_s signal.

Some of the main implications of the above observations are summarized in Figure 1.6. These are: 1) allogenic sedimentary signals are often considerably modified by autogenic processes; 2) different signals (e.g. tectonics and climate) are ultimately entangled in the same individual record, and 3) sedimentary signals are often non-unique, because they may be triggered by various forcing parameters.

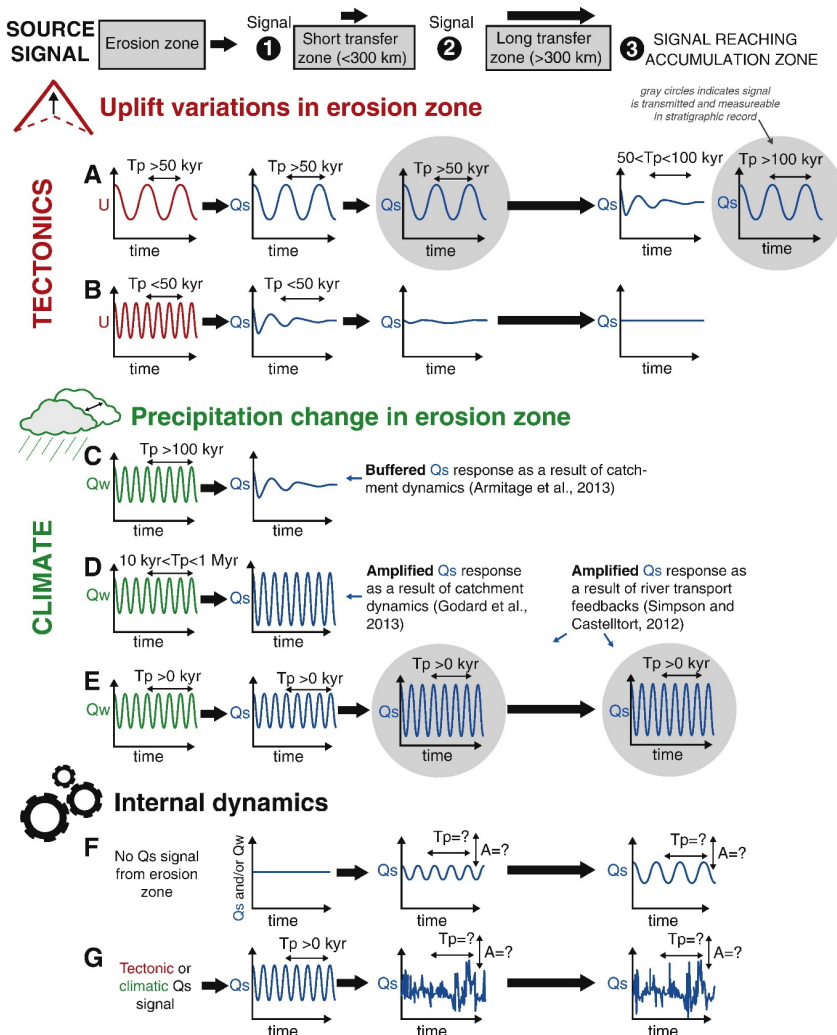


Figure 1.6 – Summary of signal propagation for perturbations of intermediate timescales (T_p , 10^2 – 10^6 yr) from Romans et al. (2016). This summary shows sediment supply (Q_s) signals induced by uplift (A, B), precipitation changes (C–E) and autogenic dynamics (F, G) and their behavior during transfer and accumulation. The gray circles indicate the cases in which the original signal has been faithfully recorded in the sink. (A) An uplift signal with a period (T_p) > 50 kyr leads to an identical Q_s signal in the erosion zone (the source-area). It may be then transmitted by a short fluvial segment and faithfully recorded in the sink or buffered by a longer transfer zone, unless the perturbation time is long enough, e.g. $T_p > 100$ kyr (Castellort and Van Den Driessche, 2003). (B) A short-period uplift signal (e.g. faulting, $T_p < 50$ kyr) is buffered in landscapes with long response times, and further buffered by autogenic processes in the transfer zone, resulting in no signal in the sink (Castellort and Van Den Driessche, 2003).

(C) A gradual, long-period water discharge (Q_w) signal ($T_p > 100$ kyr) may result in a buffered Q_s signal in the source-area due to the negative feedback between precipitation and topography and other catchment dynamics (Godard et al., 2013; Armitage et al., 2013; Tofelde et al., 2021). (D) Alternatively, an intermediate-period Q_w signal with $10 \text{ kyr} < T_p < 1 \text{ Myr}$ may result in an augmented Q_s response in the source-area due to catchment amplification, if the timing of the fluctuation matches an erodibility-dependent “optimal period” (Godard et al., 2013). (E) A Q_w signal with $T_p > 0$ kyr is transferred to the fluvial system, where it is amplified by river transport dynamics in the transfer zone (Simpson and Castellort, 2012). (F) Even without any external signals generated in the source-area, autogenic signals can be generated in the transfer zone due to the internal dynamics of fluvial sediment transport (channel meandering, delta lobe avulsion). (G) Regardless of origin, any Q_s signal can be ‘shredded’ by autogenic processes in the transfer zone if their period and amplitude match these internal dynamics.

Within this complex framework, source-to-sink studies aim to combine physical and numerical modelling with the quantification of sedimentary systems to provide data-driven ideas on the general physical nature of signal propagation. These, in turn, become useful tools for the assessment and distinction of entangled sedimentary signals in case studies throughout the planet. With this, our understanding of the preservation potential and the interaction between allogenic and autogenic signals has improved (e.g. Simpson and Castellort, 2012; Toby et al., 2019; Griffin et al., 2023 - Figure 1.6), although much work likely remains to be done (e.g. Burgess and Duller, 2024).

For this thesis, the investigation of tectonic and geodynamic signals is particularly relevant. The sedimentary record can be used to constrain changes in uplift and subsidence rates in multiple ways, and when combined these may help to navigate the uncertainty imposed by the complex nature of signal propagation. The analysis of subsidence rates is intrinsic to basin modelling studies concerned with thermal and burial histories and their applications for hydrocarbon generation (Kubala et al., 2003; Gardiner et al., 2019). Critically, deviations from predicted subsidence in many basins have been used to constrain uplift and inversion events (Nadin et al., 1997; Clift and Turner, 1998; Clark et al 2014). Uplift rates may also be derived from other observations in the stratigraphic record, such as the analysis of shorelines and unconformities (Shaw Champion et al., 2008; Stucky de Quay et al., 2017). Both are very relevant for this study and are discussed further in Chapters 2 and 3.

In this thesis, the Shetland Platform in the Early Cenozoic Northeast Atlantic is investigated to improve our understanding on the impact of superimposed geodynamic and climatic forcing on source-to-sink sedimentology. The Shetland Platform was uplifted in the Paleocene – Early Eocene, during the emplacement of the North Atlantic Igneous Province and the opening of the North Atlantic Ocean (White and Lovell, 1997; Jones and White, 2003; Stoker et al., 2018). Large Igneous Province (LIP) magmatism in the North Atlantic has been linked with the Icelandic Plume, which in turn could have generated uplift and hinterland exhumation, leading to the generation of Qs sedimentary signals and specific tectono-stratigraphic signatures (Liu and Galloway, 1997; Mudge and Jones, 2004; Smallwood, 2008). Debates surround the scales, timings and origin of uplift and exhumation (Mudge and Jones, 2004; Łuszczak et al., 2018; Stoker et al., 2018; Conway-Jones and White, 2022), the paleogeographic response to external forcing (Jones and White, 2003; Mudge, 2015) and the contribution of both long-term and sharp Paleocene – Eocene warming (*e.g.* the Paleocene – Eocene Thermal Maximum) on sedimentary signals (Sømme et al., 2019; Jin et al., 2022; Jolley et al., 2023). Importantly, this requires the consideration of entangled uplift and climatic signals (Figure 1.6). These themes are explored in three original research articles that make up the results of this thesis (Chapter 5), which will be discussed and synthesized in Chapter 6.

Objectives

The overall goal of this study was to provide data-driven observations on the influence of geodynamic processes related to LIP magmatism on source-to-sink systems, using the Paleocene-Early Eocene Shetland Platform as a case study. Because a large volume of subsurface data around the Shetland Platform is available, a main objective was to obtain a coherent and robust paleogeographic and tectono-stratigraphic overview of all basins surrounding the area, allowing both regional and detailed observations to be inserted in the same paleogeographic context. This study is a part of a larger project (S2S-Future) aimed at improving concepts, methods and training on source-to-sink sedimentology, being part of a specific work package focused on long-term step-changes in forcing factors.

The specific goals of the study were therefore to:

- Devise a detailed seismic stratigraphic framework for the East Shetland Platform based on newly acquired, high-resolution 3D data, comparing our findings with previous chrono- and biostratigraphic schemes.
- Use observations based on seismic and well data to provide new constraints on the paleogeographic evolution of the Shetland Platform.
- Use seismic stratigraphy, paleogeography and quantitative source-to-sink constraints to inform hypotheses on uplift and climate operating in the Shetland Platform during the Paleocene – Early Eocene, providing new ideas for further studies.

2. State of the art

The relationship between geodynamics and magmatism

Tectonic activity and magmatism are intrinsically linked, as geodynamic settings have a first-order control on the rates of production of magma, its sources and their consequent geochemical affinities. Plate-boundary-related processes (convergence, divergence and strike-slip motion) condition several variables involved in magma production, such as crustal and lithospheric thicknesses, geothermal gradients, strain rates and volatile enrichment (McKenzie and Bickle, 1988; Anderson, 2005). Divergence, which is more relevant for this study, enables magmatic production by favoring lithospheric thinning and adiabatic decompression of the mantle, inducing its partial melting in areas such as mid-ocean ridges (McKenzie and Bickle, 1988; White and McKenzie, 1989).

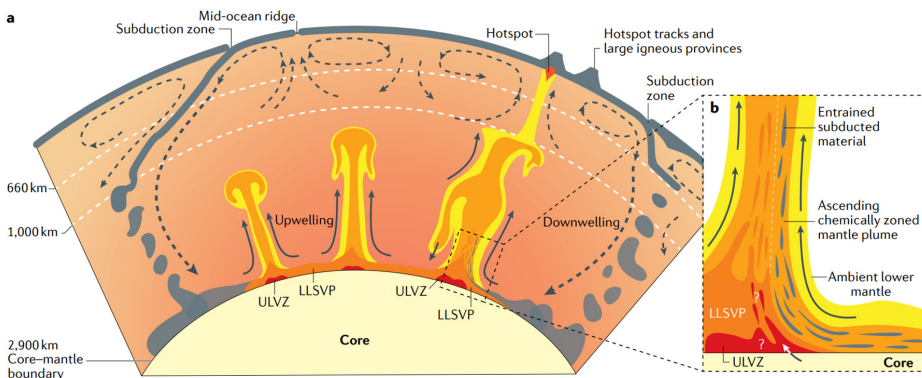


Figure 2.1 – Conceptual model of mantle convection and mantle plumes (Koppers et al., 2021). A) Cross-section of the Earth's interior, showing plume generation near large low-shear-velocity provinces (LLSVPs) near the core-mantle boundary, which may contain ultra-low velocity zones (ULVZs). Along the edges of LLSVPs, subducted material may accumulate throughout long time periods ($>100s$ Myrs). This dynamic would be connected to supercontinent cycles. B) Cross-section of a plume root, depicting entrained subducted material (gray), LLSVP and ULVZ components (including partial melt?) leading to chemical zonation of the plume.

“**Anomalous**” or **excess** magmatism that in theory cannot be explained by plate-tectonic processes alone are often attributed to the presence of **mantle plumes**, which have been linked to several large igneous provinces (Coffin and Eldhom, 1994; Dang et al., 2024). Plumes are purported bodies of rising and anomalously hot mantle that are able to increase regional temperatures above the solidus and induce great volumes of melt. Mantle plumes were initially conceived as relatively simple “mushroom”-like structures originating at the core-mantle boundary, focusing heat generated by radioactive decay (Figure 2.1; Morgan, 1971; Richards et al., 1989; Campbell and Griffiths, 1990). Such structures have been related to “hot-spots” - areas of excess intra-plate volcanism with distinct petrological characteristics, such as the abundance of picrites (high-magnesium olivine-rich basalts thought to represent high-temperature mantle melts). Hot-spots like the Hawaii archipelago have been used to interpret plume-tracks, which are trails of volcanic centres with progressive ages connected to the rate of movement of a translating plate above a stationary mantle plume (Morgan, 1971).

This view has been somewhat modified in the last decades, with the advent of improved tomographic imaging of the upper and lower mantle. Studies have demonstrated the existence of “large low-shear-velocity provinces” (LLSVPs), which are thought to be large thermochemical anomalies in the lower mantle (Garnero et al., 2016; Koppers et al., 2021) distinct from individual “simple” plumes (Figures 2.1, 2.2). Two LLSVPs are currently imaged in the Earth’s interior (Figure 2.2), and they may be rather long-lived (>300 Ma) features (Torsvik et al., 2014).

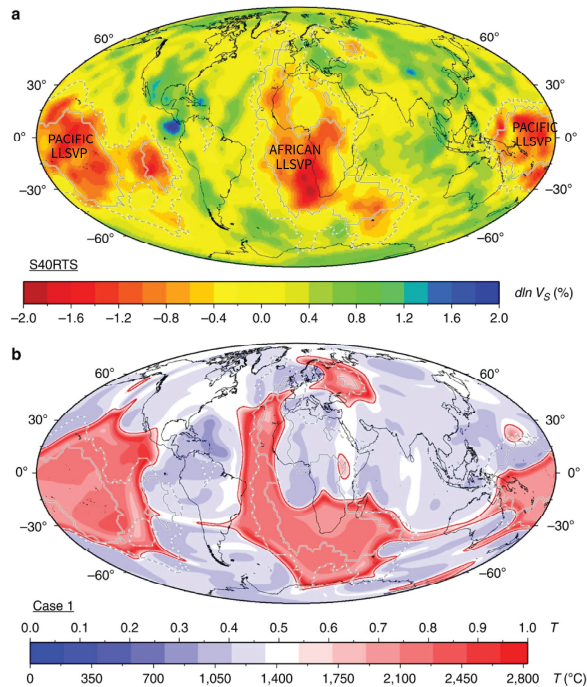


Figure 2.2 – Models of large low-shear-velocity provinces (LLSVPs), modified from Flament et al. (2017). A) Seismic velocity anomalies at 2677 km depth for tomography model S40RTS (Ritsema et al., 2011), showing the African and Pacific LLSVPs or “superswells”. B) Predicted present-day mantle temperature at 2677 km depth in one case model in Flament et al. (2017).

The existence of such bodies of upwelling mantle is central to a major debate on the nature of mantle convection and its exact ties to plate-tectonics and magmatism (Figure 2.3). Several authors hold that convection and plate-tectonics are primarily sustained by the descent of subducted oceanic lithosphere (Hager, 1984; Conrad and Lithgow-Bertelloni, 2002, 2004), meaning that direct or indirect forces related to mantle upwelling would only play a minimal (if any) role at inducing plate motion. Many authors even reject the notion that mantle plumes exist, suggesting that excess magmatism is also primarily driven by plate-tectonic processes (Anderson, 2005; Foulger et al., 2020). This view has it that the mantle only plays a passive role in supplying melt-prone material (often in the form of “geochemical reservoirs”, such as ancient subducted slabs or metasomatized lithosphere, Pilet et al., 2008), which still need to be primarily activated by plate-driven processes like extension (Figure 2.3; Anderson, 2005).

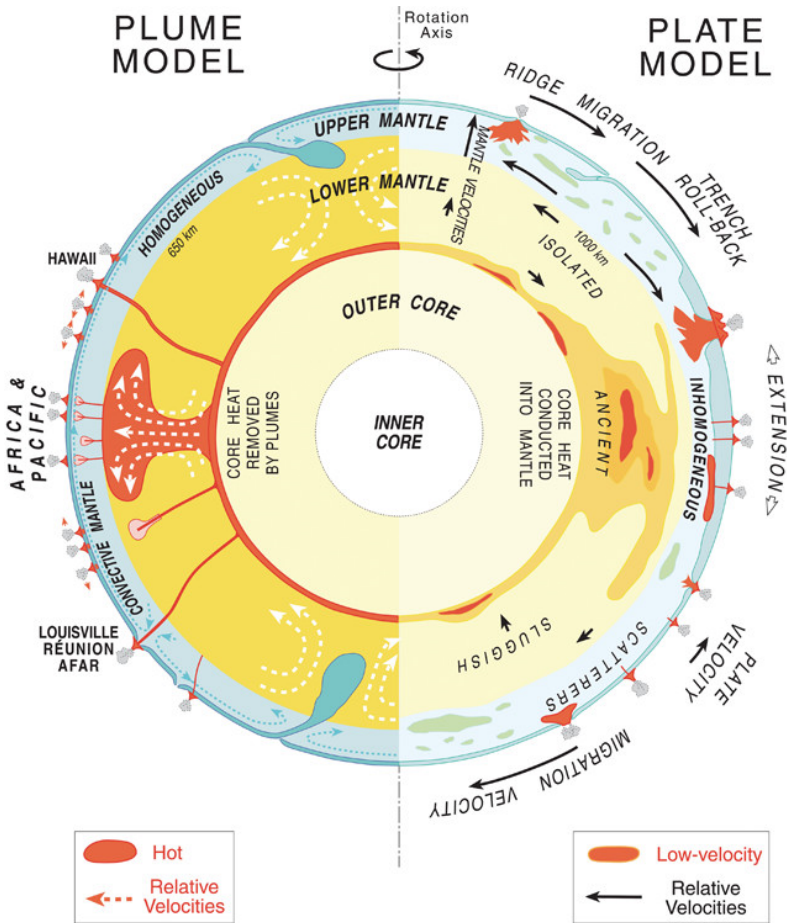


Figure 2.3 – A schematic representation of the plume (left side) and plate (right side) models of Earth's interior dynamics and magmatism from Anderson (2005). In the left side, hotspots and mantle plumes are shown. Narrow mantle plumes and giant superswells coexist and are generated by heat in the core-mantle boundary. Melting associated with plumes occurs at various levels from lower to upper mantle. Red regions are assumed to be hot and buoyant, blue regions are colder and denser than the average mantle. Other areas of the mantle are assumed to be homogeneous. In the plate model in the right side, volcanism is concentrated in tensile regions at plate-boundaries or within them. The upper mantle is heterogeneous and includes scattered portions of recycled crust and lithosphere, affecting fertility and melt generation. Melt anomalies in the upper mantle are accessed by plate stresses. The upper mantle is the main domain of mantle convection, and uplift and extension may be caused by low-density regions in the mantle. In contrast, the lower mantle is viscous, sluggish and isolated from the upper mantle. It includes ancient low velocity anomalies (which are not necessarily hotter, but may have different compositions or degrees of melt than the surrounding mantle).

On the other hand, mantle plumes and mantle upwelling in general have been recently suggested to play a larger role in mantle convection, with some authors proposing that they could also be key drivers of plate-tectonics in scenarios like the breakup of supercontinents (Zhang et al., 2018; Mitchell et al., 2021). This includes the African LLSVP, which has been related to the breakup of Pangea and could be a root for the Iceland Plume (Steinberger et al., 2015, 2019; Dang et al., 2024).

Given the open-ended nature of this debate, it is useful to approach this problem carefully. Importantly, in this study, the goal is to examine data-driven stratigraphic interpretations and their potential implications for geodynamic models, which should be relevant regardless of the chosen paradigm. This requires one to steer closer to the realm of observed phenomena, including those associated with mantle activity. One such phenomenon would be the existence of topography induced by mantle support, since its interpretation does not inherently require deciding whether mantle plumes (or which of their proposed variants) exist.

Surface effects of mantle-related topography

The topographic response to radial stresses produced by convective flow and lateral density variations in the mantle may be termed **dynamic topography** (Hoggard et al., 2021; Figure 2.4). Some authors apply this definition to deflections related to buoyancy in the lower mantle only (Hager et al., 1985), while others include the effects of shallower, sublithospheric structures (Steinberger, 2016) or even the lithosphere itself (Moucha and Forte, 2011).

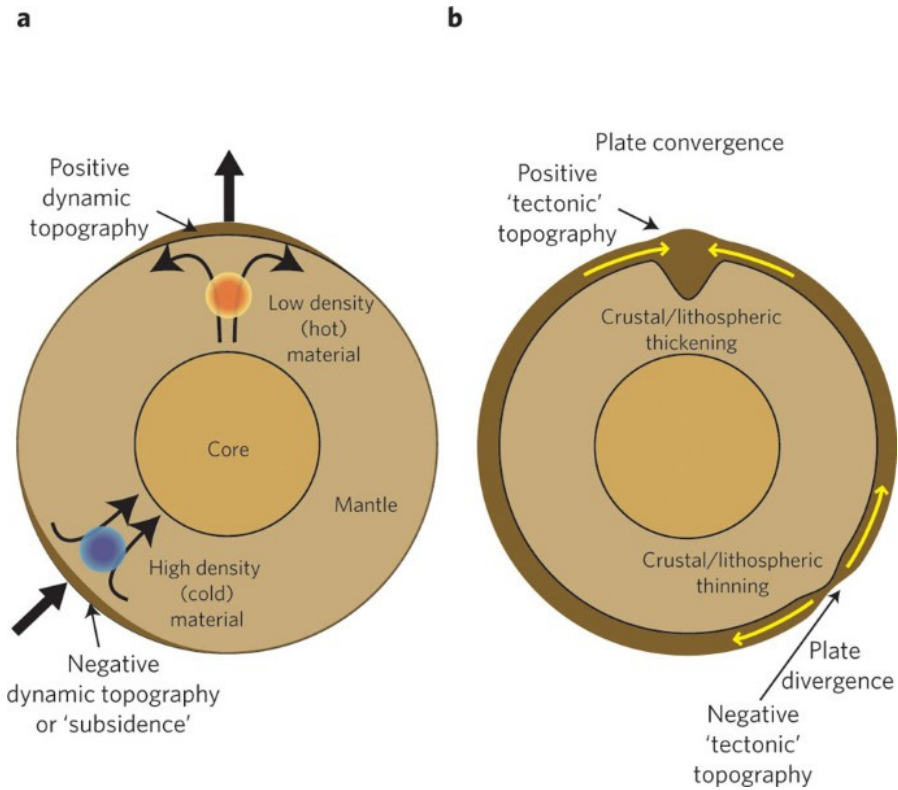


Figure 2.4 – Simplified sketch showing the effects of mantle flow on surface topography versus plate-tectonics (Braun, 2010). A) Dynamic topography can be generated by low density (red) and high density anomalies (blue) in the mantle. Black arrows represent ensuing mantle upwelling or downwelling. B) Plate-tectonic processes (convergence and divergence, yellow arrows) and isostatically compensated topography created by lithospheric thickening or thinning.

Surface dynamic topography changes through time as convection in the mantle evolves or overlying plate-tectonic configurations also change. Geodynamic processes associated with flow or density differences in the mantle produce distinct mechanisms for vertical motions in the Earth's surface (Figure 2.5). Importantly, these mechanisms can be related to different temporal and spatial scales of evolving topography. For instance, for constant rates of mantle flow below a moving tectonic plate, the rate of dynamic topographic change will depend on the speed of the translating plate itself (Figure 2.5.D). Some of the fastest rates of dynamic topography change (~ 0.5 mm/yr) have been purported to be associated with the lateral advection of thermal anomalies (Figure 2.5.F; Hartley et al., 2011) as opposed to vertical flow (Figure 2.5.E).

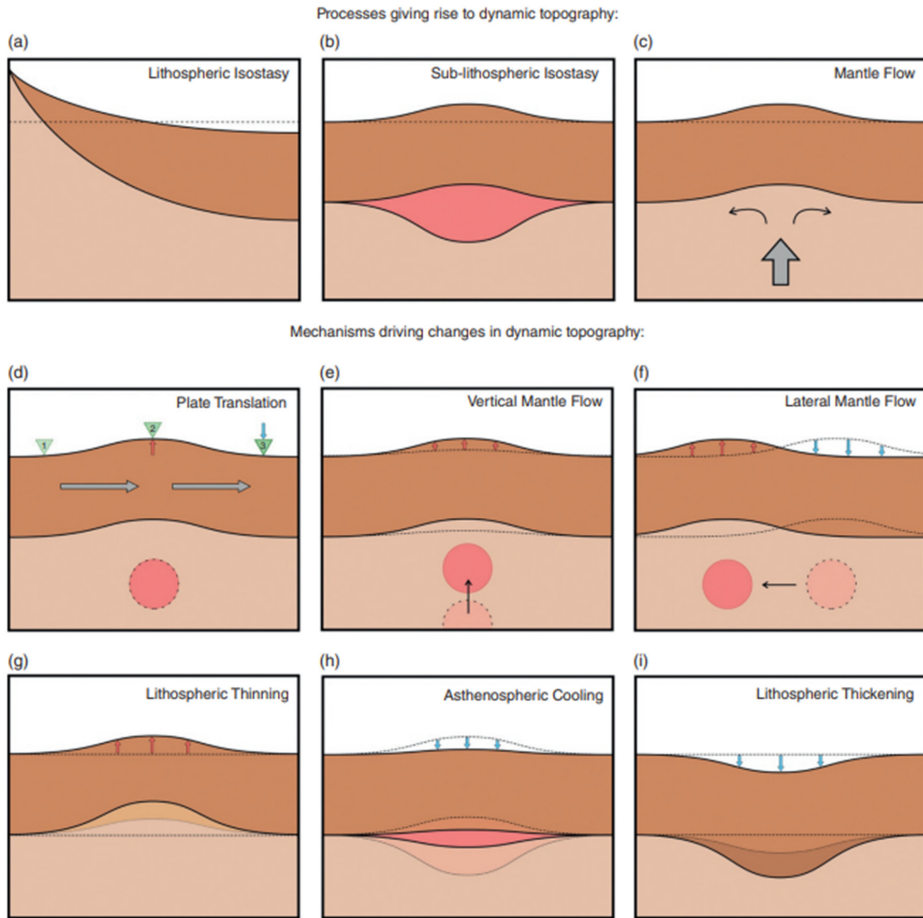


Figure 2.5 – Processes that produce dynamic topography according to different concepts for it (Hoggard et al., 2021). In the top row, three different processes and scales of dynamic topography: A) lithospheric isostasy; B) sublithospheric isostasy (including the asthenosphere); C) flow in the mantle. The lower panels show distinct vertical motions and associated processes, as described in the literature on dynamic topography: D) movement of the plate above a stable density anomaly; E) changes in the vertical flow field in the mantle; F) lateral advection of a density anomaly; G) lithospheric thinning due to e.g. thermal erosion or delamination; H) asthenospheric cooling (e.g. of underplated magma); I) lithospheric thickening due to e.g. ocean spreading.

Surface dynamic topography has great implications for source-to-sink studies, as it may be as important as plate-boundary processes and crustal isostasy for driving uplift and subsidence and generating sedimentary signals (Braun, 2010; Hoggard et al., 2021). This has been demonstrated by case studies in areas such as Austral Africa (Baby et al., 2020; Stanley et al., 2021) and Australia (Czarnota et al., 2014), which invoke one or several of the mechanisms shown in Figure 2.5 as drivers of dynamic uplift. This is also the case of the North Atlantic Ocean and the North Atlantic Igneous Province, main objects of this study.

The North Atlantic realm and the Iceland Plume

The North Atlantic Igneous Province (NAIP) includes magmatic rocks in Eastern Canada, Greenland, Iceland, the Faroe Islands, the United Kingdom, Ireland and offshore Norway (Horní et al 2017; Figure 2.6). These rocks were emplaced in distinct pulses and subprovinces starting before the opening of the North Atlantic Ocean at around ~66-63 Ma (such as in the British Tertiary Igneous Province) and continuing until the present day in Iceland (Saunders et al., 1997; Horní et al. 2017; Wilkinson et al., 2017). NAIP magmatic rocks are distributed along the conjugate margins of the North Atlantic Ocean (mostly offshore) and have provided important constraints on tectonic reconstructions since the inception of continental-drift and plate-tectonic theories (Holmes, 1918; Wegener, 1929; Morgan, 1971).

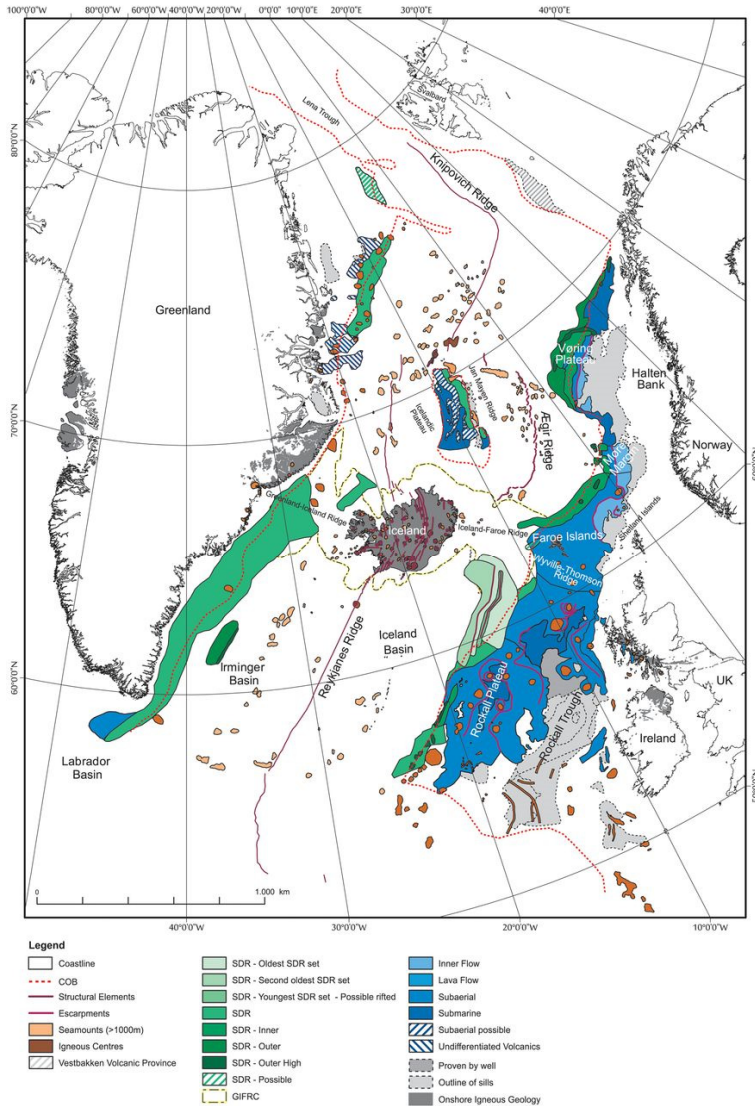


Figure 2.6 – Map of volcanic facies of the North Atlantic Igneous Province (Horní et al., 2017). COB – Continent-ocean boundary; SDR – Seaward dipping reflectors.

As in other large igneous provinces, magmatism within the NAIP has been attributed to the presence of a mantle plume, which would currently be located beneath Iceland (being therefore named the Iceland or Icelandic Plume) as suggested by seismic tomographic studies (Rickers et al., 2013; Celli et al., 2021; Figure 2.7). Many debates surrounding the nature of the Iceland Plume and its connection to geodynamics in the North Atlantic realm persist. This includes discussions regarding: 1) the proposed

simple geometry of the traditional “mushroom” model *versus* more complex geometries detected in seismic velocity reconstructions, suggesting dispersal of a low shear velocity zone along the rifting axis (Foulger et al., 2005; Rickers et al., 2013; Celli et al., 2021); 2) the depth of its mantle-source and the degree of involvement of the lower mantle (Anderson, 2005; Steinberger et al., 2015); 3) the exact magnitude of temperature anomalies (Hole and Natland, 2020) and even 4) the participation of continental crust in the makeup of the Iceland-Faroe-Greenland ridge and its relationship to normal plate-tectonic processes (Foulger et al., 2005, 2020).

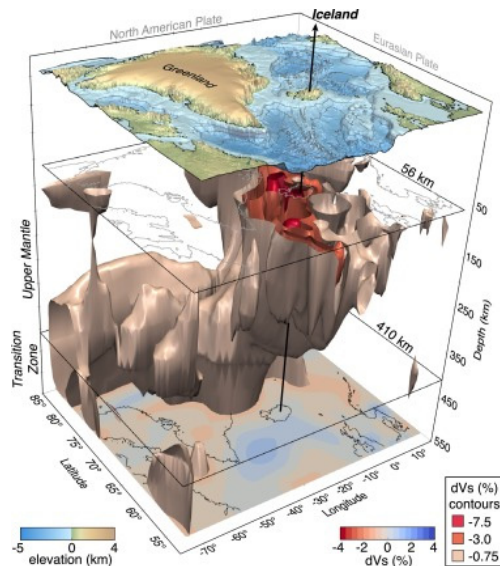


Figure 2.7 – Complex geometry of the Iceland Plume, as shown by the NAT2021 tomographic model (Celli et al., 2021). The plume itself is highlighted with velocity anomaly contours at -0.75 , -3 and -7.5% between 56 and 550 km depth.

The ascension and approach of the Iceland Plume to the base of the lithosphere has been suggested to have occurred during the Paleocene, as indicated by the age of some of the oldest volcanic rocks in the British Tertiary Igneous Province (Steinberger et al., 2019) and the onset of regional inversion, marked by a key “Atlantean” unconformity in several North Atlantic basins (Dam et al., 1998; Gale and Lovell, 2018; Patruno et al., 2022). Subsidence analyses in the North Sea and Faroe-Shetland Basin have been used to constrain subsidence anomalies, which are deviations from predicted post-rift thermal subsidence (Figure 2.8). These could be explained by uplift

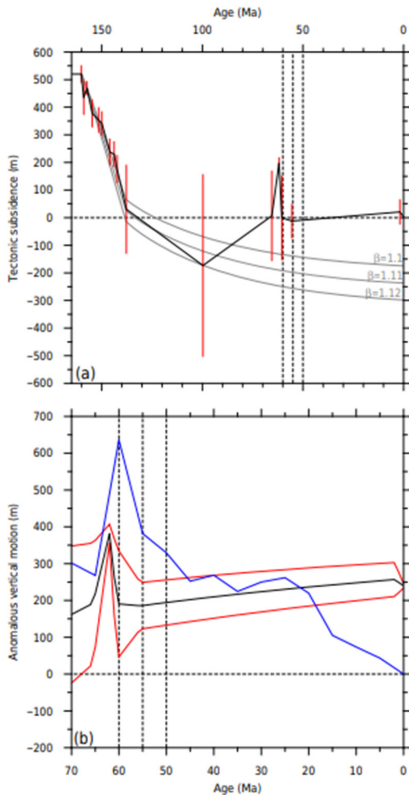


Figure 2.8 - Comparison of subsidence anomalies and modelled dynamic topography, from Barnett-Moore et al. (2017). A) Subsidence history in UK well 13/27-1 (black line) and predicted thermal subsidence (gray lines with different stretching or β factors) in Clift and Turner (1998). B) Anomalous vertical motion in the same well (black lines) and modelled dynamic topography (blue line). Red lines are uncertainty in paleobathymetry.

associated with the approach and impact of the Iceland Plume (Nadin et al., 1995, 1997; Clift and Turner, 1998), peaking either before or close to the breakup of the North Atlantic, at ~55 Ma. Uplift related to the plume would occur by igneous underplating of the crust (Figure 2.5.B) and by support related to ascending and advective flow itself (2.5.E and 2.5.F).

Recent studies on dynamic uplift in the North Atlantic have utilized preserved erosional landscapes around the Shetland Platform to reconstruct catchment morphologies and knickpoints, which allowed for the inversion of uplift rates of the Late Paleocene to Eocene (Hartley et al., 2011; Stucky de Quay et al., 2017; Stucky de Quay and Roberts, 2022). Reconstructed vertical motions point to fast, transient uplift at rates between 100 - >500 m/Myr followed by similarly fast subsidence, in variable agreement with subsidence reconstructions based on wells and seismic data (Bertram and Milton, 1988; Nadin et al., 1997; Clift and Turner, 1998). Fast pulses in uplift have been attributed to rapidly and radially-spreading

thermal anomalies emanating from the plume's center (mechanism F in Figure 2.5; Figure 2.9; Shaw Champion et al., 2008; Hartley et al., 2011; Stucky de Quay et al., 2017). Interestingly, however, the erosional landscapes used for uplift inversion in these works postdate a Selandian peak in modelled dynamic uplift (Barnett-Moore et al., 2017; Figure 2.8) and in measured sediment fluxes across the North Sea (Reynolds,

1994; Liu and Galloway, 1997; Sømme et al., 2019). Accordingly, geomorphological and stratigraphic quantifications of erosion and uplift before the Late Paleocene in the Shetland Platform remain elusive.

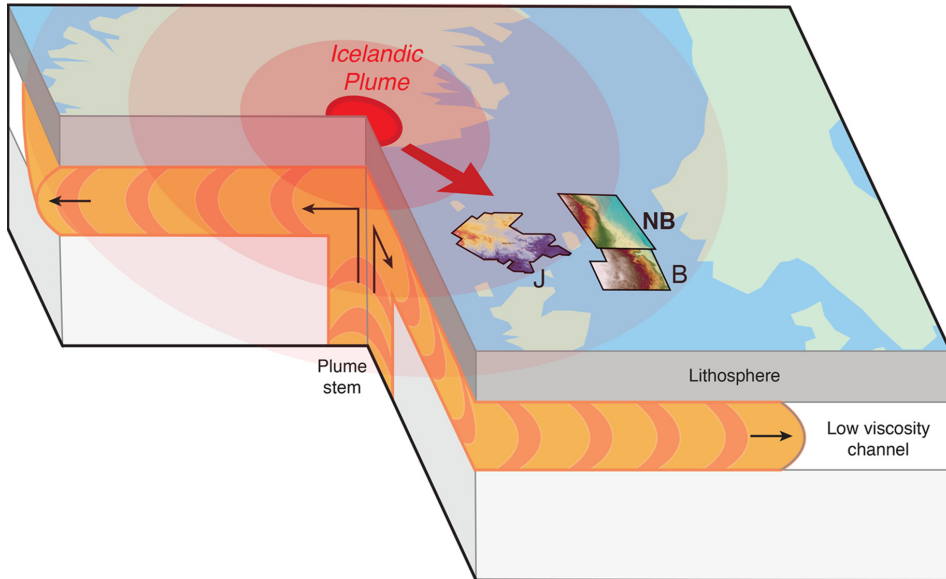


Figure 2.9 – Geodynamic model for the generation of buried erosional surfaces around the Shetland Platform (Stucky de Quay and Roberts, 2022). This model evokes the passage of laterally advecting thermal anomalies through the asthenosphere, creating waves of uplift and erosion. J – Judd catchment (Shaw Champion et al., 2008; Hartley et al., 2011); B – Bressay Channel (Underhill, 2001; Stucky de Quay et al., 2017); NB – Northern Bressay (Stucky de Quay and Roberts, 2022).

Critically, plume impingement and LIP emplacement occurred at the same time as continental rifting and seafloor spreading in the North Atlantic (Nadin et al., 1997; Fletcher et al., 2013). In the absence of clear evidence of extension in the North Sea in the Paleocene (Bertram and Milton, 1988; Milton et al., 1990), little consideration has been given to rifting as a potential driver of uplift in the studies proposing the model of Figure 2.9 (Hartley et al.; 2011, Stucky de Quay et al., 2017). However, many authors have pointed to rifting as a driver of regional tectono-stratigraphic responses in the North Atlantic (Doré et al., 1999; Fletcher et al., 2013; Stoker et al., 2018), and a potential impact in the Shetland Platform and the North Sea cannot be neglected. This is particularly true if we consider that uplift and exhumation in the observed scales are not unique responses to either tectonic or dynamic vertical motion.

3. Study areas

This project involved the analysis of a large volume of subsurface data from the Northeast Atlantic and the North Sea, focusing on sedimentary basins around the modern-day Shetland Isles and the Shetland Platform, a ~NE-SW structural high located north of the Scottish mainland (Figure 3.1).

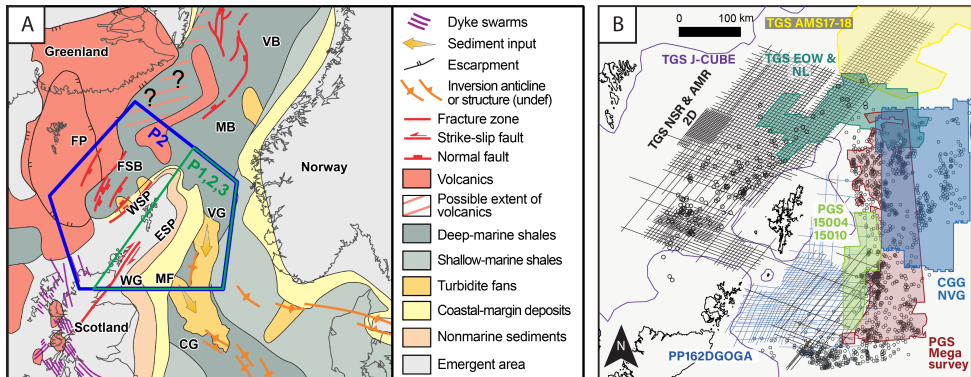


Figure 3.1 – A) Paleogeographic reconstruction of the Paleocene, modified from Coward et al. (2003). Location of the study areas in each paper shown by the black (P1,2,3) and blue (P2) polygons. While Paper 3 focuses on the East Shetland Platform, it includes a comparison of other North Atlantic basins. B) All datasets used in this thesis, including Multiclient 3D data, 2D data and more than 1700 wells available through the TGS FMB database. More details are described in the papers.

Like the main Scottish landmass, the Shetland Platform was assembled during the collision between Avalonia, Laurentia and Baltica during the Caledonian orogenic cycle, in the Late Cambrian to Early Devonian (Leslie et al., 2008; Searle, 2021). The Caledonian Orogeny entailed the amalgamation of distinct Archean-Neoproterozoic terranes throughout Scotland, with the Shetland Platform being considered part of the Northern and Grampian Highland terranes, cut-through by the Great Glen Fault (Leslie et al. 2008; Figure 3.2).

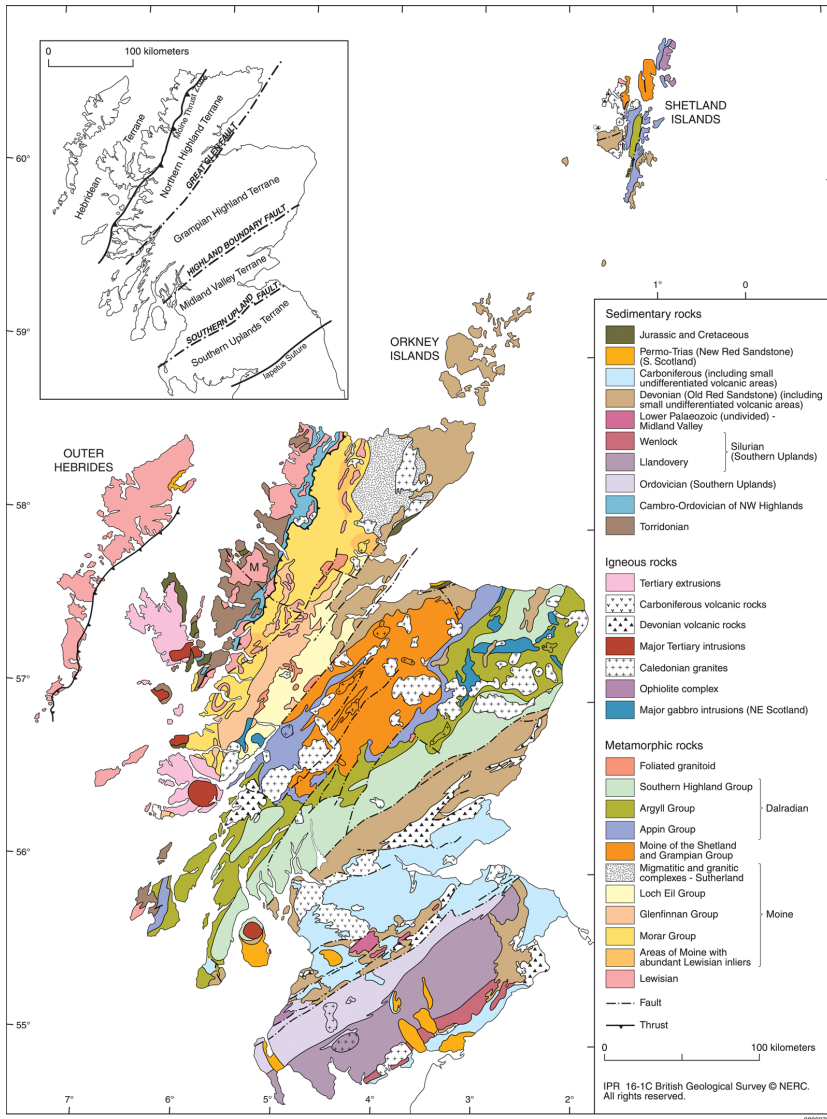


Figure 3.2 – Geological map of Scotland from Leslie et al. (2008), with tectonic terranes from Strachan et al. (2002). Reproduced with the permission of the British Geological Survey © UKRI 2024. All Rights Reserved.

Collapse of the Caledonian orogen during the Devonian resulted in overall extension, conditioning post-orogenic sedimentation around the Northeast Atlantic realm (McClay et al., 1986; Fossen, 2010; Figure 3.3). In Scotland, remnants of these post-orogenic basins form important parts of the offshore and onshore geology (e.g.

Orcadian Basin and “Old Red Sandstone”). Studies on sedimentary provenance in the North Sea Basin have demonstrated significant contributions from eroded Caledonian rocks and post-orogenic Devonian-Carboniferous sediments throughout the basin’s history (Morton, 1979; Luzinski et al., 2022), meaning that their geographic and structural configuration are particularly relevant for source-to-sink sedimentology in the North Sea (Figure 3.4).

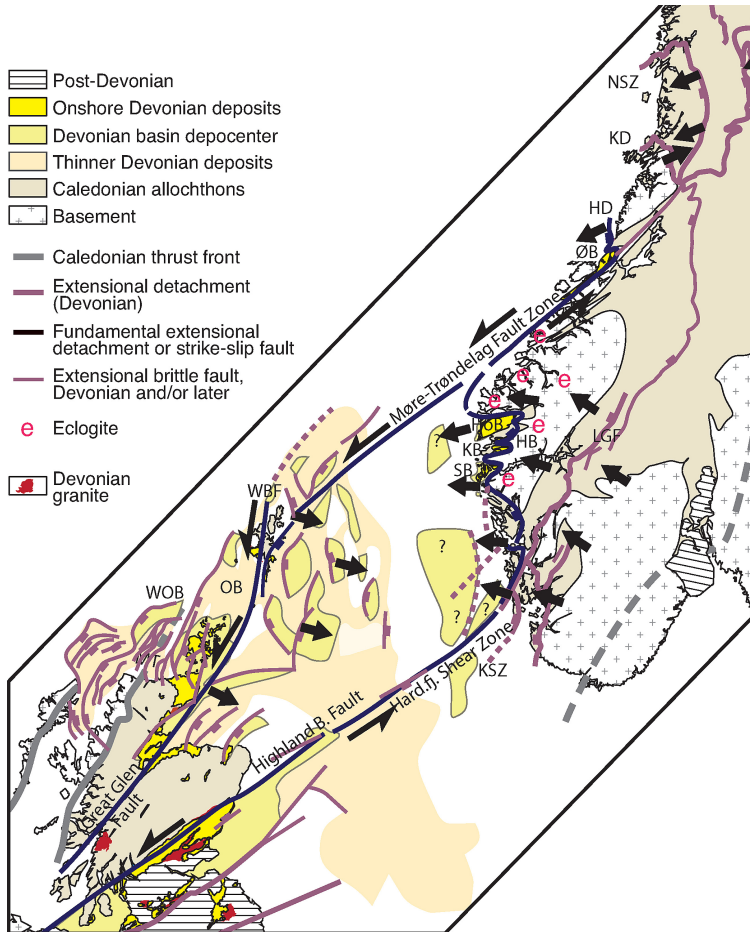


Figure 3.3 – Map of Devonian basins in the North Sea, showing their relationship to major extensional/strike-slip structures like the Møre-Trøndelag Fault Zone and the Hardangerfjord Shear Zone (Fossen, 2010).

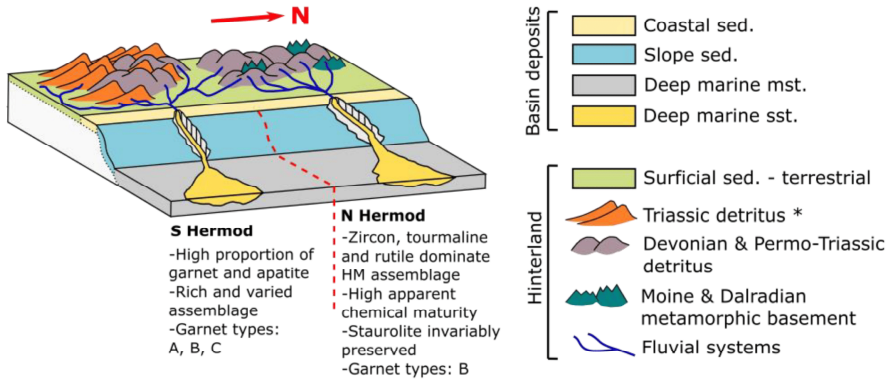


Figure 3.4 – Conceptual model for sediment provenance in the Hermod Fan (Dornoch Formation, Paleocene – Eocene transition) in the East Shetland Platform, showing different sources of Devonian and Permian-Triassic rocks, as well as smaller contributions of metamorphic basement. Modified from Luzinski et al. (2022).

Post-orogenic sedimentation was succeeded by episodes of intra-plate rifting. Broad rifting in the Permian-Triassic resulted in the inception of the current structural disposition of the North Sea (Figure 3.5; Ziegler, 1992; Færseth, 1996; Erratt et al., 1999; Phillips et al., 2019). A second rifting episode started in the Late Jurassic. Extension was responsible for more focused faulting in the North Sea, with progressive strain localization towards the axis of the Viking, Witch Ground and Central grabens until ~140 Ma in the Early Cretaceous (Zanella and Coward, 2003; Coward et al., 2003; Phillips et al., 2019). Rifting resulted in further NE-SW faulting in the Faroe-Shetland Basin, where extension lasted longer, until the Turonian – Santonian (~95-85 Ma) in the Late Cretaceous (Stoker, 2016).

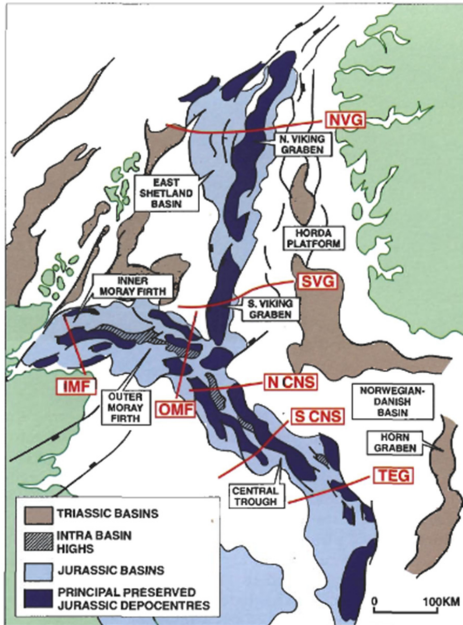


Figure 3.5 – Main basins associated with Triassic and Jurassic rifting in the North Sea (Erratt et al., 1999).

In the Norwegian Sea and virtually absent in the North Sea (Bertram and Milton, 1988; Doré et al., 1999), with the Faroe-Shetland Basin potentially situated as a failed rift in-between (Fletcher et al., 2013).

Despite the apparent absence of extension in the North Sea, the area was subject to significant inversion, detected initially by the observation of subsidence anomalies in the Paleocene, as explained in the previous chapter (Bertram and Milton, 1988; Nadin et al., 1997; Clift and Turner, 1998). Inversion and uplift have been linked to a major increase in siliciclastic input around the Shetland Platform (Liu and Galloway, 1997; Smallwood, 2008), with some studies proposing a connection between numerous pulses of sand deposition and individual inversion events (White and Lovell, 1997; Mudge and Jones, 2004). As discussed in Chapter 2, uplift has been often attributed to the impingement of the Icelandic Plume during the Paleocene and its associated thermal doming (White and Lovell, 1997; Jones and White, 2003; Coward et al., 2003; Shaw Champion et al., 2008). Other studies suggest plate-tectonic processes as alternative causes, such as potential rift-shoulder uplift or even intra-plate compression

In the North Sea, post-rift thermal subsidence resulted in the accumulation of a 1-3 km thick Cretaceous succession. In the Late Cretaceous, this was marked by chalk deposition in the southern and central North Sea (Surlyk et al., 2003), shallow equivalents of which are observable in several famous outcrops in southern England, such as the White Cliffs of Dover.

Thermal subsidence was punctuated by a third and final rifting episode occurring throughout the North Atlantic realm during the Late Cretaceous to the Early Eocene. Detected extension varied along the rift-axis, being more significant

associated with spreading (Doré et al., 2008; Mudge, 2015; Ellis and Stoker, 2018; Stoker et al., 2018; Foulger et al., 2020).

East Shetland Platform and Northern North Sea Basin

A regional unconformity between Late Cretaceous (or older) and Late Danian (or younger) deposits has been tied to the onset of inversion in the North Atlantic (“Intra-Danian Unconformity” in Mudge, 2015; “Atlantean Unconformity” in Patruño et al., 2022). In the Northern North Sea Basin, this accompanies a major change from the deposition of chalk to an increase in carbonate reworking and siliciclastic input through deep-water channel and fan systems, corresponding to the Våle Formation (Danian – Selandian), mostly recorded in the Viking Graben (Figure 3.6). The overlying Lista Formation (Selandian – Early Thanetian) is marked by prominent terrigenous input, with a more distributed and better-preserved record in the East Shetland Platform (Figure 3.6). The Lista Formation hosts major sand accumulations in submarine channel and fan settings (including those of the Heimdal, Balder and Alvheim fields), many of which are now found as injectite complexes (Bergslien, 2002). Both formations are part of the Montrose Group.

The overlying Moray Group (Late Thanetian - Early Ypresian) was deposited during a major episode of shelf progradation in the Northern North Sea, and includes sediments from non-marine to deep-water environments (Figure 3.6). This progradational episode is connected to the erosional and uplift events discussed in Chapter 2. In the North Sea, this has been captured in the uplift inversion based on knickpoints obtained by Stucky de Quay et al. (2017). Breakup of the North Atlantic initially occurred at some point during magnetic chron C24, at roughly 55-53 Ma, soon after the maximum progradation and regression of the Dornoch Formation shelf. Breakup was accompanied by widespread volcanism and ash deposition, which resulted in a key stratigraphic marker in the Balder Formation, the Balder Tuff, recognized across the North Atlantic (Figure 3.6; Watson et al., 2017).

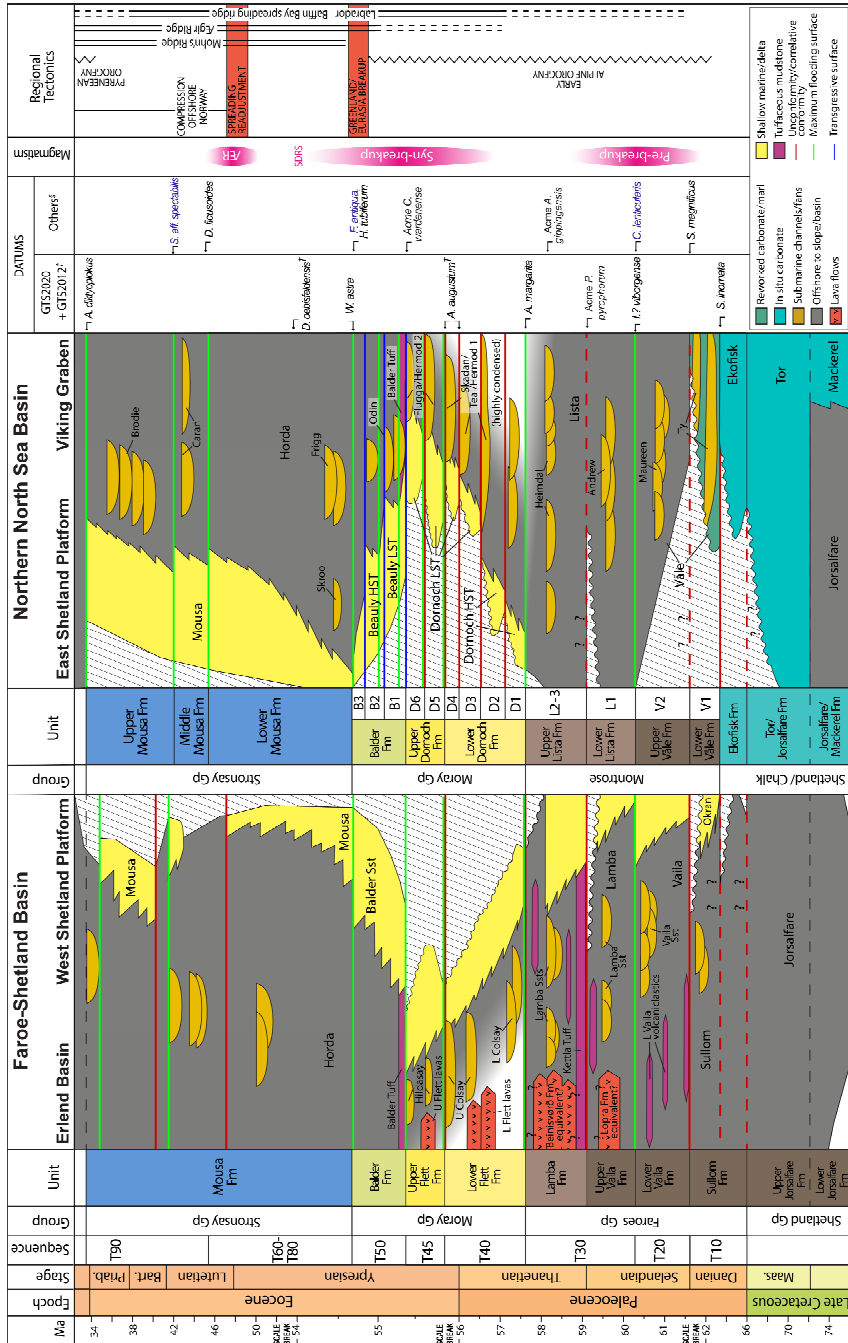


Figure 3.6 –Chart comparing litho-, chrono-, bio- and sequence stratigraphic schemes across West and East of Shetland. Biostratigraphy is based on GTS2020 (Speijer et al., 2020) or 2012 (with †, Vandenberghe et al., 2012). Other datums (§) are based on correlation of datums and surfaces in Mudge and Bujak (2001) and Brunstad et al. (2013) to ages in Mudge (2015), Speijer et al. (2020) and Jolley et al.

(2021). *Magmatism and regional tectonics from Gernigon et al. (2012). Cretaceous stratigraphy based on Surlyk et al. (2003) in the NNSB and Stoker (2016) in the FSB. Eocene stratigraphy above the Balder Formation based on Mudge and Bujak (1994) in the NNSB and Stoker et al. (2012) in the FSB. Subdivisions of the Dornoch Formation from **Paper 1. ÆR – Ægir Ridge; SDRs – Seaward dipping reflectors.***

The Faroe-Shetland Basin

Dominant slope/basin deposition took place during the Danian-Early Thanetian, resulting in the Sullom, Vaila and Lamba formations (Figure 3.6) Contrasting to the North Sea, however, studies have suggested a larger extent of preserved shelf progradation during this interval (Naylor et al., 1999; Mudge, 2015). Interestingly, a thick shelf-edge succession is not as well-imaged in the Late Thanetian – Early Ypresian Flett and Balder formations in the West Shetland Platform, although non-marine deposits are also widely recognized (Mudge, 2015; Jolley et al., 2021; Walker et al., 2022b).

An important difference in the Faroe-Shetland Basin is the presence of extrusive and intrusive volcanics and volcanoclastic units of the North Atlantic Igneous Province, which form an escarpment covering much of the Faroese platform and extending towards the north in the Norwegian Sea (Figure 3.1). Studies suggest that Paleocene – Early Eocene magmatic activity in the NAIP was focused in two distinct phases, a **pre-breakup** phase in the Selandian and a **syn-breakup** phase during or shortly after the Paleocene – Eocene transition (Storey et al., 2007; Gernigon et al., 2012; Jolley et al., 2021; Walker et al., 2022a). It has also been suggested that these phases coincide with peak tectonic activity/thermal uplift around the British Isles (Naylor et al., 1999; Shaw Champion et al., 2008; Łuszczak et al., 2018; Conway-Jones and White, 2022).

Paleocene-Eocene Thermal Maximum

Deposition of the Dornoch and Flett formations partly occurred during the Paleocene-Eocene transition, a time marked by the most intense discrete global warming event of the pre-Holocene Cenozoic - the Paleocene-Eocene Thermal Maximum (Zachos et al., 2001; Jones et al., 2013). It occurred before the Early Eocene peak of a longer-term global warming trend that ensued from Late Cretaceous to

Paleocene times (the Early Eocene Climatic Optimum, EECO – 53 to 50 Ma), after which milder, slowly cooling “hothouse” to “warmhouse” temperatures and less distinguished Eocene hyperthermals followed (Zachos et al., 2001; Abels et al., 2012). Its origin has been related to different processes in the literature, such as greenhouse gas release caused by the interaction of volcanic intrusions, carbonates and organic-rich shales (Svensen et al., 2004) and large-scale, extrusive LIP volcanism (Gernon et al., 2022).

In several basins containing Paleocene to Eocene strata in the world, it has been possible to recognize a distinct PETM signal on the basis of C13 excursions (Schmitz and Pujalte, 2007; Jones et al., 2019; Jin et al. 2022) or palynological data, for example (Sømme et al., 2019, 2023; Jolley et al., 2023). In these systems, discrete paleoenvironmental modifications associated with the PETM have been identified with multiple sedimentological and geochemical proxies. Recognized changes include varied source-to-sink responses, such as: (i) overall increased runoff and sediment flux, which influenced the architecture of fluvial deposits (Foreman et al., 2012) and are also detected by Li isotopes (Pogge von Strandmann et al., 2021); (ii) increased chemical (Ravizza et al., 2001) or physical weathering rates (Ji et al., 2023); (iii) initial increased intra-annual humidity gradients and frequency of extreme climatic events (storms) which also influenced fluvial deposition (Schmitz and Pujalte, 2007). Overall, it seems clear that much like modern global warming, the PETM has promoted enhancement of the hydrological cycle, intensification of monsoonal weather and modification of pre-existing climate envelopes (Schmitz and Pujalte, 2007; Inglis et al., 2020; Korasidis and Wing, 2023).

4. Summary of papers

The results from this thesis are divided in three original research articles covering themes surrounding the tectono-stratigraphic framework of the Shetland Platform and its surrounding basins, focusing on data-driven constraints on source-to-sink evolution during the Paleocene - Early Eocene.

Article 1: Paleogeography and 3D variability of a dynamically uplifted shelf: observations from seismic stratigraphy of the Paleocene East Shetland Platform

This study's main objective is to investigate the detailed seismic stratigraphic record of the East Shetland Platform using new, high-resolution 3D data, comparing our observations with several of the previous stratigraphic and paleogeographic frameworks suggested for the area. The goal is to obtain data-driven constraints for the paleogeographic evolution in the area, which are further used for **Papers 2 and 3** and should also be useful for future studies. This dataset covers a shallower area of the East Shetland Platform that had very limited 3D coverage before, allowing for a better grasp of proximal depositional geometries and along-strike variability, particularly relevant for the Late Paleocene – Early Eocene clinoform succession of the Moray Group (Figure 3.6).

Through detailed seismic stratigraphic mapping it was possible to assess the clinoform succession of the Moray Group in unprecedented detail, resulting in the definition of a total of 9 units with sequence stratigraphic significance, which are defined by unconformities and flooding surfaces with varied along-strike expression. These were previously grouped in three genetic sequences (bounded by maximum flooding or condensation surfaces) in many stratigraphic schemes, with only broad paleogeographic characterizations being available in the literature. Importantly, improved 3D coverage of proximal sections of the Moray Group allowed us to reconstruct the extent of 5 individual unconformities inside the Dornoch Formation, a key observation for comparisons with previous models of unconformity development and uplift events in the Shetland Platform (*e.g.* Mudge, 2015; Stucky de Quay et al., 2017).

A key observation is that the Dornoch Formation is regionally subdivided in a highstand and a lowstand succession, which are separated by >300 m of clinoform downstepping, indicative of significant relative sea-level fall. This appears to have been a regional event, although traces of its occurrence were likely removed by erosion in the Bressay High, an area in the center of the ESP where the Dornoch highstand is absent.

Higher vertical resolution and more extensive lateral coverage of our 3D dataset allowed the improved extraction of seismic geomorphological constraints on depositional environments. Observations from this study confirm the prevalence of shallow marine and paralic deposits in the Moray Group, with evidence of distinct fluvial sediment entry points feeding wave-dominated shores subject to lateral sediment redistribution. This has important implications on the balance between accommodation and supply, as it potentially modulated the development of individual sequence stratigraphic responses, such as the internal flooding surfaces of the Beaulieu Member and the prominent but localized lowstand Beaulieu. Additionally, the development of unconformities and forced regressive wedges (progradational wedges with downstepping clinoforms and truncated topsets) is markedly varied along-strike. This indicates that local variability of accommodation space was also important, as also highlighted by the complete erosion of the Dornoch highstand in the Bressay High mentioned above.

In general, results indicate that a complex interplay of laterally varied accommodation, sediment entry point distribution and along-shore transport resulted in significant along-strike variability. Paleogeographic constraints suggest a first-order control on erosion promoted by transient uplift, with what appear to be short-wavelength and short-period variations in uplift rates (or relative sea-level fall, as we do not attempt to quantify uplift itself in this contribution). Importantly, this contrasts with typical characteristics of dynamic topography and with proposed models of fast transient uplift caused by laterally advecting asthenospheric anomalies (Stucky de Quay et al., 2017, referred in Chapter 2). Lastly, our constraints on paleogeography and stratigraphic evolution indicate that the effects of the PETM for sediment routing are

likely masked by contemporaneous uplift, which had a first-order control on shelf reworking and submarine fan distribution.

Article 2: High-resolution sediment mass-budget of the Shetland Platform during the Paleocene – Early Eocene - constraints on external forcing by uplift and climate

This paper aims to provide the first high-resolution sediment mass-budget of the Shetland Platform using seismic and well data, building on the stratigraphic observations and framework of **Paper 1** for a coherent tie across different basins. We used an even larger 3D seismic dataset that continuously covers the Faroe-Shetland, More and North Sea basins for a total of >200 000 km², also including large extents of tightly spaced 2D data (Figure 3.1). This dataset allowed unprecedented control of depositional extents and sediment transport directions, key for determining the contributions of different source-areas in the studied basins. It also allows a better constraint of volcanic rocks in the Faroe-Shetland Basin, which need to be removed before calculations of deposited sediment masses are carried out. We also include constraints from a large well dataset including >1700 wells in the North Atlantic realm, from which we extract stratal thickness maps that can be compared to thickness measurements obtained from seismic data. We can then compare results with the previous highest resolution estimate for sediment volumes in the North Sea, which was mostly based on well data (Liu and Galloway, 1997).

In order to obtain a *mass-budget*, constraints on the masses of eroded material in the source-area are evidently needed in addition to measurements of deposited sediment masses. Estimates of eroded material are often obtained by low-resolution low-temperature thermochronology in the North Atlantic (Anell et al., 2009; Wilkinson, 2017), which suggests between 1-3 km of denudation in the British Isles during the Cenozoic (Holford et al., 2010; Łuszczak et al., 2018). For a higher-resolution constraint, however, we use the empirical-statistical “BQART” model of Syvistki and Milliman (2007) to estimate the mass of generated sediments in each time-step of interest. Importantly, this model allows first-order constraints on the impacts of lithology, water discharge, catchment area, maximum relief, and atmospheric

temperatures on sediment generation, which are key aspects we aimed to evaluate. This has key implications for any attempts to distinguish between uplift-related signals and climate-related ones, particularly the PETM.

Results suggest that two peaks in sediment accumulation rates occurred in both the Faroe-Shetland and North Sea basins, which were fed by catchments in the West and East Shetland Platform, respectively. These peaks occurred in the Mid-Paleocene (during the Selandian – Early Thanetian) and close to the Paleocene – Eocene transition. Importantly, these peaks are captured in slightly different ways in both basins, because the well and seismic datasets have different temporal resolutions. One resulting contrast is that a syn-PETM unit and peak can be distinguished in the North Sea (based on seismic data), but a syn-PETM unit could not be subdivided in the Faroe-Shetland Basin (where the peak detected with well data postdates the hyperthermal). This is consistent with previous constraints based on well data in Liu and Galloway (1997).

Based on our sediment mass-budget, estimates of catchment parameters derived from a combination of literature data and our own interpretation are broadly reasonable, resulting only in a slight mismatch for predicted and observed sediment masses for a few intervals. This small mismatch is mainly due to the broad range in relief values, which can be better investigated using an inverted approach, where we solve for maximum relief in the BQART equation using measured sediment accumulation rates as an input. Constraints on maximum relief provide a remarkably good fit with independent estimates of uplift rates and relative sea-level fluctuations, suggesting that uplift is the main cause for both peaks in sediment production and deposition, with a minor contribution of both long-term and sharp warming at the Paleocene – Eocene transition.

Article 3: Short-wavelength uplift variations in the Paleocene East Shetland Platform reveal lithosphere imprint on plume-related support

Paper 3 is a short-form study where we further develop the relatively broad constraints on relative sea-level fluctuations obtained in **Paper 1** and explore their direct implications for geodynamic models. To do so, we focus our observations on

three specific seismic transects in the East Shetland Platform. We then use high-resolution backstripping to extract relative sea-level curves based on clinoform rollover bathymetries and trajectories, removing the effects of incremental sediment loading, compaction, thermal subsidence and eustasy. This results in a more robust assessment of tectonic/dynamic vertical motions (uplift and subsidence) that are not predicted by thermal subsidence models (*i.e.* subsidence anomalies). As in **Paper 1**, the analysis of different transects distributed along-strike in the East Shetland Platform allowed us to constrain the spatial and temporal variability of these vertical motions in detail.

Our results confirm that transient uplift in the Late Paleocene – Early Eocene occurred at a rate faster than what is suggested by models of long-term dynamic topography (Barnett-Moore et al., 2017). After comparison with uplift events and curves proposed across the North Atlantic, we find that uplift was overall synchronous in different basins, as constrained by the resolution of biostratigraphic data used for such comparison. Uplift appears to have occurred in two distinct phases, one in the Late Thanetian (Late Paleocene) and another in the Early Ypresian (Early Eocene), confirming a recent suggestion for the West Shetland Platform (Jolley et al., 2021; Walker et al., 2022b). In East Shetland, regional uplift occurred in the Late Thanetian, corresponding to the relative sea-level fall event between the Dornoch highstand and lowstand referred to in **Paper 1**. The same event has been hinted at in West Shetland, although its record is more elusive there due to significant erosion of the equivalent Late Thanetian Flett Formation (Jolley et al., 2021). Uplift appears to have been localized in the Early Ypresian East Shetland, with significant lateral variability in magnitude, whereas in the West this event appears to be predominant in the stratigraphic record.

Important conclusions can be drawn. A comparison of events throughout the North Atlantic does not confirm diachronic uplift from west to east, a requirement of the “blob model” (Hartley et al., 2011; Stucky de Quay et al., 2017). However, even if synchronous, fast uplift also does not agree with “traditional” models of dynamic topography (*e.g.* Barnett-Moore et al., 2017). Furthermore, marked variability in the magnitude of the Early Ypresian uplift event suggests local controls on its expression, which also contrasts with long-wavelength dynamic topography. These observations

point to a more complex scenario, where any dynamic topographic input could have been modulated by lithospheric or crustal features. For instance, mantle-lithosphere interaction could allow for the generation of short-wavelength and short-period topography (*e.g.* Burov and Guillou-Frottier, 2005; Burov and Gerya, 2014). Any primary long-wavelength dynamic uplift signal could also be affected by differential isostatic responses, an idea that is currently undertested (Hoggard et al., 2021).

6. Synthesis

Main results and implications

Paleogeography and tectono-stratigraphic framework in the Shetland Platform

In all **3 papers**, paleogeography and tectono-stratigraphy were approached in different scales. The stratigraphic and paleogeographic framework obtained in **Paper 1** is of a higher resolution than ever before, and offers greater insight into proximal and non-marine depositional environments in the Moray Group (particularly in the Dornoch Formation lowstand units). Using seismic geomorphology and well data, the locations of fluvial sediment entry points can be suggested. These appear to be related to distinct drainage networks and incised valley systems formed during forced regression. These drainage networks were formed during relatively fast erosion of the Dornoch coastal plain, subsequently flooded and buried during overall transgression. This is consistent with previous observations of intermittent and fast fluvial erosion across the Shetland Platform in the Late Paleocene – Early Eocene, which has been associated with fluctuations in dynamic uplift (Underhill, 2001; Hartley et al., 2011; Stucky de Quay et al., 2017; Hardman et al., 2018).

In addition to forming coastal plain incised valleys, fluvial systems of the Dornoch Fm lowstand may have been connected to larger drainage systems updip of the intermittently exposed coastal plain, some of which are partially preserved and imaged in seismic data (the newly described West Bressay drainage system **Paper 1**). This is also suggested by the large extent and distribution of individual submarine fan systems in comparison to the size of the incised valleys. Broader erosion and shelf reworking are also indicated to be important sources of sediment for some of these fan systems, such as in the pre-PETM Hermod fan in the Central Viking Graben, likely fed by erosion in the Bressay High. Importantly, shelf reworking here was also favored by uplift.

The stratigraphic contribution of individual sediment entry points and sediment routing systems distributed along strike appears to have varied through time, controlled by differential supply and accommodation (Figure 5.1). Another key factor for

differential supply along-strike could be sediment redistribution in wave-dominated shores, as shown by differences in the development of deltas and clinofans in the Dornoch lowstand (Figure 5.1.C). or, even more strikingly, the stark along-strike contrasts of the Beaulieu member (Figure 5.1.D).

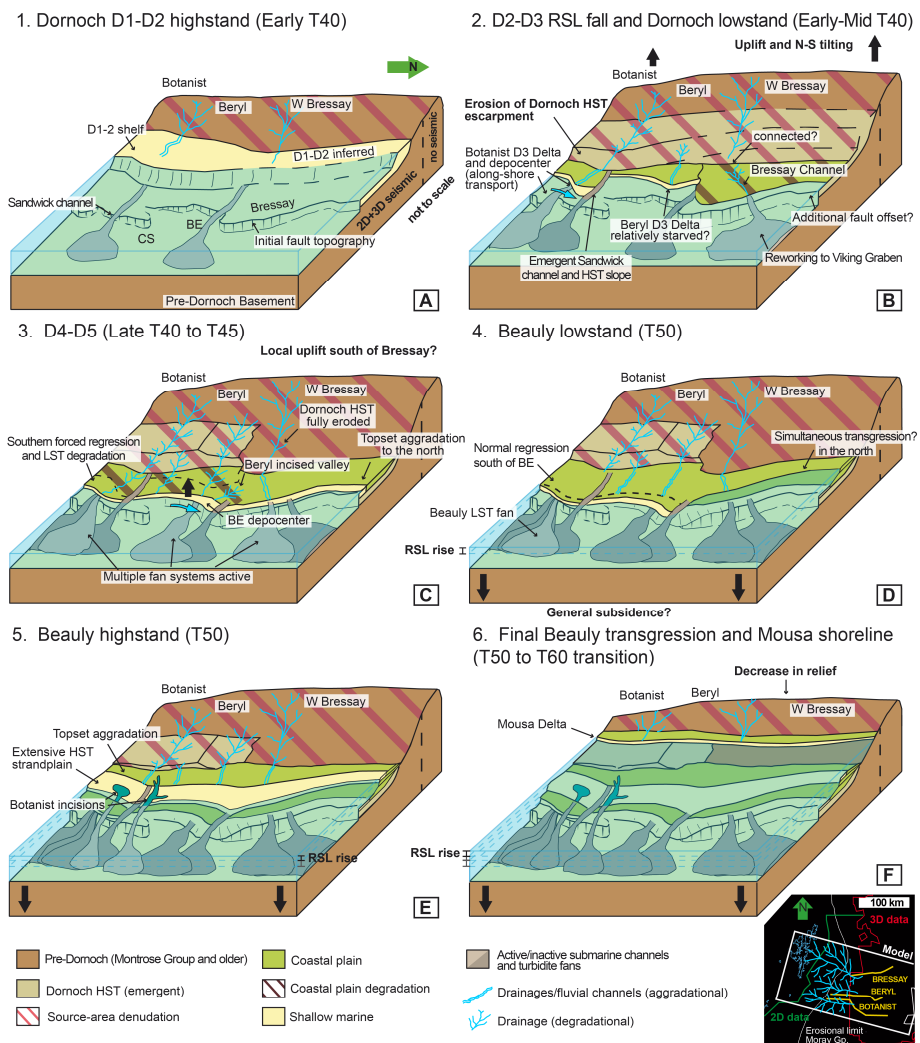


Figure 5.1 – Simplified model of the paleogeographic evolution of the East Shetland Platform in the Paleocene – Early Eocene. After uplift and tilting of the platform, forced regression, normal regression and transgression were varied along-strike.

This has key implications for along-strike sequence stratigraphy, as differential supply and accommodation modulate the development of sequence stratigraphic surfaces

(unconformities, transgressive and maximum flooding surfaces). Three-dimensional mapping of seismic surfaces and their correlation to biostratigraphy suggests that these surfaces can be reliably tracked along-strike, even when their stratigraphic character changes laterally (see also Paumard et al., 2019). However, the reliability of individual stratigraphic frameworks based on regionally extrapolated sequence stratigraphic surfaces can be hindered by such local variability. For instance, a sequence stratigraphic framework derived from an individual 2D line such as the South ESP transect would be intrinsically different from one obtained in the Bressay area in **Papers 1 and 3**, because the number of recognizable systems tracts, unconformities and maximum flooding surfaces within the same time-interval varies between these areas. A key conclusion is that, in the presence of high-quality and wide-coverage data, this problem can be minimized or eliminated by employing seismic stratigraphic mapping and calibration to a coherent paleogeographic and time-stratigraphic framework (Paumard et al., 2019; Sylvester et al., 2024).

This has some consequences for our comparison with the West Shetland domain in **papers 2 and 3**. Our analysis suggests that key stratigraphic trends and surfaces can be sufficiently well correlated across West and East Shetland, especially those tied to clear biostratigraphic markers. However, uncertainties in biostratigraphic ages and in the exact nature of certain stratigraphic surfaces should be carried forward during such correlation. For instance, while the presence of a “Top Selandian” subaerial unconformity has been hinted at in the West (Mudge, 2015; Conway-Jones and Roberts, 2022), any preserved record of erosion in the East would be likely submarine, and the Selandian could rather have been a time of accommodation increase in the East Shetland Platform (Figure 2 in **Paper 2**). Therefore, a *sequence stratigraphic* correlation between West and East is not straight-forward in this interval, while biostratigraphic or chronostratigraphic correlation provide better results. On the other hand, due to the low temporal resolution of biostratigraphy in the Late Thanetian, is not possible to match the exact ages of the Upper Thanetian Unconformity in the West and the D2-D3 RSL fall event in the East, and they can only be included together in the same 57.6 – 56 Ma interval (Figure 3 in **Paper 3**), even if subdivisions in the East have a higher temporal resolution.

The goal in **Paper 3** is to obtain quantitative constraints on vertical motions using the clinoform stratigraphic framework of **Paper 1**. This follows previous backstripping efforts on the Shetland Platform based on seismic data (Bertram and Milton, 1988; Nadin et al., 1997). However, whereas many papers only obtain one bathymetric constraint for the entire Paleocene, **Paper 3** aims to extract a detailed curve for the Late Thanetian – Early Ypresian Moray Group, investigating all forced regression events observed in **Paper 1**. Our results are broadly similar to previous uplift curves extracted in West and East Shetland based on the geomorphology of buried erosional surfaces (*e.g.* Hartley et al., 2011; Stucky de Quay et al., 2017). Critically, however, the occurrence of two uplift events is confirmed, rather than just one - the first in the Late Thanetian (pre-PETM) and the second in the Early Ypresian (syn-post-PETM), as recently suggested by Jolley et al. (2021) and Walker et al. (2022b). This provides independent tectono-stratigraphic confirmation of geomorphological and qualitative indications of uplift across the Shetland Platform, showing that clinoform backstripping can be a powerful tool for the “deconvolution” of relative sea-level curves and for uplift analysis.

Sediment mass-budget and source-to-sink significance

Results from **Paper 2** confirm several of the broader implications of the tectono-stratigraphic framework proposed in **Paper 1**, also allowing independent insights for intervals with unclear sequence stratigraphic significance, such as the Selandian, mentioned in the paragraphs above. Our sediment mass-budget and constraints on catchment parameters confirm the importance of increases in maximum relief values to drive two distinct pulses of high sedimentation rates in the West and East Shetland systems. These increases in relief can be thought of as proxies for uplift (*e.g.* Okwara et al., 2023), although they are not direct measurements of uplift because relief is also modulated by erosion (as discussed in Chapter 1). It is reasonable to assume that changes in topography will broadly follow uplift-driven signals as the landscape evolves toward steady-state, but this will also be modulated by any transient states induced by climate (Figure 5.2).

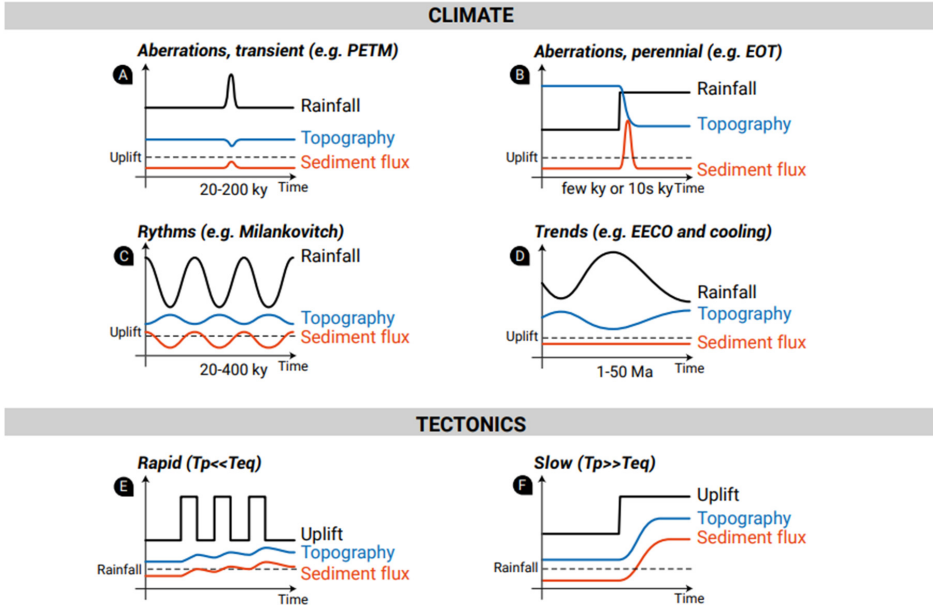


Figure 5.2 – Characteristics of different tectonic and climate signals and their expression as topographic and sediment flux changes (Castelltort et al., 2023). T_p – Timing of perturbation; T_{eq} – equilibrium (steady-state) time.

Importantly, however, the LRoART equation employed in **Paper 2** has no intrinsic consideration of landscape transience and steady-state, something that is relevant for both tectonic and climate-driven signals, but perhaps even more so for the latter (Romans et al., 2016). As explained in Chapter 1, it is possible that long-term climatic trends are underrepresented in the sedimentary record, because the landscape has time to adjust to changes in the erodibility parameter, leading to stable Q_s outputs over time (Figure 5.2.D; Tofelde et al., 2021, Castelltort et al., 2023). It may be difficult for the LRoART method to accurately represent this form of long-term climatic change, as increases in T (temperature) and R_o (runoff) in the LRoART equation will always lead to an increase in Q_s output - in other words, the self-limited behavior of a discharge signal shown in Figure 5.2.D will not be obtained. Perhaps this could be solved by including a corresponding decrease in R (relief) values (as shown in the lower topography induced by higher rainfall in Figure 5.2.D), but it is not clear how this value of R would be determined *a priori*. However, it is possible that inverse reconstructions

of R (such as in **Paper 2**) actually provide an approximation of this effect. For instance, if increasing T or R_o values are used as an input in the inverse method along with fixed values for Q_s (and for all other parameters) then the predicted R will necessarily decrease to maintain proportionality.

On the other hand, the expression of rapid climate modifications (such as the PETM) may be better captured by the LRoART. In nature, the expression of any Q_s signal depends on landscape response and signal transfer times. Systems with long transfer zones (300-1000 km) have slower response times, leading to buffered or out-of-phase output signals (Romans et al., 2016; Castellort et al., 2023). In the Shetland Platform, relatively short catchments and transfer zones (<300 km total lengths) point to fast landscape response and signal transfer times, with increased potential for preservation of a fast signal induced by the PETM. However, results from **papers 1 and 2** suggest that concurrent uplift overshadowed any contributions of the hyperthermal, both in terms of whole-system Q_s signals (**Paper 2**) and as a modulator of shelf reworking and fan deposition (**Paper 1**).

Overall implications for geodynamic models

When combined, the overall tectono-stratigraphic and paleogeographic observations of **papers 1, 2 and 3** and the sediment mass-budget of **Paper 2** all offer insights for geodynamic evolution. First, the absence of clear indicators for shallowing and subaerial exposure before the Late Thanetian in East Shetland can be highlighted, something that is dealt with in **Papers 1 and 2**. Several inferences have previously been made on unconformity development during the Danian and Selandian (often based on biostratigraphy or analysis of well-logs, Mudge and Jones, 2004), and these unconformities have also been linked with uplift (Mudge and Jones, 2004; Mudge, 2015; Conway-Jones and White, 2022). Results from **Paper 1** suggest that if these unconformities are present in East Shetland, they may be developed in a submarine environment, meaning that any ties to uplift are less straightforward. On the other hand, results from **Paper 2** indicate a gradual increase in maximum relief from the Danian onwards, broadly matching hypotheses for the timing of initial dynamic uplift in the Shetland Platform (Ahmadi et al., 2003; Figure 15 in **Paper 2**). Tying all of these

observations, the **Mid-Paleocene** uplift event is proposed to be more restricted to the west and central Shetland platforms, implying a W-E or NW-SE tilted topography at the time (Figure 15 in **Paper 2**). This is similar to the reconstructed Paleocene topography showcased in Smallwood (2008), constrained by estimates of denudation and dynamic topography (Figure 5.3). However, it is unclear how this is related to potential uplift in other areas in the North Atlantic with similar stratigraphic records like Norway (Sømme et al., 2019, 2023) and how exactly tilting would be expressed there. Additionally, while the timing of this Mid-Paleocene event broadly fits the modelled dynamic topography in Barnett-Moore et al. (2017), it is unknown whether real dynamic topography could have operated at the same temporal and spatial scales invoked by the Mid-Paleocene uplift event.

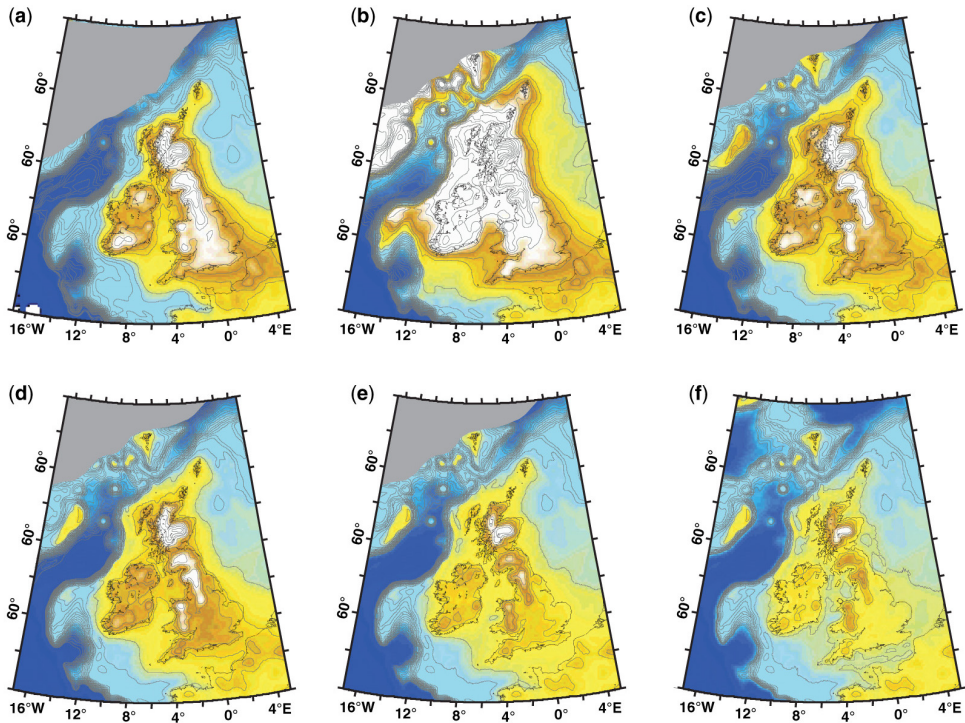


Figure 5.3 – Topographic reconstruction of the British Isles and Faroes during the Cenozoic, using igneous underplating, dynamic support and erosion as function of elevation and erodibility (Smallwood, 2008). A) 65 Ma, earliest Danian; B) 57 Ma, latest Thanetian; C) 50 Ma, Ypresian; D) 40 Ma; E) 20 Ma; F) 1Ma.

After Mid-Paleocene tilting, uplift appears to have resumed just before the Paleocene-Eocene transition, causing the D2-D3 relative sea-level fall event (at ~56.6 Ma? in the Late Thanetian). This appears to have been a broad event over the Shetland Platform, although it is better preserved and distinguished in the East, as discussed in **papers 2 and 3**. Another uplift event occurred in the Early Ypresian, corresponding to the Flett Unconformity in the West Shetland area (Jolley et al., 2021; Walker et al., 2022b). In East Shetland, it can be further distinguished in a syn-PETM (D4) and a post-PETM (D5) pulse, the former being restricted to the southern ESP and the latter being restricted to the Beryl Embayment area. Both events appear to be less significant than Early Ypresian uplift in the West. Overall, this points to spatially variable uplift in both the W-E and N-S directions, but suggests synchronous uplift across West and East.

As explained briefly in **Paper 3**, fast and spatially varied uplift in the Late Thanetian – Early Ypresian is not consistent with traditional models of dynamic topography involving long-wavelength and long-period fluctuations (*e.g.* Hager et al., 1985; Barnett-Moore et al., 2017). The marked coincidence between crustal features and the patterns of spatially varied uplift suggests some involvement of lithospheric or crustal features. If uplift is induced by vertical flow in the mantle, perhaps the interaction between mantle and a rheologically heterogeneous lithosphere could allow for short-wavelength and short-period variations in vertical motions (Figure 5.4; Burov and Gerya, 2014). Furthermore, perhaps even a primary long-wavelength dynamic uplift signal could be affected by differential isostatic responses and rebounds, resulting in short-wavelength variations consistent with the crustal configuration of East Shetland.

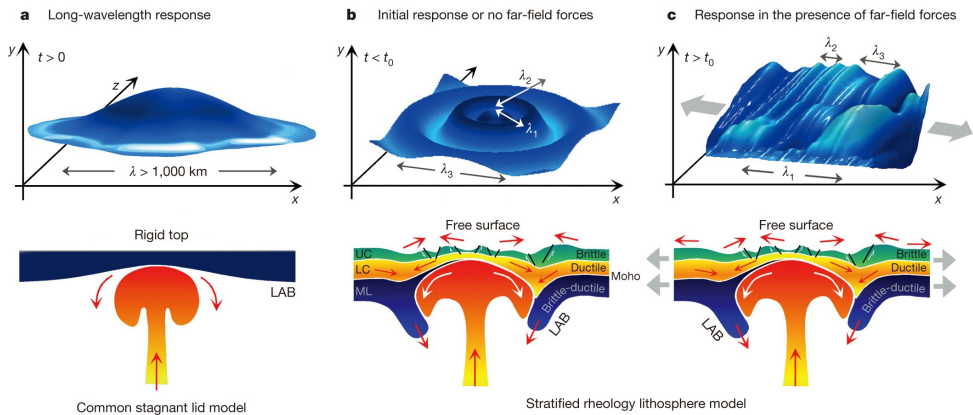


Figure 5.4 - Conceptual model of dynamic topography and mantle-lithosphere interaction (Burov and Gerya, 2014). A) Traditional model of long-wavelength response with a rheologically homogeneous lithosphere (stagnant lid model); B) Complex geometries and wavelengths (λ) generated by the inclusion of the free-surface response of a rheologically stratified lithosphere. C) Asymmetric topography created by inclusion of far-field tectonic stresses in case B, shortly after t_0 . Notice the presence of short-wavelength ridges and basins in the upper crust. The location of these will be influenced by pre-existing structural features. LAB – lithosphere-asthenosphere boundary; LC – lower crust; ML – middle crust; UC – upper crust.

Future work

3D sequence stratigraphy

The topic of along-strike or 3D variability inside sequence stratigraphy has received significant attention in recent years (review in Zecchin and Catuneanu, 2020), although arguably few attempts have been made to propose formal methodologies to tackle it. In short, along-strike variability offers significant additional complexity for the interpretation of sequence stratigraphic frameworks, mainly because the identification and correlation between variably expressed, amalgamated, or omitted surfaces is often challenging (Figure 5.5). The extent of how relevant this is to a particular analysis depends on the scales and resolution of observation, with complexity tending to increase with higher resolutions (e.g. Zuchuat et al., 2023).

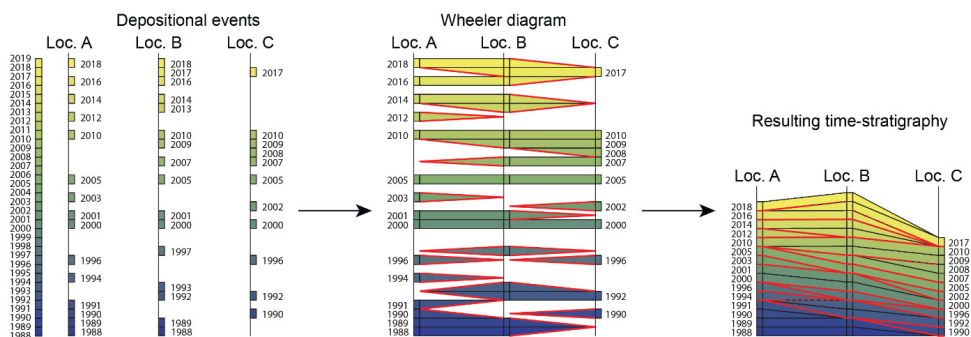


Figure 5.5 – Diagram showing the time-stratigraphic record of different locations along the coast of the Gulf of Carpentaria, Northern Australia, from 1988 to 2018 (Zuchuat et al., 2023). The pseudologs on the left represent depositional events in three different locations, with yearly resolution. Erosion events and hiatuses have varied expression along-strike, and result in complex amalgamation of erosive surfaces, shown in the Wheeler diagram. Only some events can be traced in all three locations, and all hiatuses are strongly diachronic.

Constraints on along-strike variability of depositional systems and seismic-scale sequence stratigraphy have been the focus of studies carried out in offshore Australia (Paumard et al., 2019; Shepherd et al., 2023) and the Santos Basin (Steventon et al., 2020), which are similar in scope and approach to **Papers 1 and 3**. In offshore Australia, the applicability of automated 3D interpretation for seismic stratigraphic analysis has been shown to be significant (Paumard et al., 2019). Automated interpretation essentially consists in AI-assisted ultra-high-resolution seismic interpretation, where all seismic reflectors inside a studied succession of interest are interpreted in a continuous 3D volume. Automation allows the faster extraction of a larger number of seismic stratigraphic surfaces in 3D, increasing the temporal resolution of seismic attribute maps and seismic geomorphological constraints. The extraction of multiple Wheeler chronostratigraphic diagrams using seismic data is also made easier by automated processes (Figure 5.6; Guillon et al., 2013; Sylvester et al., 2024), and can reduce any model-driven bias for picking stratigraphic surfaces.

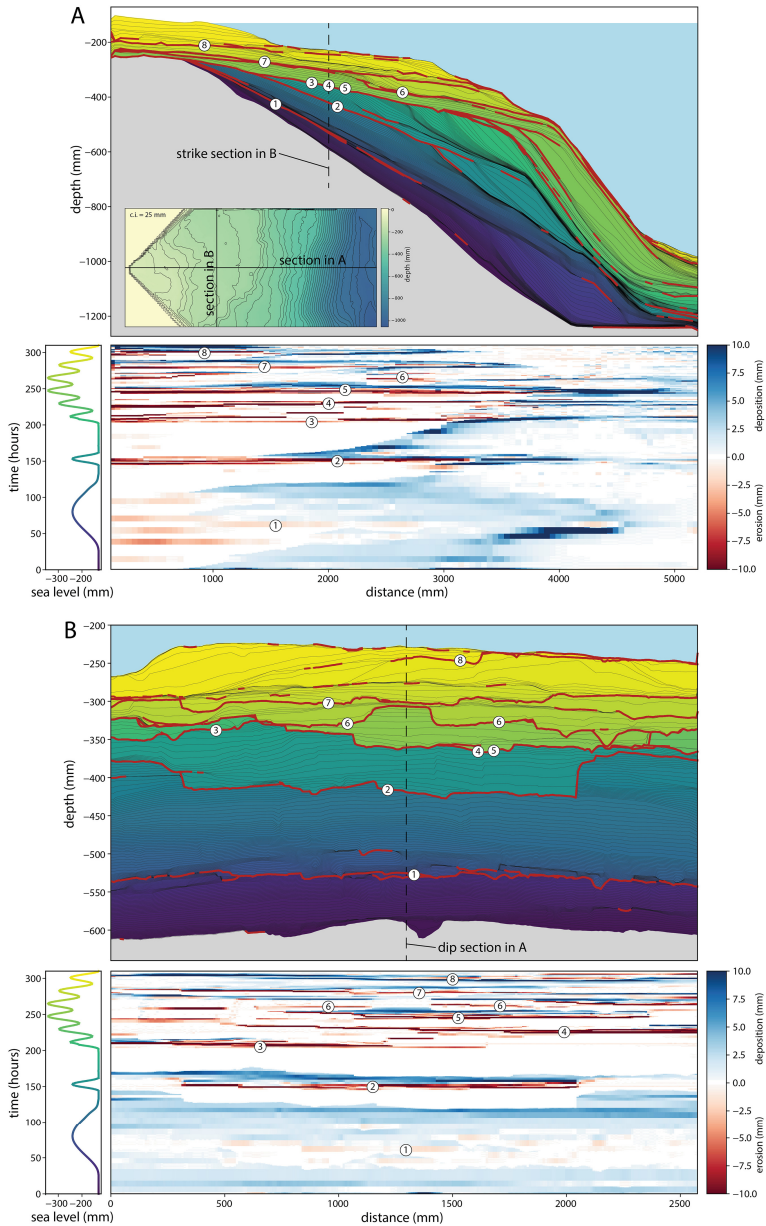


Figure 5.6 – Automated chronostratigraphic diagrams from Sylvester et al. (2024) made from flume-tank experiment (Martin et al., 2009). A) Dip section close to the apex of the “shelf”, colors represent time. In the Wheeler diagram, red colors represent erosive events (unconformities), while blue represent deposition and white represents non-deposition or bypass. Main unconformities are numbered. B) Cross section of the same succession. Notice the variable extent of unconformities and their time-transgressive character along-strike.

However, in the East Shetland Platform, the elevated complexity in seismic imaging caused by soft-sediment deformation complicates automated interpretation. **Paper 1** aims to obtain constraints that are similar to those in Paumard et al. (2019) using traditional, manual seismic interpretation. This is evidently a very time-consuming process for such very large areas. The Wheeler diagrams shown in Figures 8-12 in **Paper 1** were devised with a similar “horizon-flattening” approach as in automated interpretation, resulting in data-driven (albeit complex) depositional geometries.

The stratigraphic framework proposed in **Paper 1** can benefit from further testing and experiments using automated interpretation, but this would likely require significant preparation of the seismic data for this goal. Automated interpretation may be used to independently test the proposed stratigraphic framework, while also providing a framework to investigate the effects of human bias during interpretation. Additionally, higher-resolution seismic horizon extraction should allow even greater insights based on seismic attribute analysis.

Furthermore, it would be interesting to use a quantitative approach to elaborate on the qualitative observations of **Paper 1** on along-strike variability (in addition to clinoform rollover trajectories and RSL curves). This could be done by quantifying supply and accommodation along the main sediment entry points and associated sediment routing paths, including corresponding submarine fans downdip. This would allow a greater understanding of the exact roles of laterally varied supply and accommodation on the generation of sequence stratigraphic responses.

Another quantitative aspect that could be extracted from this dataset is the geomorphological information contained in buried erosional surfaces, namely the West Bressay drainage described in **Paper 1**. This erosional surface could be used for the analysis of uplift-related knickpoints, as in Hartley et al. (2011) and Stucky de Quay et al. (2017), or perhaps for other constraints on geomorphic processes operating during erosion (Figure 5.7). Preliminary processing of this surface suggests that deformation associated with sediment remobilization and differential compaction can pose an additional challenge for the extraction of relevant geomorphic data, but this may also

provide a framework for comparison with other drainage networks examined in the area.

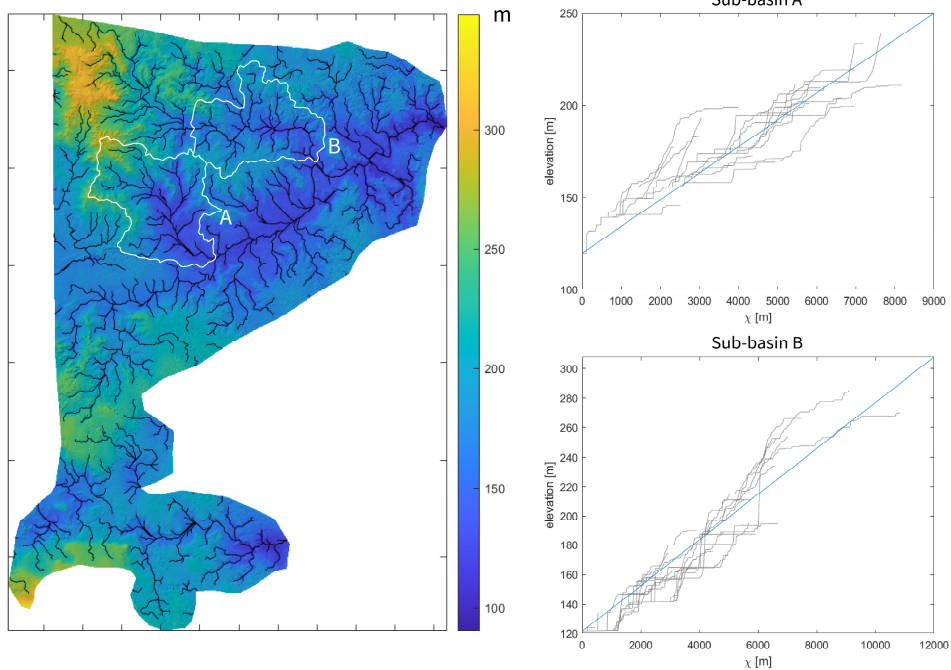


Figure 5.7 – A) Drainage reconstruction of the 3D backstripped West Bressay catchment. B) Elevation against χ plots of two drainage subbasins. χ refers to the integral function of position with normalization by drainage area in a channel network (Perron and Royden, 2013). These “chi-plots” can be used to investigate the state of bedrock incision and proximity to steady-state river profiles (Willett et al., 2014).

Forward modelling of the stratigraphy and sediment flux signals around the Shetland Platform

Last but not least, it should be highlighted that all of the observations obtained here can be used to calibrate forward models of the stratigraphy around the Shetland Platform. Previous forward modelling attempts have successfully employed low resolution constraints on the regional paleogeography and sediment fluxes to evaluate the feasibility of the dynamic uplift hypothesis (Jones and White, 2003; Smallwood, 2008, Figure 5.3). It is likely that similar studies can benefit considerably from the higher resolution observations here. A forward modelling framework could attempt to recreate: sediment fluxes measured in **Paper 2**; the broad stratigraphic framework and

paleogeography described in **papers 1 and 2**; the vertical motions described in **papers 2 and 3** (relief values in Paper 2).

Forward modelling could employ an accurate recreation of the topography in the Shetland Platform to test exact constraints on catchment parameters and stratigraphy obtained here (*e.g.* Jones and White, 2003; Smallwood, 2008). Alternatively, a modelling study could use a simplified model meant to represent the boundary conditions in the Shetland Platform to test generalized assumptions (*e.g.* Ding et al., 2019), such as the superimposition of uplift and the PETM during the Paleocene-Eocene transition. Regardless of format, the advantage of a coupled landscape evolution + sediment deposition model would be a more intricate assessment of tectonic and climatic signals than the LRoART mass-budget, for example. Additionally, a coupled thermomechanical + surface processes model (*e.g.* Wolf et al., 2022; Pichel et al., 2023) could be used to investigate if and how different geodynamic processes such as dynamic uplift and rifting could be distinguished in the Shetland Platform based on their corresponding signals, an important standing question that few studies have attempted to tackle.

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Appendix

Appendix A: Stratigraphic expression of the Paleocene-Eocene Thermal Maximum climate event during long-lived transient uplift—An example from a shallow to deep-marine clastic system in the Norwegian Sea.

Appendix B: Supplementary material for Paper 1.

Appendix C: Supplementary material for Paper 2.

Appendix D: Supplementary material for Paper 3.

Appendix A: Stratigraphic expression of the Paleocene-Eocene Thermal Maximum climate event during long-lived transient uplift—An example from a shallow to deep-marine clastic system in the Norwegian Sea.

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Stratigraphic expression of the Paleocene-Eocene Thermal Maximum climate event during long-lived transient uplift—An example from a shallow to deep-marine clastic system in the Norwegian Sea

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Seismic geomorphology and stratigraphic analysis can reveal how source-to-sink systems dynamically respond to climatic and tectonic forcing. This study uses seismic reflection data from the Norwegian Sea to investigate the stratigraphic response to a short-lived (0.2 Myr) period of climate change during the Paleocene-Eocene Thermal Maximum (PETM), superimposed on a long-lived (~8 Myr) period of hinterland uplift. The data show that long-term uplift resulted in ~300 m of relative sea-level fall, forced regression and formation of incised valleys during the latest Paleocene-earliest Eocene. The short-lived PETM climate perturbation at ~56 Ma changed the transport dynamics of the system, allowing sediment to be bypassed to wide channel complexes on the basin floor, feeding a large mud-rich basin-floor fan more than 50 km into the basin. Our analysis also suggest that sediment supply was up to four times higher during the PETM compared to earlier and later periods. Maximum regression at ~55.5 Ma resulted in the formation of a subaerial unconformity. The style of subaerial incision was dictated by shelf accommodation and proximity to the area of direct sediment input. Out-of-grade shelves and slopes sourced by littoral drift were prone to incision, but directed and graded shelves and slopes were not. Despite maximum regression, sediments were not transported significantly beyond the toe-of-slope aprons, suggesting that rapid climate change was more efficient in bypassing sediment to the deep-water than low stands of sea level. As long-term accommodation increased after the PETM, deltas were still able to reach shelf edge, but periods of maximum regression were not associated with deep incisions along the outer shelf and only smaller canyons and gullies formed. The shelf-slope wedge was finally transgressed at ~51 Ma. The age of deep valley incisions overlaps with the time of subaerial erosion in the East Shetland and Faroe-Shetland basins, suggesting a common mechanism for North Atlantic uplift around 55–56 Ma. Other seismic stratigraphic surfaces do not seem to be regionally time-equivalent, highlighting the importance of local controls on internal architecture of shelf-slope wedges. This study demonstrates the high-resolution stratigraphic response to long- and short-term external forcing together with intrinsic processes and can help identify similar relationships in other areas.

KEYWORDS

climate signal, PETM, sediment routing, shallow and deep-marine clastics, seismic geomorphology

Introduction

The stratigraphic record along continental margins captures the long-term interplay between allogenic tectonic and climatic forcing, autogenic dynamics, and environmental processes acting upon the sedimentary routing system. The environmental signals that result from such allogenic forcings or controls are expressed by changes in erosion, transport and deposition of sediment along the system (Tofelde et al., 2021). Although the 'ideal' stratigraphic response to allogenic forcing can be investigated in numerical stratigraphic models or flume tank experiments where boundary conditions are well known (Jerolmack and Paola, 2010; Armitage et al., 2011; Straub, 2019), natural systems are often more difficult to decipher. However, several studies have documented Milankovitch-driven climate signals in Pliocene-Pleistocene deep-water systems where the impact of tectonic processes often is small (Goodbred, 2003; Ducassou et al., 2009; Gong et al., 2018; Cullen et al., 2021). But separating short-lived climate signals from more long-lived tectonic signals in ancient greenhouse systems can be more challenging since the configuration of the original sediment routing system changes with time and because signals tend to overlap in time and space and become difficult to distinguish from more local environmental and autogenic processes (Muto et al., 2007; Hajek and Straub, 2017; Blum et al., 2018). In addition, the role of shallow-marine systems as conveyers of environmental signals to deep-marine basins may have been different during greenhouse times since shelf configurations and the amplitude and frequency of climatic fluctuations have varied throughout the Phanerozoic (e.g., Sømme et al., 2009a; Harris et al., 2020; Burgess et al., 2022).

Seismic stratigraphy and seismic geomorphological have the potential to image the interaction between signals resulting from allogenic forcing and more local autogenic and environmental processes (e.g., Bullimore et al., 2005; Ryan et al., 2009; Paumard et al., 2020; Posamentier et al., 2022). The strength of three-dimensional (3D) seismic geomorphology is that it provides: 1) a regional view of the strike and dip response of the routing system, 2) a linkage between sedimentary processes and stratigraphic architecture, and 3) high temporal and spatial resolution, which allows snapshots of closely spaced geological events to be imaged. When geomorphology from extensive seismic reflection datasets is combined with lithological and chronostratigraphic data from wells and two-dimensional (2D) seismic cross-sections, showing changes in stratigraphic stacking pattern and internal architecture, a three-dimensional image of the dynamic response to external forcing can be achieved.

The North Atlantic margin provides an ideal opportunity to apply this approach, as it experienced overlapping climatic and tectonic perturbations during the Paleocene and early Eocene. From ~65 Ma, the region experienced uplift on the order of hundreds of meters, commonly linked to dynamic topography associated with the Iceland mantle plume (e.g., Nadin and Kusznir, 1995; White and Lovell, 1997; Skogseid et al., 2000; Hartley et al., 2011; Stucky de Quay et al., 2017). The consequence of this regional uplift was a significant increase in sediment supply to the North Atlantic margins, manifested as progradation of thick shallow to deep-marine clastic wedges (e.g., Hartog Jager et al., 1993; Liu and Galloway, 1997; Martinsen et al.,

1999; Henriksen et al., 2005; Hovikoski et al., 2021). At the same time, expulsion of methane and carbon dioxide prior to North Atlantic break-up is believed to have been the main driver for the PETM (Svensen et al., 2004; Kjoberg et al., 2017), which was a short-lived (100–200 kyr) climate perturbation (Sluijs et al., 2007a; Li et al., 2022) superimposed on the longer term marginal uplift. Studies of the PETM suggest that this hypothermal event was associated with increased onshore weathering rates, higher seasonal variability in fluvial discharge and change in sediment delivery to shallow and deep-water basins (Pujalte et al., 2015; Carmichael et al., 2017; Rush et al., 2021; Jin et al., 2022), producing a stratigraphic response that can be distinguished from more long-term tectonic forcing (Samanta et al., 2016). Even though the PETM has been recognized several places in wells and outcrops in the North Atlantic (e.g., Beerling and Jolley, 1998; Dypvik et al., 2011; Harding et al., 2011; Kender et al., 2012; Eldrett et al., 2014; Stokke et al., 2021; Jolley et al., 2022), little is known about how the PETM influenced the dynamics of sedimentary systems in this region. However, recent studies from the North Sea suggest increased sediment supply and a change in depositional style during the event (Sømme et al., 2019; Vieira and Jolley, 2020; Jin et al., 2022).

This study aims to investigate the seismic architecture and seismic geomorphology of a ~8 Myr shallow to deep-marine clastic system in the Froan Basin on the eastern margin of the Norwegian Sea (Figure 1). The intention was to analyze the stratigraphic response of long-term uplift and short-lived climate change in shelf, slope and deep-water settings in a source-to-sink perspective. Individual seismic stratigraphic units, their well-log response, seismic facies, and seismic geomorphology were described and linked to a robust chronological framework. An analysis of the sequence stratigraphic and geomorphological characteristics was performed, and the dynamic changes of the sediment routing system were captured. This dataset represents the only known preserved record of thick Paleocene-early Eocene shallow-marine and shelf sediments on the Norwegian continental shelf and the first documentation of the shallow to deep-marine response to PETM climate change using seismic geomorphology. The stratigraphic response observed here will act as a valuable reference point for other areas that experienced similar changes in forcing conditions in the North Atlantic or elsewhere. Improved knowledge of system response to natural climate fluctuations can also help to understand how anthropogenic influence and global warming may affect active sedimentary systems in the future.

The Froan Basin

As a part of the North Atlantic rifted margin (Doré et al., 1999), the Froan Basin and the greater Norwegian Sea (Figure 1) experienced three phases of rifting. The first phase took place in the Late Permian to Early Triassic and was associated with the formation of local rift basins filled by continental and marginal-marine sediments (Blystad et al., 1995; Bunkholt et al., 2021). The second phase occurred during the Middle Jurassic to earliest Cretaceous, resulting in the formation of deep rift basins, hyperextension and local mantle exhumation along

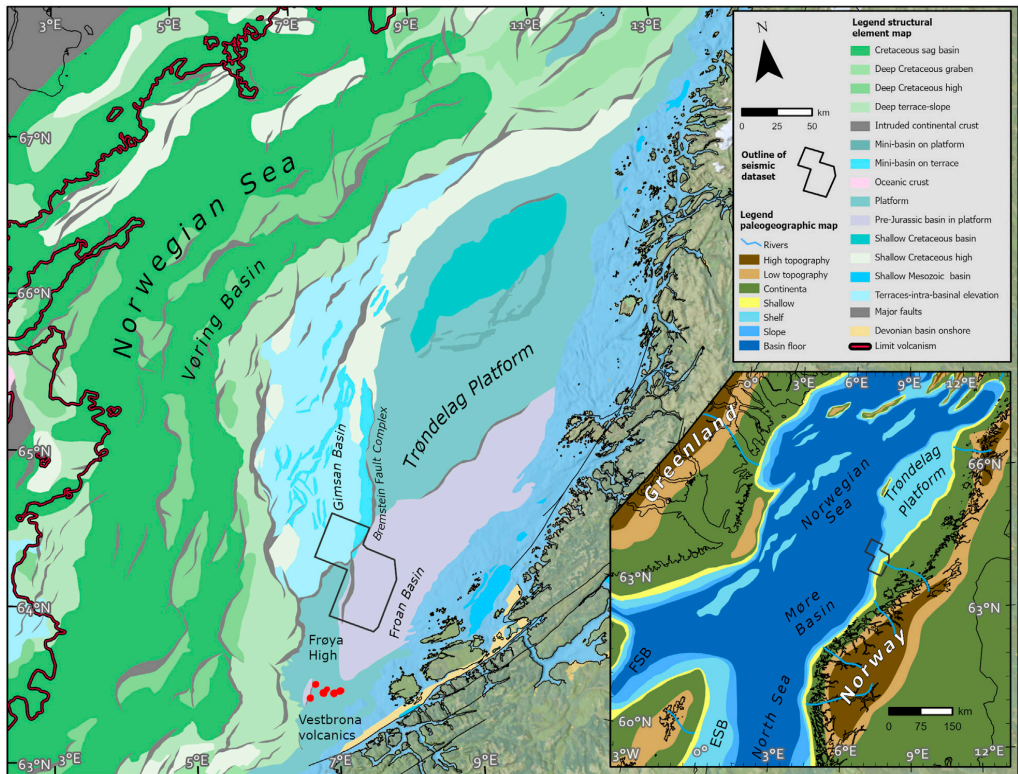


FIGURE 1
Location of the study area in the Froan Basin. Inset map show the paleogeographic situation for the Norwegian Sea and northern North Sea prior to break-up at ~55 Ma. Structural elements from [Gernigon et al. \(2021\)](#). FSB = Faroe-Shetland Basin, ESB = East Shetland Basin.

the central part of the basin ([Osmundsen et al., 2017](#)). These rift basins were filled by shallow and deep-marine clastics derived from local fault blocks and the surrounding hinterland on the Greenland and Norwegian margins (e.g., [Lien, 2005](#); [Martinsen, 2005](#)). The third and last rift phase commenced in the middle Campanian and continued until continental break-up and seafloor spreading around 55 Ma (e.g., [Eldholm et al., 1989](#); [Færseth and Lien, 2002](#); [Zastrow et al., 2020](#)). Most of the deformation during this last phase of extension was centered along the axis of future break-up ([Skogseid et al., 2000](#)), and there is little evidence of rift-related (footwall) uplift of the hinterland east of the rift. Long-lived extension and crustal thinning resulted in thermal subsidence of the Froan Basin and the surrounding Trøndelag Platform, and the area was expressed by a low-relief basin tilted toward the west in the latest Cretaceous to earliest Paleocene.

As in many other North Atlantic regions (e.g., [Jolley and Whitham, 2004](#); [Ryan et al., 2009](#); [Mudge, 2014](#); [Hovikoski et al., 2021](#)), the base Paleogene unconformity marks a shift from overall transgression to regression along the Norwegian margin ([Martinsen et al., 1999](#); [Gjelberg et al., 2001](#); [Henriksen et al., 2005](#); [Dmitrieva et al., 2017](#); [Sømme et al., 2019](#)). The angular unconformity and the

abrupt increase in sediment supply observed in the Paleocene succession are often attributed to dynamic topography and the Icelandic mantle plume (e.g., [Dam et al., 1998](#); [Petersen, 2019](#); [Sømme et al., 2019](#)), but other mechanisms such as rift-related deformation and thermal effects related to pulses of volcanism could also have played a role, especially in the vicinity of the outer margin ([Guarnieri, 2015](#); [Stoker et al., 2017](#); [Jolley et al., 2021](#)).

High sediment supply and progradation in the North Atlantic was followed by a phase of transgression and backstepping around 55–54 Ma ([Mudge and Jones, 2004](#); [Henriksen et al., 2005](#); [Hardman et al., 2018](#); [Sømme et al., 2019](#)). This transgression occurred just after sea-floor spreading at 55 Ma, and regional subsidence following break-up and deflation of the Iceland plume has been suggested to be a driving mechanism for this relative sea-level rise ([Nadin and Kuszniir, 1995](#); [Hardman et al., 2018](#)).

Prior to the PETM, the Paleocene Froan Basin experienced relative cooling and overall sea-level fall after the warm and humid Late Cretaceous period ([Miller et al., 2020](#)). The Froan Basin itself was situated between the relatively steep and high-relief Møre Basin to the south and the low-relief Trøndelag Platform in the North. The proximal part of the marine basin was characterized by a ramp-like

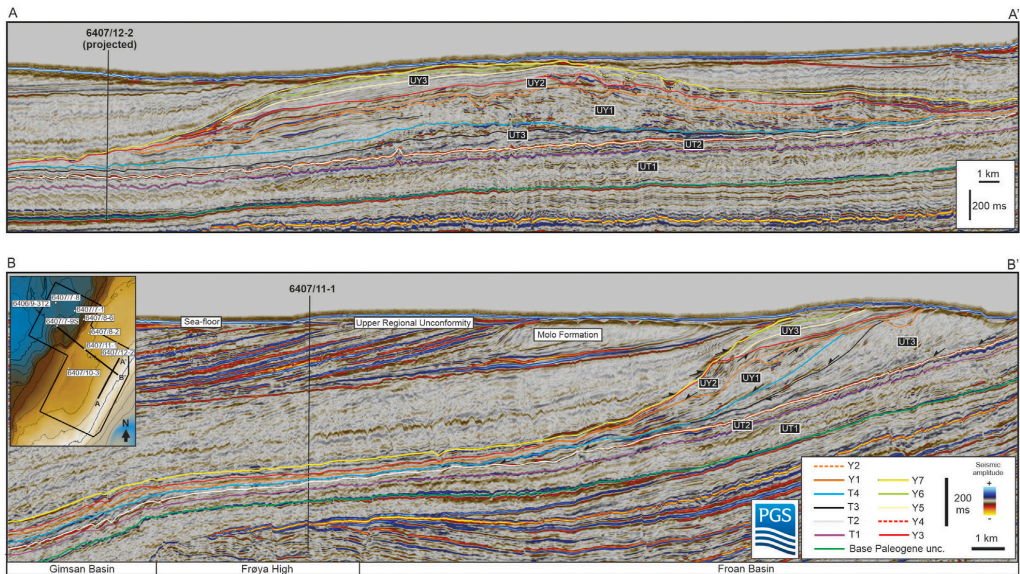


FIGURE 2
 (A) Strike- and (B) dip-line across the study area covering the main Paleocene to Eocene depocenter. Units UT1-UY3 are defined by bounding surfaces expressed by downlap, onlap, toplap, and truncation relationships.

morphology, bounded to the west by the Frøya High and the Bremstein Fault Complex, which created a steeper slope gradient into the deeper Gimsan Basin on the down-faulted Halten Terrace (Figure 2). Rivers feeding the Paleocene shoreline in the Froan Basin are believed to have had catchment areas on the scale of 2×10^4 km² and the full source-to-sink length was approximately 150 km (Figure 1) (Sømme et al., 2009b).

Data and methodology

High-resolution 3D seismic reflection datasets provide an unparalleled opportunity to analyze sediment routing systems in time and space (Posamentier and Kolla, 2003; Posamentier, 2005). Initially these data sets covered only smaller portions of sedimentary basins, while in recent years, larger datasets and regional megamerages >1,000 km² have become the norm in many areas, not least offshore Norway (e.g., Keller et al., 2022). The seismic reflection data utilized in this study comprise a ~2,700 km² subset of PGS18M05 and PGS17M05. Both datasets have a line spacing of 12.5 × 12.5 m. The dominant frequency is ~25 Hz and the velocity of the studied succession is ~2000 m/s, which gives a vertical resolution of ~20 m. The data are zero phase and an increase in acoustic impedance is shown as a (blue) peak (Figure 2).

Seismic well-tie was based on time-depth relationships from well 6407/12-2 (Figure 3). Well 6407/12-2 penetrates the toe-of-slope succession of the Paleocene wedge and has a full suite of electrical logs and biostratigraphic coverage. Palynological data from this well have been linked to the NW Europe dinoflagellate zonation scheme from Speijer et al. (2020). In addition, electrical logs and palynological

data from several other wells in the area have been used to correlate and describe the lithology and log-response in the distal part of the Froan and Gimsan basins.

No carbon isotope data are available in this study and the recognition of the PETM in well data is based on the occurrence of the dinocyst species *Apectodinium augustum* in palynological data (Sluijs et al., 2007b). A global review by Denison (2020) shows that the main body of the Carbon Isotope Excursion (CIE) defining the PETM is coeval with the acme of the genus *Apectodinium*. Recent work from the North Sea (Jin et al., 2022) and from Svalbard (Harding et al., 2011) also confirms that the onset, the main body and the recovery of the PETM CIE are reflected by the initial presence, acme and last occurrence of *A. augustum*.

Seismic stratigraphic surfaces (BPU-Y7; Figure 2) were mapped based on stratal terminations (toplap, downlap, onlap and truncation) (Mitchum et al., 1977), and were used to define the top and base of six seismic units (UT1-UY3). Changes in accommodation (A) and sediment supply (S), described as $\delta A / \delta S$, were investigated using shelf-break trajectory trends (Helland-Hansen and Martinsen, 1996), and geometrical descriptors of Aali et al. (2021). Here the geometries are described based on the shelf-break trajectory trend (down-stepping, stationary, up-stepping) and the landward and seaward termination styles (fore-stepping or back-stepping). Transgression occurs when the rate of accommodation creation is higher than the rate of sediment supply ($\delta A > \delta S$), normal regression occurs when sediment supply is higher than the rate of accommodation creation ($\delta A < \delta S$), and forced regression occurs when falling base level ($A < 0$) together with sediment supply forces the shoreline basinward (Posamentier et al., 1992; Catuneanu, 2002).

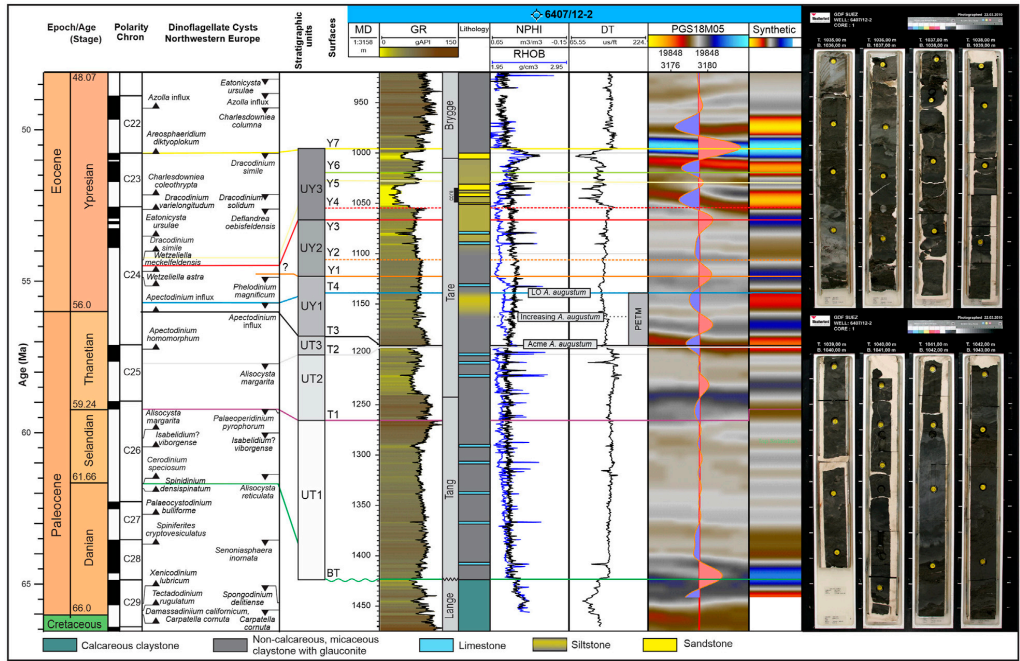


FIGURE 3 Seismic well tie for 6407/12-2 and dinoflagellate zonation in NW Europe (from Speijer et al., 2020). The PETM is defined by the presence of *A. augustum*. Position of core photos between 1,035 and 1,045 m are indicated on the lithology column. Core photos from www.npd.no.

To preserve original shelf-edge trajectories and to quantify relative sea-level fluctuations, sequential backstripping was conducted following the methodology of Beelen et al. (2019), where individual clinotherm units were backstripped, decompacted and adjusted for local and flexural subsidence. Lithology distribution and porosity-depth trends (for foresets and bottomsets) were derived from well 6407/12-2. Since topsets have not been penetrated by wells, lithology distribution was inferred from seismic facies and seismic geomorphology and assumed to comprise 70% sandstone. Thicknesses and slope gradients reported in this study are values after decompaction.

Seismic geomorphology of each seismic unit was extracted with two different attributes: i) a co-blend of Root Mean Square (RMS) and Variance (edge effect), which is useful for highlighting lateral changes in impedance (proxy for lithology), faults and channels etc.; and ii) frequency decomposition (FDC) and RGB color blending, which are useful for imaging subtle and lateral changes in depositional systems. For the RGB color blending, frequency bands of 15, 30 and 45 Hz were used.

Depositional environments present in the prograding wedge vary from coastal to deep-marine. We do not attempt a full hierarchical description of depositional elements (see Cullis et al. (2019) for a recent overview), but apply general terms linking scale and depositional setting. The term ‘channel’ is used to describe single-cycle features of erosion or deposition at the lower scale of seismic resolution (tens of meters wide) and ‘channel complex’ to describe up

to 2 km wide erosional or depositional features on the slope and basin floor. ‘Lobe’ is used for single, sheet-like depositional features in front of channels, canyons and/or incised valleys, and ‘toe-of-slope apron’ is used for stacking and coalescing of channels and lobes near the advancing shelf margin. ‘Basin-floor fan’ is reserved for depositional features larger than lobes deposited tens of kilometers into the deep-water basin.

Results

Stratigraphic units

We split the Paleocene to lower Eocene succession in the Froan Basin into six units (Figure 4): three in the Selandian to Thanetian (Units T1-T3) and three in the Ypresian (Units Y1-Y3). The depositional setting and evolution of these units are described and interpreted in the following section using thickness and seismic attribute (RMS) maps (Figure 5) together with well log data (Figure 3, Figure 6). The seismic geomorphology is described from frequency decomposition (FDC) maps in Figure 7.

Unit T1—Selandian-lowermost Thanetian (BPU to T1)

Unit T1 (UT1) is bounded below by the base Paleogene unconformity (BPU) and above by surface T1 (Figure 4). A Selandian age of this unit is constrained by the presence of

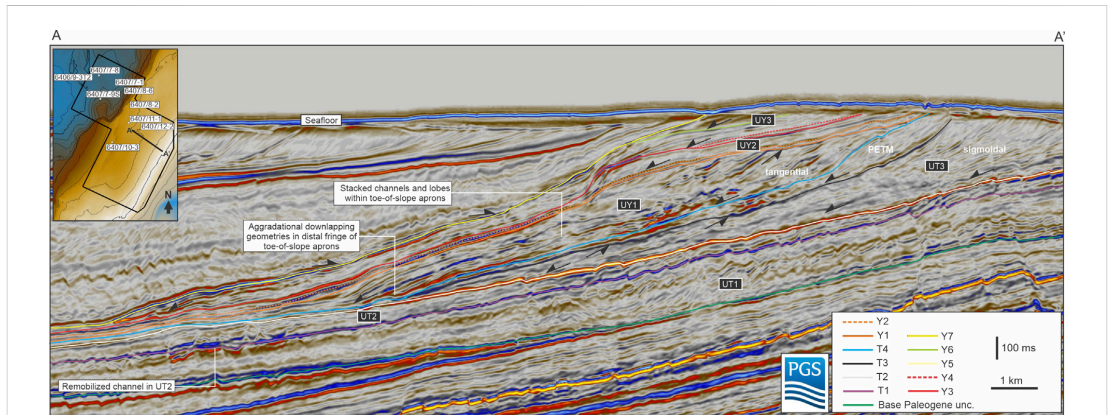


FIGURE 4

Key units and surfaces in the proximal part of the shelf-slope wedge with stratal terminations. First phase of progradation is defined by sigmoidal clinoforms. Main phase of progradation of UY1 is characterized by tangential clinoforms. The PETM is represented by the stratigraphic unit between surface T3 and T4.

Cerodinium speciosum in the lower part and the last occurrence of the dinoflagellate *Palaeoperidinium pyrophorum* at the top of the unit (Speijer et al., 2020) (Figure 3). In seismic cross section (Figure 4) and in RMS map view (Figure 5), the unit is locally characterized by low amplitude reflections and high-density faulting with a polygonal expression. In well logs, the unit is marked by higher gamma ray (GR) values compared to the Upper Cretaceous unit below and the unit above. Internally, the unit shows increasing GR values upward. The lithology also changes from carbonaceous claystones in the Cretaceous below, to mica-rich claystones with stringers of limestone in the Paleocene above. The thickness map shows a distinct north-south trend, reaching up to ~200 m in the southern Froan Basin (Figure 5).

Interpretation

Selandian strata are interpreted to represent mudstones deposited in a deep-marine setting. The thickness trend reflects fault-related subsidence east of the Frøya High. The angular character of the BPU and the hiatus to Cretaceous sediments below (e.g., in well 6407/12-2) suggest a major unconformity at the base of the unit. At the time of deposition of UT1, the shoreline was located farther to the east and the deposits are thus interpreted to represent the distal equivalent to Selandian progradation further landward (later eroded by the URU (upper regional unconformity) and at the present-day seafloor) (Figure 4).

Unit T2—Lower Thanetian (T1 to T2)

Unit T2 (UT2) is bounded below by surface T1 and above by surface T2 (Figure 4). An early to middle Thanetian age of this unit is suggested by the last occurrence of *Alisocysta margarita* and *Areoligera gippingensis* at the top of the unit (Speijer et al., 2020; Vieira et al., 2020) (Figure 3). The unit is dominated by moderate to low amplitude reflections characterized by high-density faulting in the western part of the study area (Figure 4). The fault pattern appears polygonal in the RMS and FDC attribute maps (Figures 5, Figure 7A). In the eastern part of the study area, the RMS, thickness and FDC maps show channel complexes and lobe-like features; however, the easternmost part of the system is not visible due to bypass or

subsequent erosion by the overlying unit. In the area between 6407/10-3 and 6407/11-1, a well-defined channel complex and terminal lobe is observed tapering out towards the Frøya Fault in the west (Figure 5). In seismic cross section, the channel complexes are locally associated with differential compaction (Figure 4, Figure 7A, section A), and locally also with wing-like intrusions in the upper part. A fault-controlled depocenter to the east of the Frøya Fault is suggested from the thickness map. In well logs, the unit generally shows an upward decrease in GR values and the lithology is dominated by claystones (Figure 6), but in 6407/10-3 (Figure 5), a 11 m thick sandstone unit is present.

Interpretation

UT2 is interpreted to represent the first input of sandy gravity flows to the Froan Basin during the Paleocene. The feeder system has a SE-NW orientation indicating sediment supply from the hinterland to the east, but no time-equivalent shoreline or shelf deposits are preserved. A common feature with this unit is that the channel complexes and lobes are strongly deformed and partly injected into the overlying sediments, resulting in the chaotic expression on the seismic attribute maps and in cross sections. A similar expression of Paleocene sands also occurs farther south in the North Sea (Dixon et al., 1995; Dmitrieva et al., 2017; Sømme et al., 2019). This deformation appears to have occurred soon after deposition, because there is no evidence of significant injection into overlying stratigraphic units and because later progradation was partly affected by the sea-floor topography caused by the injectites.

Unit T3—Uppermost Thanetian (T2 to T3)

Unit T3 (UT3) is bounded below by surface T2 and above by surface T3 (Figure 4). The upper T3 surface correlates with the influx of the dinoflagellate cyst *Apectodinium augustum*, suggesting a latest Thanetian (pre-PETM) age of this unit (Speijer et al., 2020; Vieira et al., 2020) (Figure 3). In seismic, the basal T2 is a regional downlap surface (Figures 2, 4). The first clinoforms to downlap the T2 surface in the easternmost part of the study area are 300–400 m thick and show

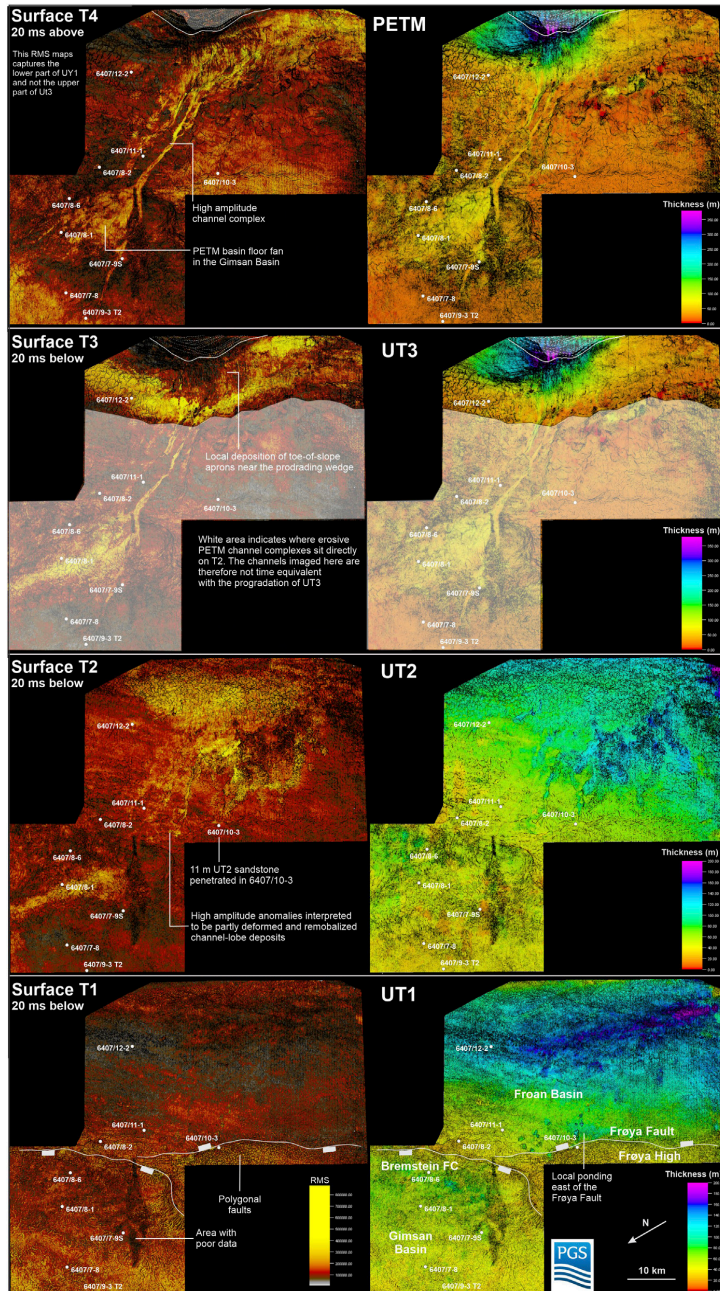


FIGURE 5
(Continued).

sigmoidal geometries with foresets dips of 4–8 degrees (Figure 4). The clinothem thickness decreases rapidly as the toesets pinch out and downlap T2, and the thickness map in Figure 5 shows that the

UT3 depocenter is constrained to the proximal part of the study area. The intra UT3 FDC map (Figure 7B) shows several narrow slope channels basinward of the shelf break feeding sediment to toe-of-slope

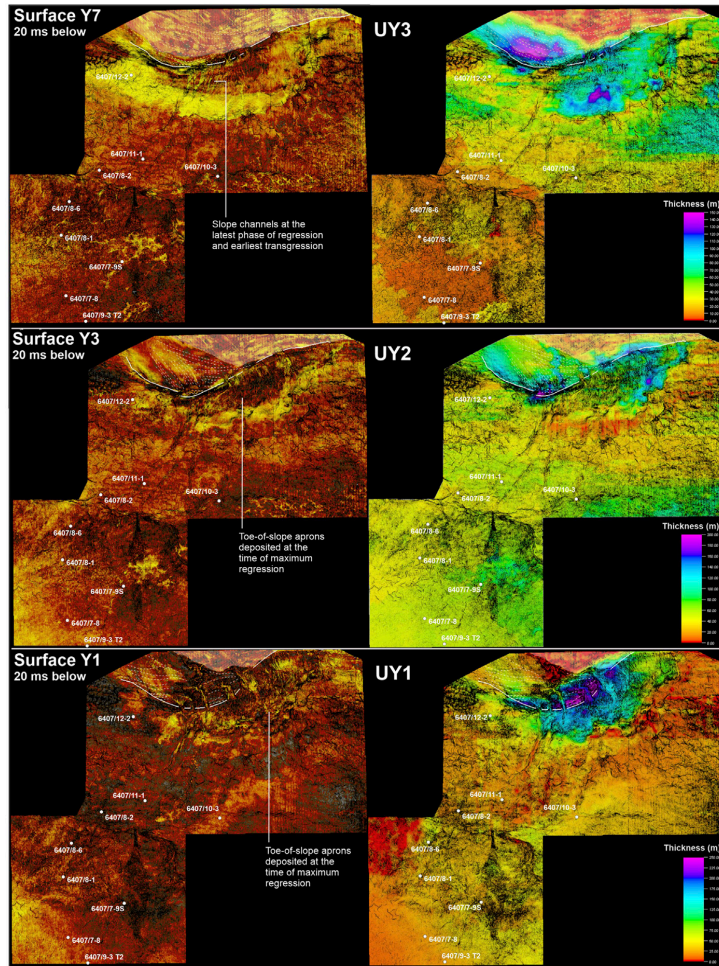


FIGURE 5 (Continued). Co-blend of RMS amplitude and variance attributes, and thickness map for units UT1 to UY3. Since the PETM occur the base of UY1, UY1 has been split to show the PETM and the post-PETM part of UY1 separately. Continuous white lines show the shelf break at the top of each unit, hatched lines show curvilinear features indicating progradation direction. Note that the scale of the thickness maps varies between units.

aprons. In cross section these are characterized by chaotic and discontinuous seismic facies (Figure 7B, section A).

In all wells within the study area, the basal downlap surface is marked by increasing GR values (Figure 6). In 6407/12-2, which is located at the flank of the high-amplitude channelized area, UT3 is represented by an upward decrease in GR and density, and is dominated by silty, micaceous claystone with abundant glauconite (Figure 3).

Interpretation

The up to 400 m thick clinothems in UT3 represents a prograding shelf margin (Patrino and Helland-Hansen, 2018) and the first arrival of the prograding shelf-slope wedge in the study area, sourced from the hinterland in the SE. The variable amplitude character and seismic

geomorphology suggests mixed lithologies where slope channels formed along a relatively narrow section of the slope. The thickness map in Figure 5 and the abrupt basinward thinning of the unit in cross section in Figure 4 show that sediments were largely trapped within the prograding wedge and at the proximal toe-of-slope during this early phase of progradation with no evidence of bypass to the more distal part of the basin.

Unit Y1—Ypresian (T3–Y1)

Unit Y1 (UY1) is bounded at the base by surface T3 and at the top by surface Y1 (Figure 4). The base coincides with the first occurrence of *Apectodinium augustum*, but except for the internal top *A. augustum* marker (surface T4), there are no major bioevents separating the top of UY1 from UY2. The top of UY2 coincides

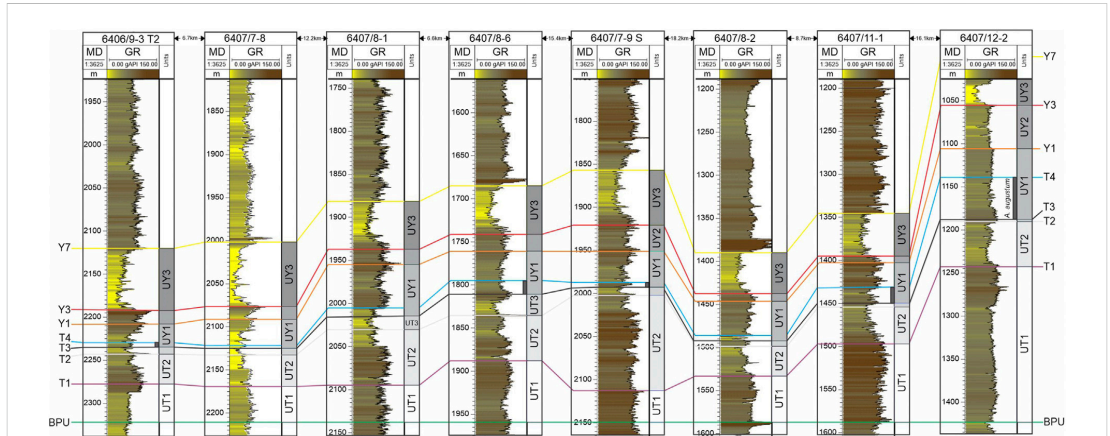


FIGURE 6
 Log correlation from proximal (right) to distal (left) showing the gamma ray (GR) character of each unit and key bounding surfaces. Surface T2 is the regional downlap surface and Y7 is the uppermost onlap surface. Despite high variability in GR, the succession is generally mud-dominated and low GR signature tend to reflect siliceous ooze and volcanic tuff. Where biostratigraphic data are available, the presence of *A. augustum* is shown with a grey bar, indicating the thickness of the PETM interval in each well. See Figure 5 for well locations. BPU = Base Paleogene unconformity.

with the first occurrence of the dinoflagellate *Cerodinium wardenense*, suggesting a latest Thanetian to early Ypresian age for UY1 and UY2 (Bujak and Mudge, 1994; Brunstad et al., 2013). The PETM, defined by the presence of *A. augustum*, is thus present at the base of UY1, between surface T3 and T4 (Figure 3).

Surface T3 at the base of UY1 is a downlap surface, which locally truncates reflections below (Figure 7C, section B). The upper Y1 surface is an erosional toplap surface. Tangential clinothems of UY1 are initially 200–300 m thick, but thin basinward to less than 50 m and have foreset dips of 3–5 degrees. Internally, the unit has chaotic seismic facies in the central part, becoming more continuous as the reflections downlap to the north, south and west.

Figure 7C shows the seismic geomorphology at the T3 surface, which correlates with the lower part of the PETM in 6407/12-2 and 6407/11-1. The map shows a series of slope channels that merge into two NW-trending, low-sinuosity channel complexes (up to 1.8 km wide) (Figure 7C, section A). These channel complexes extend ~20 km westward into the Gimsan Basin where they feed a basin floor fan (~450 km²). Several smaller channels are observed in the medial and distal part of the fan. The fan outline is defined by moderate amplitudes in cross section and RMS maps, decreasing towards the NW (Figure 5). West of the Bremstein Fault Complex, UY1 is limited to a single peak-response in the seismic data (Figure 2). The FDC map in Figure 7D shows the intra UY1 geomorphology during continued progradation after the PETM. Sediment delivery to the outer shelf and slope appears more localized and several slope channels (Figure 7D, section A) converge at the base-of-slope.

The upper bounding surface of UY1 cuts as deep as 130 m into underlying strata of UY1 and UT3, forming several incised valleys, gullies and erosional networks along the shelf (Figure 7E). At the distal end of the valleys, the seismic facies appear chaotic in cross section (Figure 7E, section A), but the FDC map demonstrates a network of coalescing channels and lobes. In the northern part, the terminal lobes in front of the erosional networks are expressed at the upper part of the

unit where they have a continuous seismic expression. In general, the amplitude strength often increases in the distal part and the seismic reflections show apparent aggradational downlap geometries (Figure 4). Farther west and into the basin, the attribute maps show a homogenous attribute expression and areas with polygonal faulting.

Biostratigraphic data from 6407/12-2 and 6407/11-1 show that the lower part of UY1 (between T3 and T4) is characterized by the presence of *A. augustum* and corresponds to the PETM (Figure 3). In 6407/12-2, the acme of *A. augustum* occurs just above the basal T3 surface and the diagnostic fossil is present within a 54 m thick interval. The basal part of this unit is also characterized by an increase in amorphous organic matter and a marked decrease in microfauna diversity and absence of agglutinated foraminifera. Wells in the distal part of the system show a similar trend, where the presence of *A. augustum* coincides with increased input of amorphous organic matter and absence of foraminifera. Here the PETM succession is more condensed and typically <20 m thick (Figure 6).

Across the study area, the pre-PETM Thanetian lithology is generally characterized by grey claystones with limestone beds. In 6407/12-2, which is located at the flank of the high-amplitude channelized area, the transition into the lower PETM part of UY1 is expressed by an increase followed by a decrease in GR and density, and the lithology is dominated by silty, micaceous claystone with abundant glauconite (Figure 3). In 6407/11-1, located between the two high-amplitude channel complexes, the unit is dominated by silty claystone. The distal part of the PETM basin floor fan is penetrated by 6407/8-1 (Figure 5), documenting claystone with traces of siltstone. Similarly, all other wells positioned along the flank of the system also show dominance of claystone (Figure 6) (Supplementary Table S1). Low GR in the distal wells reflect increasing content of siliceous ooze.

Interpretation

UY1 documents ~7 km of continued shelf-slope progradation relative to the previous UT3 shelf break. The base of the unit correlates with the

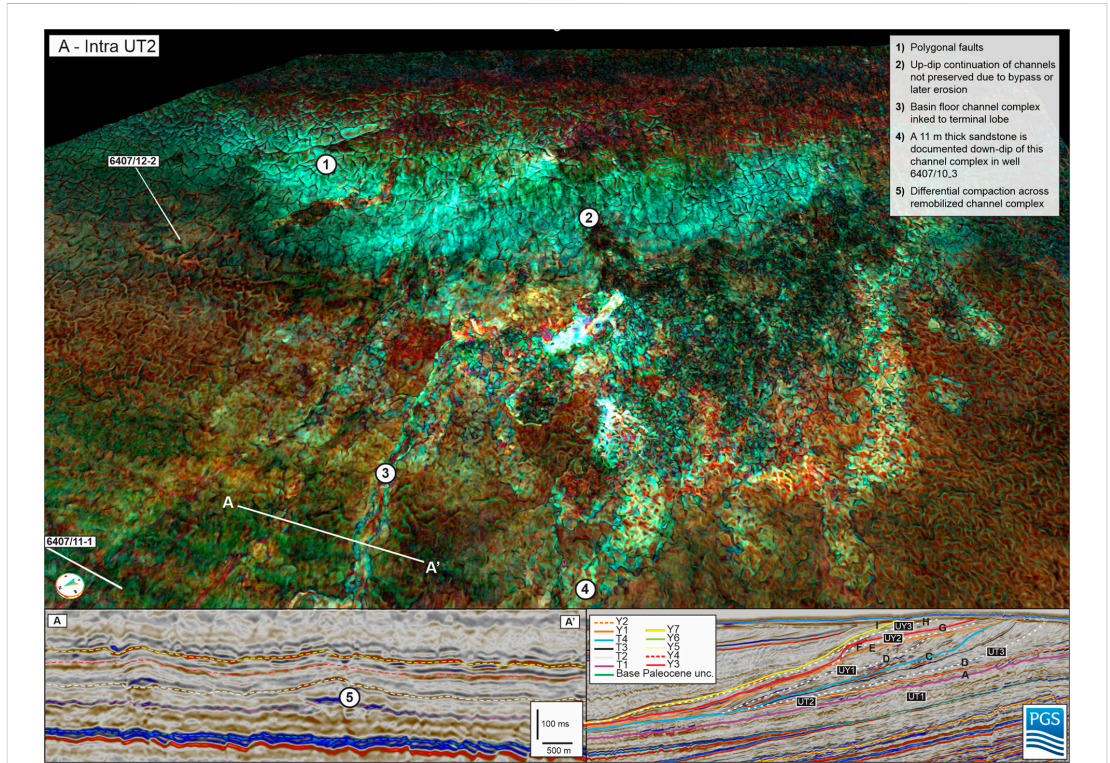


FIGURE 7

Frequency decomposition maps (FDC) with RGB blending showing seismic geomorphology within and between UT2 to UY3. (A) First phase of sand delivery to the Froan Basin. The map shows channel complexes and terminal lobes with poorly defined margins due to later remobilization. Line (A–A′) shows cross-section of remobilized and partly intruded channel sands into the overburden. (B) Pre-PETM map showing slope and toe-of-slope deposition during normal regression. The deposits downlap T2 and the channel complex in front of the system is younger. (C) Syn PETM deposition is characterized by direct sediment delivery to the shelf edge and slope bypass in wide erosive channel complexes. Channels merge at the toe-of-slope where they form larger channel complexes that bypass sediment to the basin floor and the Gimsan basin-floor fan. Inset map shows the full PETM shelf and deep-water system. Line (A–A′) shows the tie to well 6407/11-1 and the PETM interval defined by the presence of *A. augustum*. Line (B–B′) shows erosive bases of slope channel complexes. This indicates that the channels accommodated larger and more erosive gravity flows during the PETM. (D) Immediately after the PETM, sediment supply to the shelf break appears to have been more restricted. Narrow slope channels line (A–A′) show that these confluence at the toe-of-slope. (E) At the time of maximum regression of UY1, the subaerial unconformity was associated with incised valleys and erosional networks. These incisions formed during and at the time of maximum regression and some may have been modified during subsequent transgression. The map show that the main form of sediment delivery to the toe-of-slope was by unconfined, fluvial dominated shelf-edge deltas in the south and longshore transport to canyons in the north. Line (A–A′) shows a cross-section through the forced-regressive toe-of-slope aprons. (F) During progradation of UY2, there was a change in sediment transport direction and the northern part of the shelf was deposited as a result of longshore transport from the south Line (A–A′). (G) During maximum regression of UY2, sediment was delivered directly to fluvial dominated shelf-edge deltas in the south. Littoral currents transported sediment northward and was funneled through canyons, which formed during transgression. Line (A–A′) shows that UY2 canyons formed laterally from UY1 valleys suggesting that they formed during transgression. (H) Progradation during UY3 was characterized by significant aggradation on the shelf and sediment delivery was unconfined at the shelf break. Slope and toe-of-slope deposition was not as affected by bottom currents and have more radial depositional geometries. Line (A–A′) show the relatively dim toe-of-slope seismic facies compared to the higher amplitude forced regressive facies. (I) Ypresian backstepping resulted in transgressive erosion as shown in Line (A–A′). Immature slope channels reflect fine-grained sediment delivery during transgression. Transparent polygons show where units have been eroded by the upper regional unconformity or the seafloor, or where the image show younger stratigraphy as in (B). Main surfaces are shown as hatched lines in seismic cross sections to improve visibility. Dotted lines indicate shelf-break location. The seismic line in (A) shows the stratigraphic position of each FDC map (A–I). The stratigraphic position of extractions between main surfaces are indicated as a white hatched line and annotated.

transition from sigmoidal to tangential clinoforms in the proximal part of the system (Figure 4). In FDC maps, this transition is marked by a change from more localized sediment delivery to the shelf edge (Figure 7B), to a wider zone of direct sediment delivery to the shelf-edge delta during the PETM (Figure 7C).

Differential compaction across the PETM channel complexes suggests that they are sand-dominated in the proximal part (Figure 7C, section A), but gradually becoming less sand-rich

basinward (higher aspect ratio basinward). Even though the wells show that the basin floor fan is mud-rich, the channel complexes did not develop pronounced levees, suggesting rapid deposition and down-dip changes in lithology. The transition from grey claystone with limestone beds to more micaceous claystones with local glauconite and siltstone is interpreted to reflect shoaling of the carbonate compensation depth and input of more terrestrial material during the PETM (Carmichael et al., 2017). The channel-

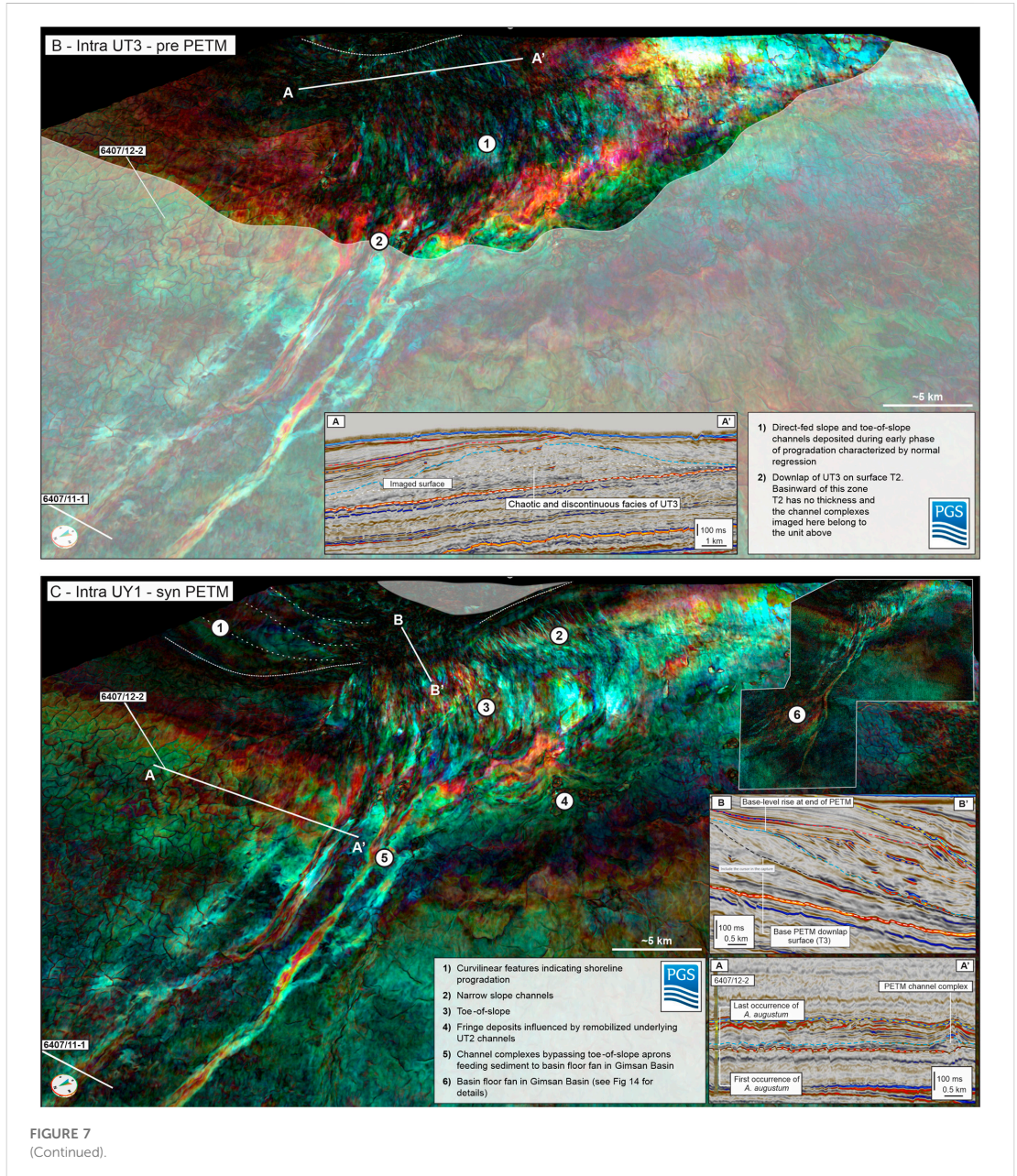


FIGURE 7
(Continued).

lobe transition zone coincides with a decrease in basin-floor gradient west of the Bremstein Fault Complex, where the PETM fan widens and eventually tapers out toward the northwest.

Continued progradation above the PETM wedge resulted in the deposition of thick slope and toe-of-slope succession within ~12 km of the advancing shelf-edge. Incised valleys, canyons, gullies, and erosional networks in Figure 7E formed at the last stage of

UY1 progradation and during subsequent transgression within UY2. The curvilinear features seen in the FDC maps indicate that the early phase of progradation was toward the NE (Figure 7D), but that the orientation changed to a more northern direction at the last phase of regression (Figure 7E). The shelf-slope topography at the end of UY1 progradation formed a bathymetric template for later UY2-UY3 progradations.

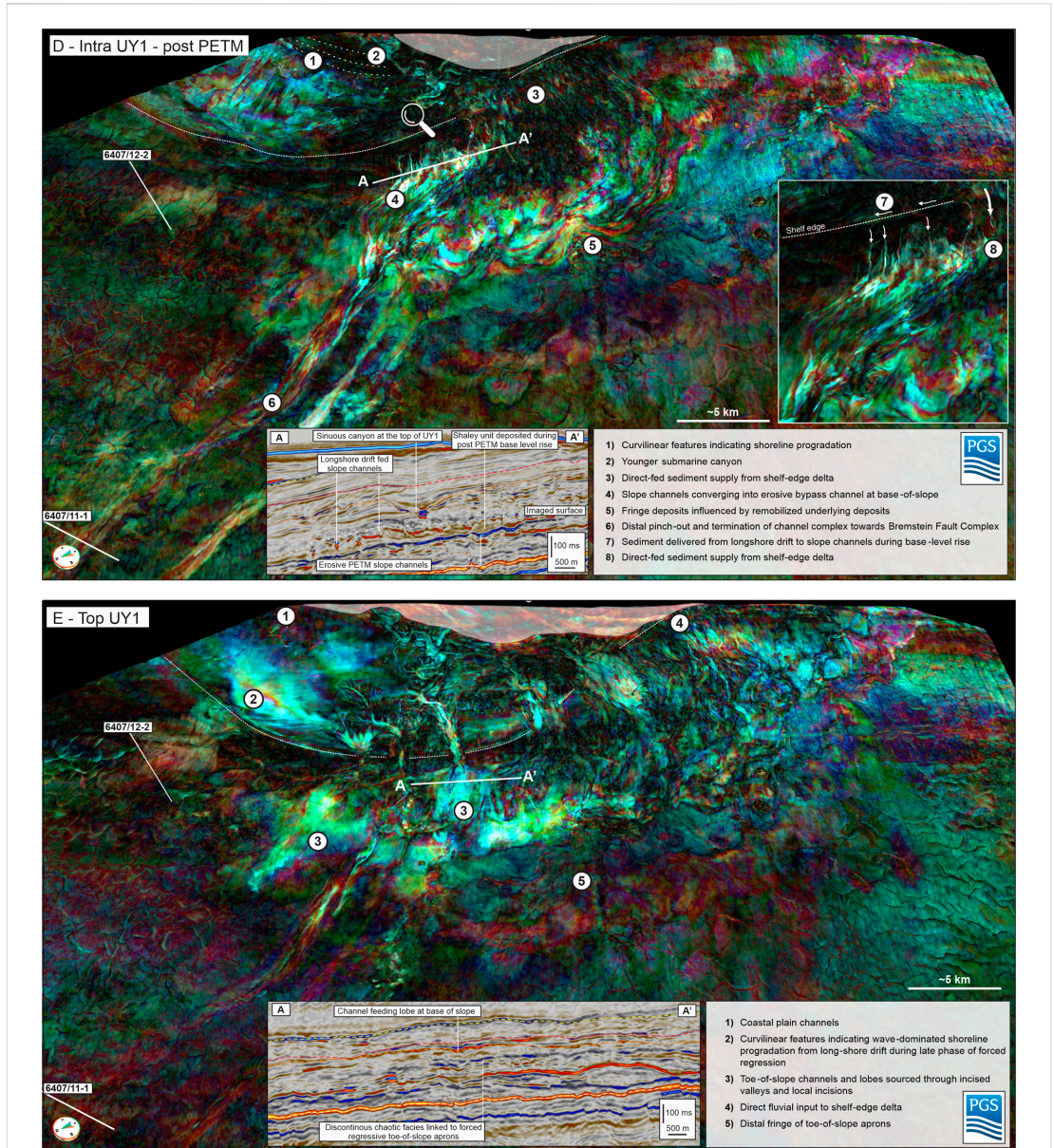


FIGURE 7 (Continued).

Unit Y2—Ypresian (Y1-Y3)

Unit Y2 (UY2) is bounded below by surface Y1 and at the top by Y3 (Figure 4), and is of early Ypresian age (see above) (Figure 3). The basal Y1 surface is generally downlapped by a single set of tangential clinofolds up to 80 m thick with foreset dips of 5–9 degrees (except for inside the main canyons where it is an onlap surface). The clinofolds of UY2 did generally not advance beyond the UY1 shelf

break; however minor progradation of up to 1 km beyond the previous shelf break occurred locally (Figure 2). Similar to the unit below, the clinofold set is marked by a sharp toplap surface. The RMS and FDC maps in Figures 5, 7F,G show similar characteristics as the underlying unit, with incised valleys and gullies on the shelf (Figure 7G, section A) transitioning into channels and chaotic attribute expressions to the west. In contrast to UY1, the thickness of UY2 is more variable with a local

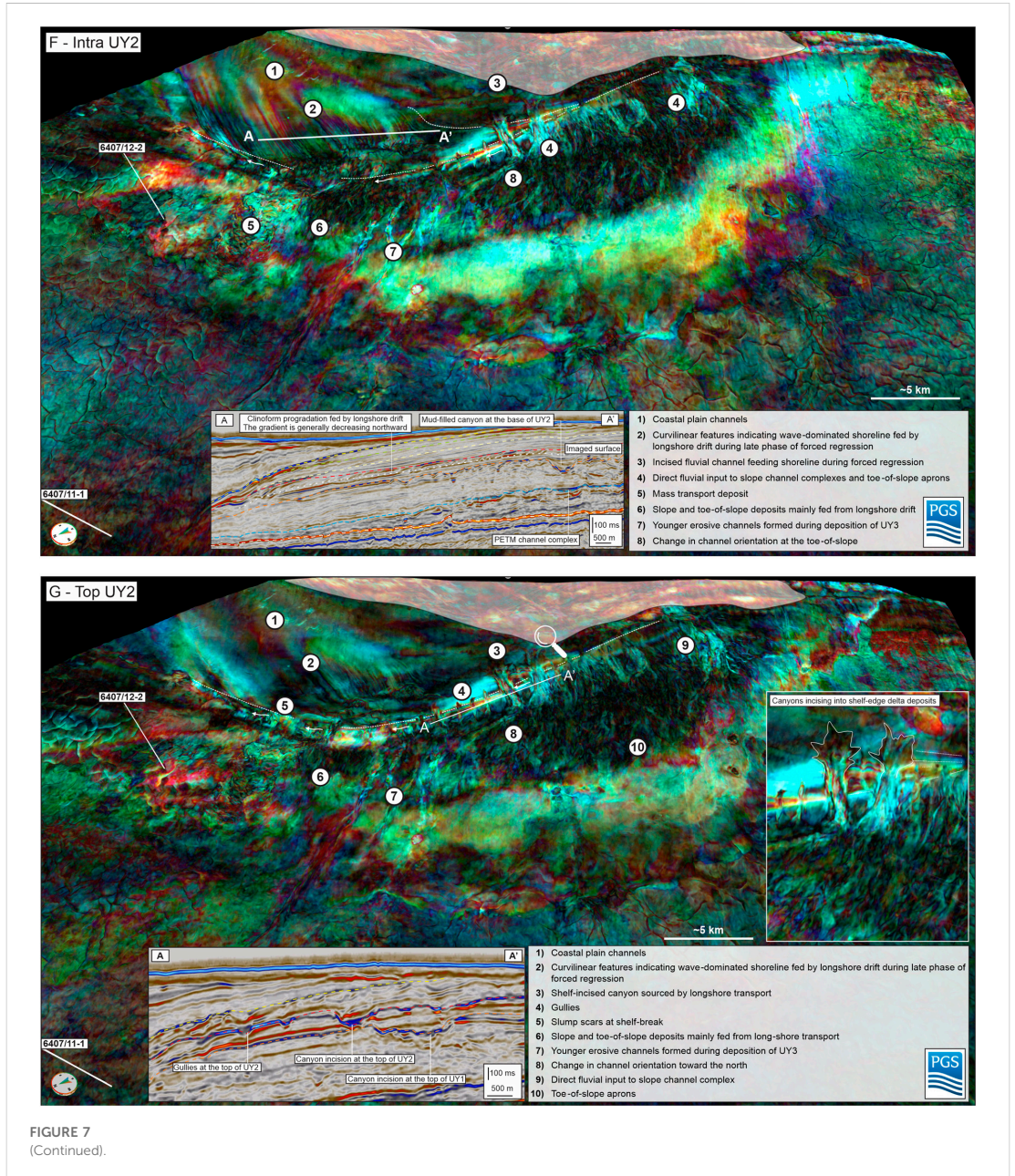
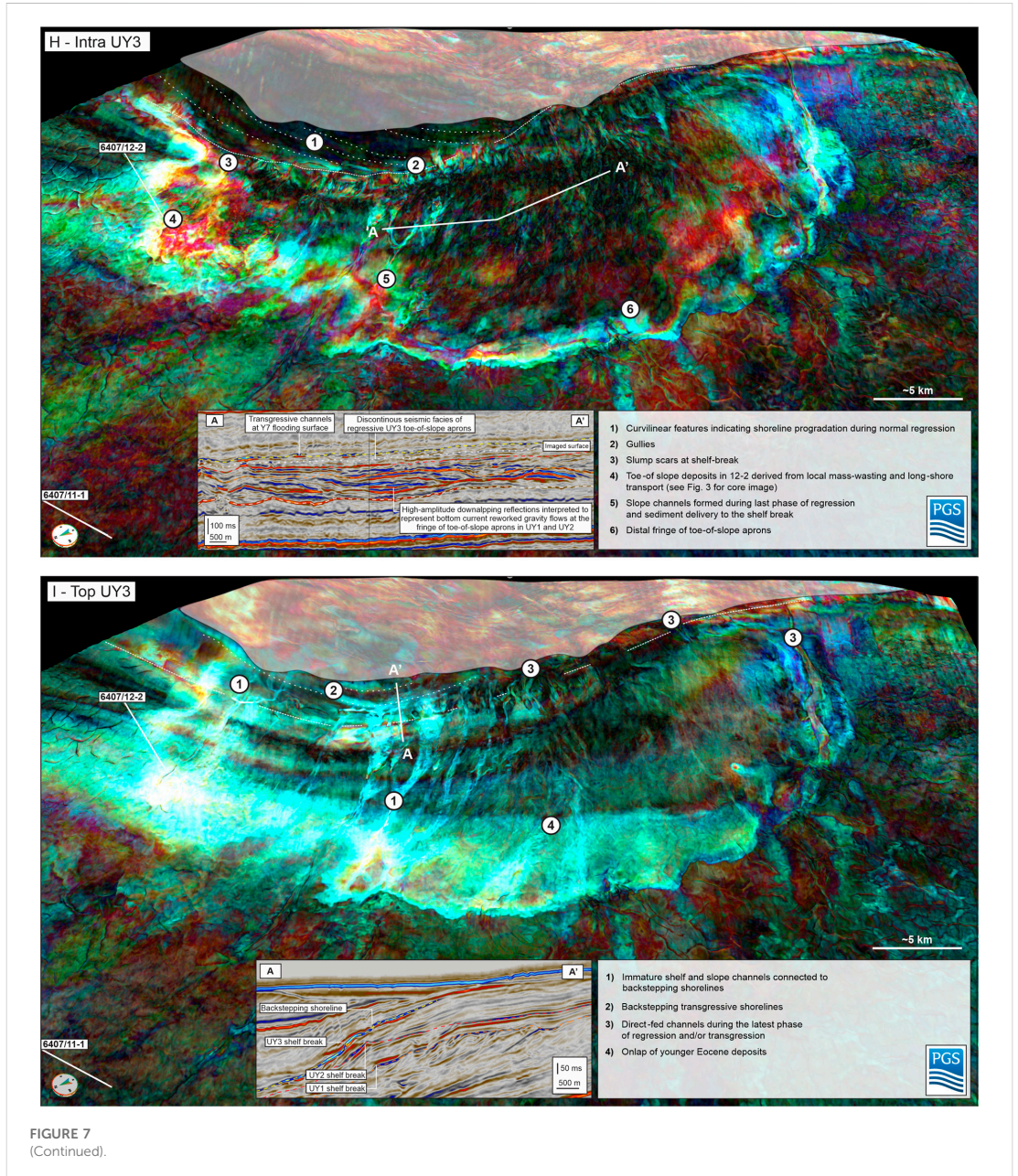


FIGURE 7 (Continued).

deponenter in the north. Also in contrast to the unit below, the intra UY2 FDC map in Figure 7F shows well-defined curvilinear features suggesting significant northward shelfal progradation during the last phase of regression. The foreset angles of these clinofolds decrease northward, from as much as 9 degrees near the main canyon to less than 5 degrees in the northernmost part of the shelf (Figure 7F, section A).

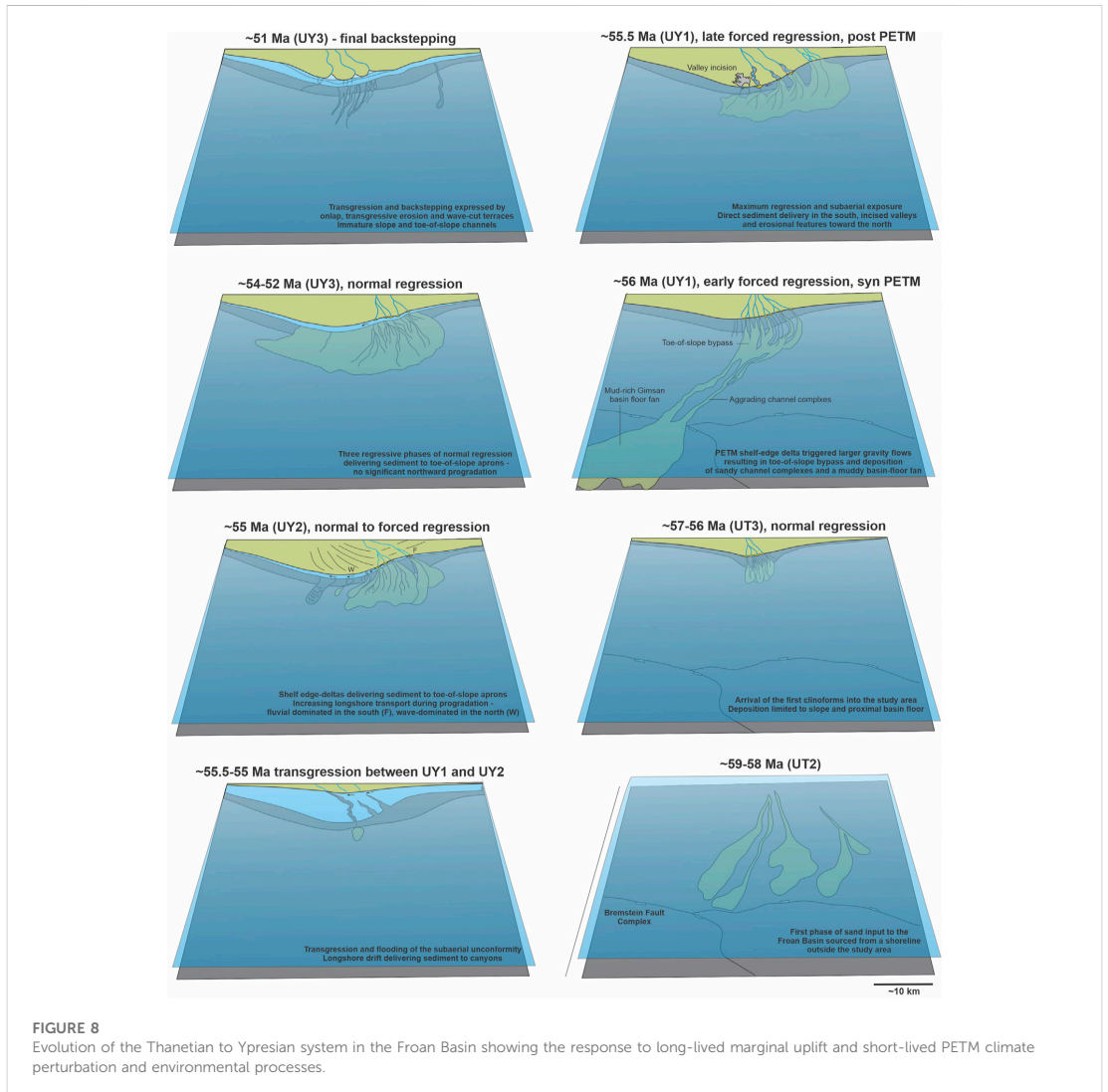
In 6407/12-2, the unit shows a minor decrease in GR toward the top, comprising claystone with siltstone and sandstone stringers and minor limestone beds (Figure 3). In the distal part of the system, the unit is typically <20 m thick and comprise claystone with traces of siltstone, limestone and tuff. Low GR values reflect high content of silicious ooze (Figure 6).



Interpretation

The lowermost part of UY2 represents backfilling of submarine canyons formed during incision of surface Y1. Transgression associated with this surface shifted the shoreline at least 4.5 km to the east. The main body of UY2 is interpreted to represent a renewed phase of progradation following flooding of UY1. The single clinothem set represents a delta advancing across the

flooded shelf (Patrino and Helland-Hansen, 2018), eventually reaching the UY1 shelf break. The FDC maps show that shelf progradation was mainly toward the north (Figure 7F); except for an MTD south of 6407/12-2, there are no indications of fluvial channels feeding the shoreline and no indications of slope channels and toe-of-slope aprons beyond the shelf break as observed in the south. The curvilinear features in the north are

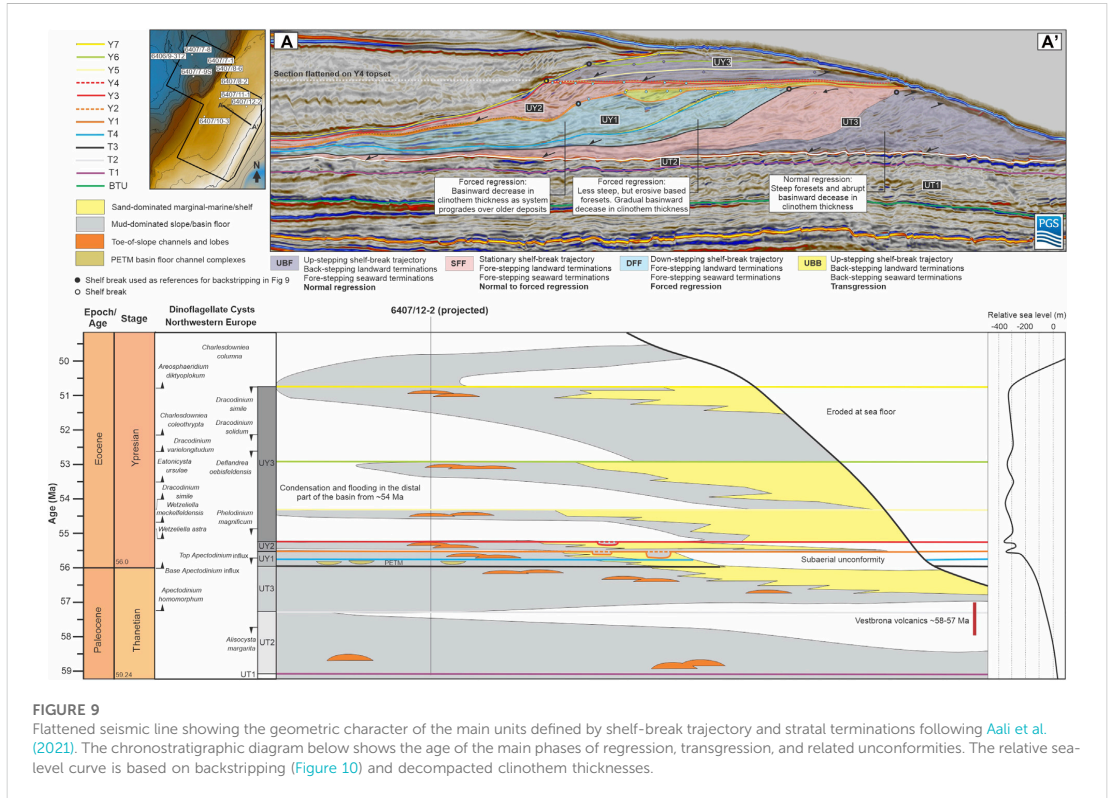


interpreted to reflect a wave dominated coastline (Nienhuis et al., 2015; Berton et al., 2021). The main point of sediment delivery to the outer shelf was in the south, where slope channels seem to connect to the shelf break (Figures 7F,G). This is evident from an incised channel linking up with the southern canyon in Figure 7F, indicating direct fluvial input to a shelf-edge delta. A lateral change in processes regime from fluvial dominance in the south to more wave dominance to the north is also expressed by the change in gradient of the delta clinofolds. To the south, direct supply from rivers is expressed by steep delta foresets up to 5–9 degrees, decreasing northwards to ~5 degrees (Figure 7F, section A). A northward decrease in delta foreset gradient indicates increasing distance from the active fluvial system (Paumard et al., 2020; Zimmer

et al., 2020). The progradation direction and the lack of feeder systems in the north suggest that the wave-dominated coastline was mainly supplied by long-shore sediment transport. Both the thickness maps and the seismic geomorphology show that most of the sediments were deposited within ~11 km of the shelf break and that little sediment escaped to the Gimsan Basin in the west. This interpretation is also supported by the wells which show predominantly claystone with traces of siltstone, limestone and tuff (Figure 6).

Unit Y3—Ypresian (Y4–Y7)

Unit Y3 is bounded by surface Y4 at the base and by a regional flooding surface Y7 at the top (Figure 4). An Ypresian age is

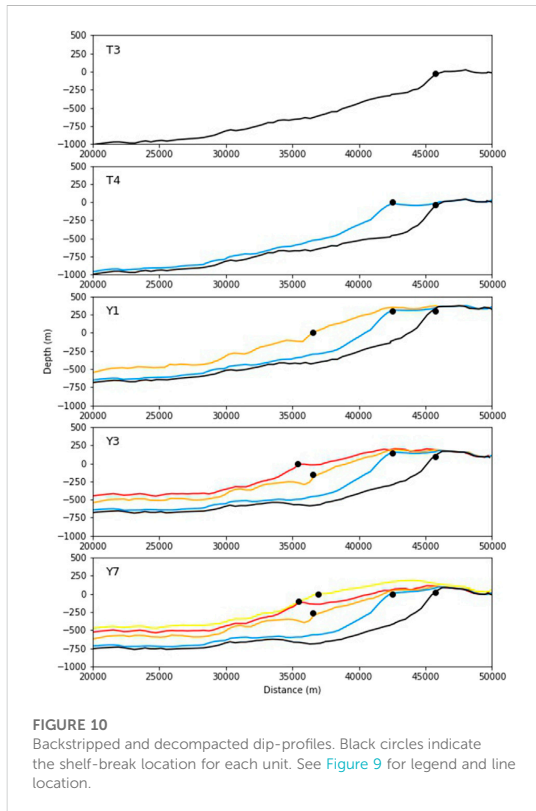


suggested as the top of the unit coincides with last occurrence of *Dracodinium simile* (Speijer et al., 2020) (Figure 3). The thickness map shows that the main depocenter was on the slope and toe-of-slope region in the south, but the unit also has a prominent depocenter in the north with thicknesses up to 150 m (Figures 2, 5). In cross sections the unit shows an overall aggradational character with three 30–70 m thick clinothem sets stacked on the shelf (Figure 4). These sets are not regional across the shelf and the foreset angle is lower than in the units below (2–4 degrees). They are all constrained to the topsets of the underlying unit and none of the clinothem sets prograded significantly beyond the underlying UY1 or UY2 shelf breaks. The upper bounding surface is a regional onlap surface, but it is also erosive near the outer shelf where several gullies formed along the outer shelf. The intra UY3 FDC map in Figure 7H shows the last regressive phase of UY3, imaging a wide zone of slope channels feeding toe-of-slope aprons (Figure 7H, section A). The aprons have a more radial geometry compared to the units below and there is less evidence of northward sediment transport. This shift in transport direction is also expressed by the curvilinear features on the shelf indicating more radial progradation towards the NW. Similar to the units below, the northern part of the system is less channelized and the FDC map in Figure 7H shows small gullies and slump scars in addition to mass-wasting deposits at the toe-of-slope. The mass wasting deposits were penetrated and partially cored by 6407/12-2, documenting an 80 m thick interval of interbedded sandstones and claystones (Figure 3). Specifically, the 16 m thick cored interval document fine to medium-

grained sandstone beds deposited as turbidities, interbedded with layers of claystone, tuff and slumped heterolithic deposits with high terrestrial organic content. In the distal wells, the unit is dominated by low GR tuff and non-calcareous claystone (Figure 6). The character of the uppermost Y7 onlap surface is imaged by the FDC map in Figure 7I. The map shows several low-sinuuous channels extending landward of the shelf break. The upper surface also has a terraced appearance in cross sections (Figure 7I, section A).

Interpretation

Unit Y3 represents several phases of shoreline regression and transgression following flooding of UY2. The main depocenter was on the shelf and seismic cross sections show that the individual phases of regression were able to reach the shelf-break, at least locally. Similar to the unit below, most of the gullies, slope channels and toe-of-slope aprons are located in the southern part of the system, suggesting that this was the area of direct sediment supply as before. The heterolithic tuffaceous sandstones in 6407/12-2 were deposited at the last stage of regression or earliest transgression when most of the sediment was trapped in topsets. The final backstepping of the system is documented by the local terraced morphology (Figure 7I), interpreted to be wave erosion along a backstepping shoreline. The low-sinuuous channels seen in FDC map are interpreted to represent relatively immature shelf and slope fairways (Maier et al., 2013), formed during transgression and backstepping of the system.



Sequence stratigraphy and relative sea-level fluctuations

The Late Cretaceous and earliest Paleocene coastline was located landward of the Froan Basin and seismic data show that the basin had a ramp-like geometry with gradual deepening toward the Bremstein Fault Complex in the west (Figure 2). Initial hinterland uplift in the earliest Danian is documented by input of clastic material to the margin south of the study area (Gjelberg et al., 2001; Sømme et al., 2013). The first phase of sand delivery to the Froan Basin was in the early Thanetian (UT2) (Figure 8), but the shallow-marine system feeding these channels and lobes is not preserved. When the coastline reached the Froan Basin in the late Thanetian (easternmost UT3 clinoforms) (Figure 8), progradation was characterized by an up-stepping shelf-break trajectory, backstepping landward terminations and fore-stepping seaward terminations (UBF, sensu Aali et al., 2021) (Figure 9), indicating moderately high $\delta A/\delta S$ conditions during normal regression. Moderately high $\delta A/\delta S$ is also expressed by sigmoidal clinoforms showing rapid basinward thinning and downlap onto the T2 surface (Figure 4). Continued progradation is characterized by a static shelf-break trajectory and fore-stepping landward and basinward terminations (SFF, sensu Aali et al., 2021), indicating gradual decrease in accommodation. This SFF unit represents a transitional phase between normal and forced regression.

Backstripping and decompaction of the dip-transect in Figure 9 and the individual clinothem provide an indication of the water depth in front of the advancing system, suggesting ~350 m near the shelf break and ~800 m at a reference location 20 km along the dip transect (near well 6407/11-1) during this time (Figure 10).

Progradation at the base of UY1 is dominated by a down-stepping shelf-break trajectory and fore-stepping landward and seaward terminations (DFF, sensu Aali et al., 2021), indicating negative $\delta A/\delta S$ conditions and forced regression. The unit is also marked by a basinward increase in thickness of the lower foresets and bottomsets pointing to loss of accommodation at the topset and increasing sediment accumulation on the slope and toe-of-slope region (Figure 9). The lower part of this unit coincides with the PETM (Figure 8). The FDC map in Figure 7C shows that the longest run-out of channel complexes and development of basin-floor fans occurred during the PETM, at the early phase of forced regression. This forced regressive unit is succeeded by a thin package (above T4) showing up-stepping and backstepping terminations (UBB, sensu Aali et al., 2021), high $\delta A/\delta S$ conditions, and base-level rise. Backstripping and decompaction suggests water depths of 150–250 m near the shelf break and ~750 m at the reference location at the PETM (Figure 10).

Subsequent progradation of UY1 above T4 and after the PETM is characterized by continued forced regression (DFF). The clinothem thicknesses rapidly decreased as relative sea-level fell and as the system prograded over older slope and toe-of-slope deposits (Figure 4), indicative of negative accommodation (Posamentier and Morris, 2000). Backstripping and decompaction of clinothem (Y1) suggests that the water depth was less than 50 m at the shelf edge and ~345 m at the reference location during maximum regression of UY1 in the earliest Ypresian (Figure 10). At the reference location near 6407/11-1, this relative sea-level fall of ~460 m reflects ~260 m of deposition post PETM combined with ~200 m of sea-level fall due to tectonic uplift.

Valley incision (up to 130 m deep) and erosional features on the outer shelf developed as a part of the subaerial unconformity (Y1) and the expression of this surface is shown in Figure 7E. At the time of maximum regression and subaerial exposure, ~0.5 Myr after the PETM, deposition was largely constrained to toe-of-slope aprons within ~15 km of the shelf break. The aprons were fed directly by unconfined shelf-edge deltas in the south and incised valleys and erosional features in the north (Figure 9). Valley incision ~130 m below the shelf break at maximum regression suggests a total relative sea-level fall of more than 300 m between ~57 and ~55.5 Ma (between the earliest phase of UT3 progradation and maximum regression of UY1).

Y1 is overlain by another UBB (transgressive) unit indicating high $\delta A/\delta S$ conditions and flooding of the shelf (Figure 9). Inside the canyons, the upper Y2 maximum flooding surface is marked by a trough in seismic reflection data and can be traced from the outer shelf to the distal toe-of-slope (Figure 4). This flooding resulted in a landward shift of the shoreline of at least ~5 km as suggested by the most proximal clinoform above Y2.

Progradation of UY2 is characterized by a stationary shelf-break trajectory and fore-stepping landward and seaward terminations (SFF, sensu Aali et al., 2021), indicating negative to balanced $\delta A/\delta S$ conditions and normal to forced regression (Figure 9). In contrast to UY1, UY2 deposits are preserved over a larger part of the shelf indicating higher accommodation during this phase of deposition. The decompacted clinothem are up to ~85 m thick near the shelf break, showing that significant

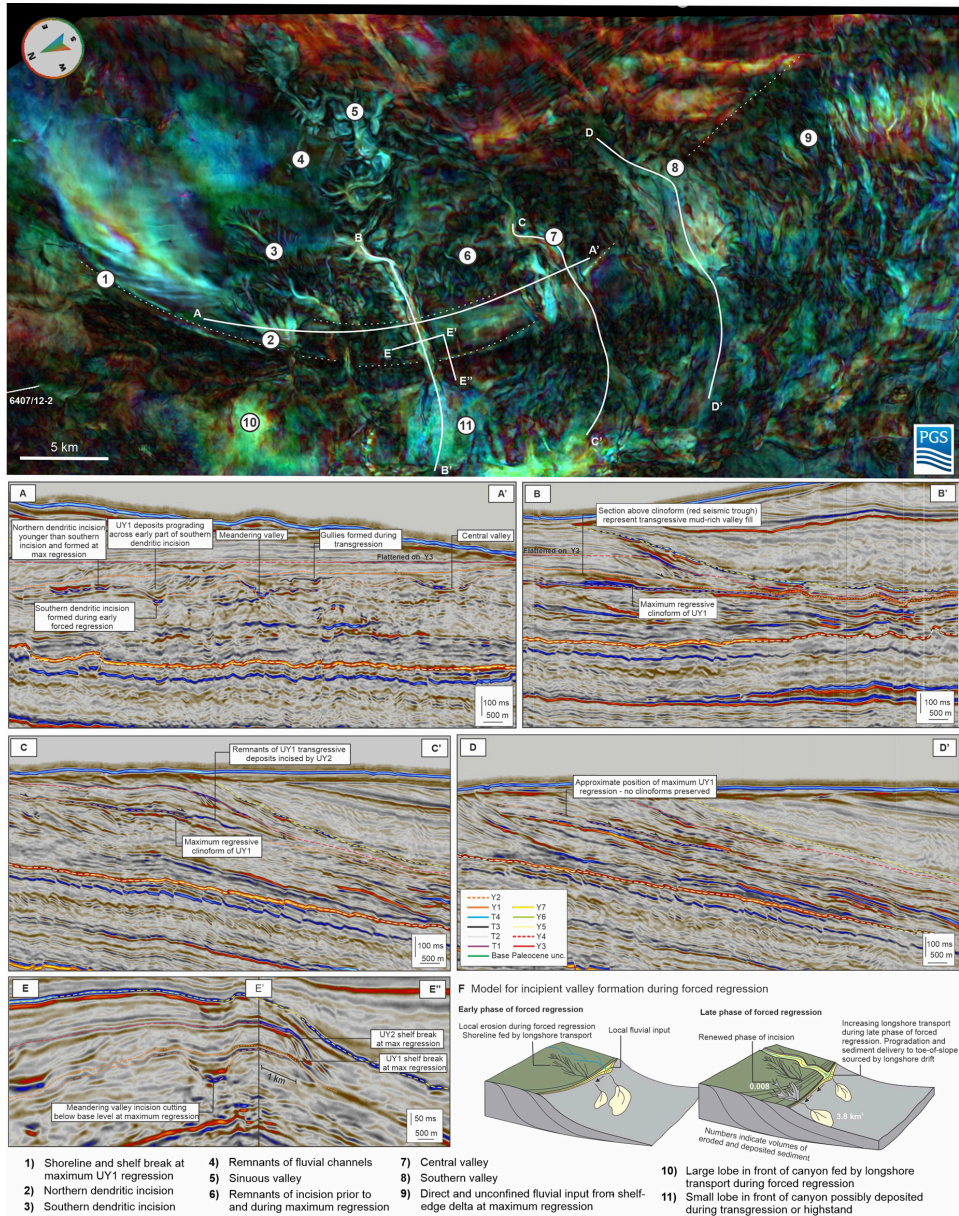


FIGURE 11

FDC map at the top of UY1 (surface Y1) showing the expression of the subaerial unconformity and associated erosional networks and incised valleys. Section (A–A') how the different generations of erosional network incision during forced regression. The clinoforms overlying the southern incision show that longshore drift advanced the margin after the incision and prior to erosion of the northern network. Sections (B–B'), (C–C'), and (D–D') show dip-sections through respective valleys. The recognition of clinoforms within the valleys suggest that they formed at maximum regression. Section (E–E') show the deep incision of the sinuous valley below the shelf break. F illustrates the suggested model for network incision during forced regression observed in seismic section (A). Numbers indicate eroded and stored sediments volumes in the northern incision and its frontal lobe, respectively. Excess volume in front of the northern incision show that the lobe must have been supplied also by longshore drift during forced regression.

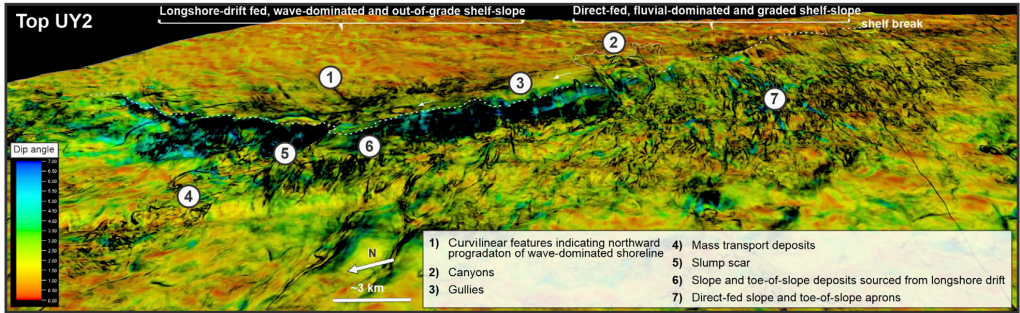


FIGURE 12

Dip-angle map (in degrees) from the top of UY2 showing along-strike variation in slope gradient. The southern area of direct fluvial input to the slope and toe-of-slope aprons has lower gradients reflecting a graded slope profile in the area of active sediment delivery. To the north, shelf progradation was mainly driven by longshore transport. Here the gradient is higher reflecting the out-of-grade margin development and inheritance from UY1 regression. Mass wasting processes are the dominant mechanism of sediment delivery to the toe-of-slope and canyon incision is deeper in this area.

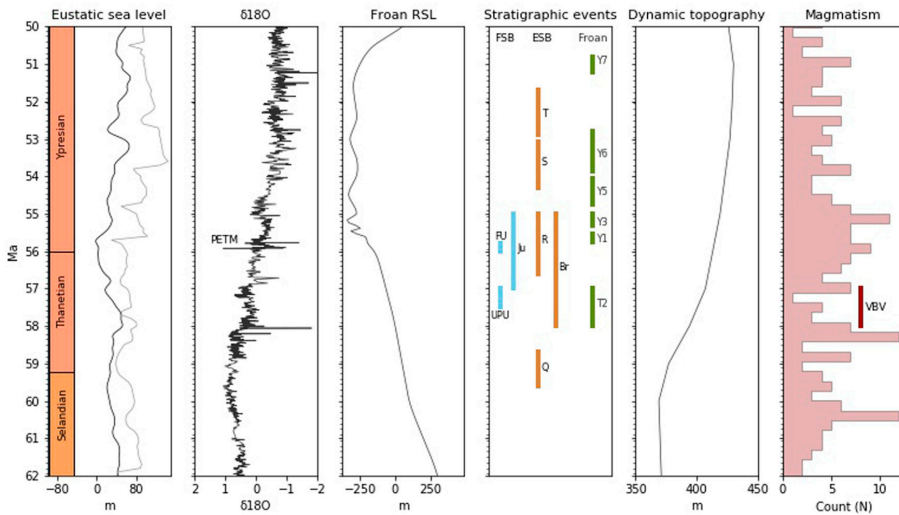


FIGURE 13

Comparison between key stratigraphic events in the Faroe-Shetland Basin (FSB), East Shetland Basin (ESB) and Froan Basin and other external tectonic and climatic controls. UPU = Upper Paleocene unconformity and FU = Flett unconformity from Jolley et al. (2021). JU = Judd unconformity from Hartely et al. (2011). Q-T landscapes from Conway-Jones and White (2021). Br = Bressay landscape from Stucky de Quay et al. (2017). Eustatic sea level curves from Kominz et al. (2008) (grey) and Miller et al. (2020) (black). $\delta^{18}O$ curves from Miller et al. (2020). Relative sea level (RSL) from the Froan Basin (Figure 9). Relative dynamic topography modified from Flament et al. (2014) with Iceland plume response from Skogseid et al. (2000). North Atlantic magmatism from Steinberger et al. (2019). Age of Vestrobrona volcanics (VBV) from Hafeez et al. (2017).

accommodation existed on the shelf after flooding of UY1. UY2 progradation stopped as the shoreline reached the previous shelf break. Similar to UY1, forced regression resulted in the formation of a subaerial unconformity; however, no erosional networks, valleys or large shelf-indented canyons formed at this time. The lack of shelf erosion and significant valley incision at the end of UY2 deposition is interpreted to reflect higher accommodation conditions compared to UY1. The FDC maps in

Figures 7F,G show that the main area of deposition during maximum regression was within ~15 km of the shelf break, similar to UY1, and there is no evidence of sediment bypass to the more distal Gimsan Basin. Backstripping suggests that the water depth at the reference location was ~340 m during the end of UY2.

UY2 is locally overlain by another thin UBB package indicating high $\delta A/\delta S$ conditions and transgression. The upper Y4 surface represents another maximum flooding surface, which can be

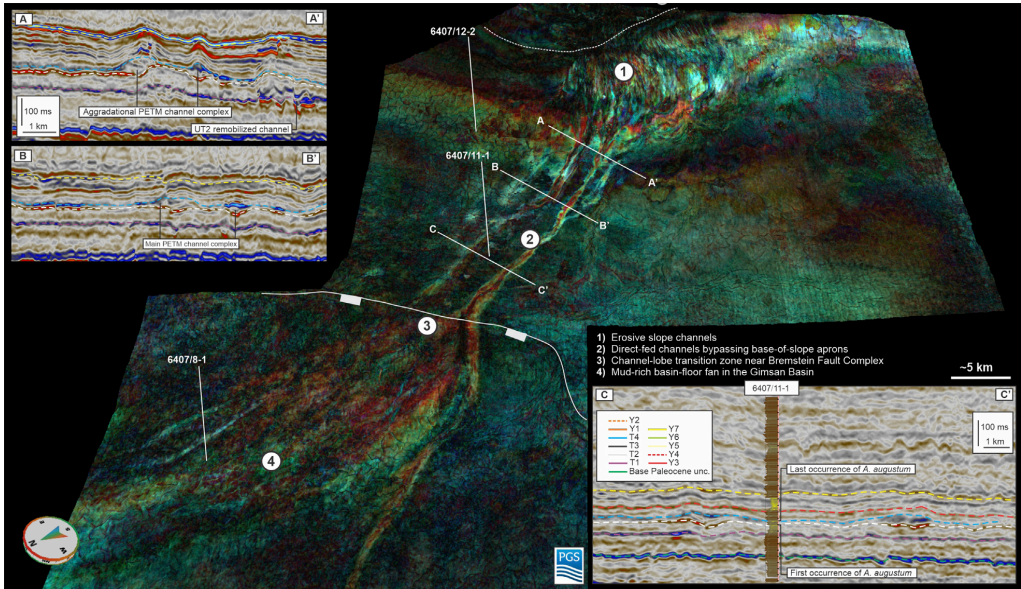


FIGURE 14 Oblique view of the PETM shelf, slope, and basin-floor fan system in the Froan and Gimsan Basin. Line (A–A') show proximal channel complexes basinward of toe-of-slope aprons. Differential compaction and aggradational internal geometries suggest that they are sand-rich in the proximal position. Channel complex aspect ratio increases rapidly down dip Line (B–B'), and near well 6407/11-1 only local internal channel geometries can be observed Line (C–C'). The tie to 6407/11-1 show that the channel complexes are linked to the PETM as they coincide with the occurrence of *A. augustum* in the well. The channel-lobe transition is located near the Bremstein Fault Complex. In the Gimsan Basin, the basin floor fan is characterized by distributary channels which terminates towards the north. Well 6407/8-1 (Figure 6) show that the fan is mud-dominated.

mapped in canyons and locally on the shelf (Figure 7F, section A). This transgression shifted the shoreline at least 6 km landward. Above the Y4 surface, UY3 comprises three clinofolds sets (bounded by downlap surfaces Y5 and Y6) showing up-stepping shelf-break trajectories, backstepping landward terminations and fore-stepping seaward terminations (UBF) indicating high $\delta A/\delta S$ conditions and normal regression (Figure 7F). UY3 is deeply eroded at the seafloor in the southern part of the study area and the stacking pattern cannot be determined. Decompacted clinothem thicknesses are 30–70 m for the three sets, giving an indication of the water depth at the outer shelf during each regressive phase. Despite higher shelf accommodation during UY3 deposition, each regressive phase was able to extend to the previous shelf-break before transgressing and backstepping. No deep valleys or canyons are observed within UY3 and the outer shelf is only cut by gullies that did not develop into larger canyons (Figure 7I). Backstripping and decompaction suggests that the water depth was 360 m at the reference location at the upper Y7 flooding surface.

Flooding and condensation commenced at ~54 Ma in the distal part of the system as UY3 topsets started to aggrade on the shelf (Figure 9). This flooding is marked by increasing GR values, more diverse microfossils indicating open marine circulation and the absence of terrestrial derived pollen and organic matter. On the toe-of-slope and on the shelf, final flooding took place around

51 Ma (Figure 8), marked by the last input of sand to the basin (Figure 3). This transgression is expressed by backstepping shorelines and local wave-cut terraces during step-wise transgression, similar to what has been documented in Pleistocene systems (Lebrek et al., 2022).

In summary, the UT1–UY3 succession documents a long-lived (~8 Myr) phase of relative sea-level fall of at least ~300 m followed by aggradation, backstepping and abandonment (Figure 9). Within this long-term trend, at least four short-lived (0.5–3 Myrs) phases of regression and transgression occurred, reflecting relative sea-level fluctuation of 30–80 m, suggesting higher-frequency fluctuations in sediment supply and accommodation along the margin.

Controls on incised valley and canyon formation

Several dendritic incisions, valleys, canyons, and gullies formed during development of UY1 and UY2 (Figures 7E,G; Figure 11). A seismic strike-line through UY1 shows that the northern and southern dendritic incisions did not form at the same time (Figure 11, section A). The southern incision formed during early forced regression, but a short base-level rise and a shift in progradation direction caused the network to be abandoned and

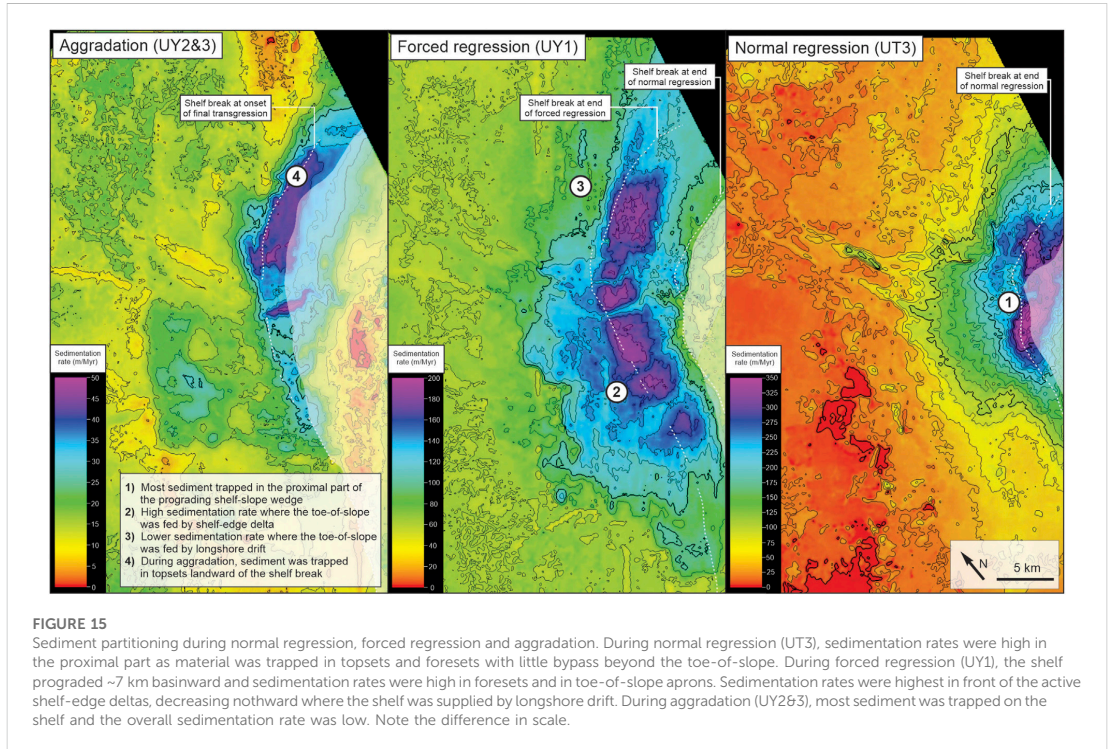


FIGURE 15

Sediment partitioning during normal regression, forced regression and aggradation. During normal regression (UT3), sedimentation rates were high in the proximal part as material was trapped in topsets and foresets with little bypass beyond the toe-of-slope. During forced regression (UY1), the shelf prograded ~7 km basinward and sedimentation rates were high in foresets and in toe-of-slope aprons. Sedimentation rates were highest in front of the active shelf-edge deltas, decreasing northward where the shelf was supplied by longshore drift. During aggradation (UY2&3), most sediment was trapped on the shelf and the overall sedimentation rate was low. Note the difference in scale.

overlay by shallow-marine deposits. This shift was driven by littoral sediment transport, evident from the observed change in progradation direction along the outer shelf. As the system approached maximum UY1 regression, the northern dendritic incision formed. This development is evident from the different incision depths in the seismic cross section (Figure 11F). Both incisions occurred landward of the shelf break at maximum regression, and both are associated with low-sinuuous, up to 150 m wide channels bypassing erosional products to terminal lobes at the toe-of-slope (Figure 11). These observations suggest that incision commenced *during* formation of the subaerial unconformity. Remnants of other dendritic incisions also occur farther south, pointing to a common development in areas away from main sediment entry points. Similar incisions have also been documented in Quaternary systems along-strike of large sediment entry points (Nordfjord et al., 2006).

Farther south, the FDC map shows a ~12 km long and up to 700 m wide sinuous valley incising up to 130 m into the underlying shelf and slope deposits. Remnants of fluvial channels on the subaerial unconformity and toe-of-slope channels within UY1 suggest that the area occupied by the sinuous valley also received sediment during forced regression. A seismic dip-line through the valley (Figure 11, section B) shows that the high amplitude reflections at the base of the valley terminate near the area of maximum regression, where a clinoform is observed. This clinoform is interpreted to represent the coastline at maximum

forced regression. Basinward of this clinoform, the erosive valley changes into a channel attached to a terminal lobe in its distal end. The valley fill is expressed by a low-amplitude trough in seismic data, suggesting backfilling by fine-grained material during subsequent transgression. Despite being primarily formed towards the end of forced regression, it is possible that headward erosion extended the valley landward during subsequent transgression.

Toward the area of direct sediment input to the south, the UY1 valleys have a more funneled morphology (Figure 11). The central valley is up to ~80 m deep and ~1.5 km wide at the shelf break where it terminates. The valley head is located ~5 km landward of the shelf break. Basinward, the FDC maps show a network of distributary channels (tens of meters wide) where the valley fed sediments to the toe-of-slope aprons. A cross section along the valley (Figure 11, section C) shows a clinoform at the position of maximum regression, indicating that valley incision primarily occurred during the last phase of forced regression. However, the canyon shape may also here have changed during subsequent transgression, resulting in the funneled morphology typical of transgressive estuarine systems (Dalrymple et al., 1992).

The southern valley has no preserved clinoforms at maximum regression but shows similar characteristics as the central valley. Incision is highest landward of the maximum regressive shoreline, reaching as deep as ~70 m. The valley widens to ~2.4 km near the shelf break, where it changes into submarine channels feeding sediment to

toe-of-slope aprons. Up-dip, the valley head is located ~5 km from the shelf break.

In the southern part of the shelf, the maximum regressive shoreline is eroded at the seafloor and the shelf break is not preserved. However, the FDC map shows that the area is dominated by nested channels up to 150 m wide instead of valleys, suggesting unconfined sediment supply from shelf-edge deltas to the slope during maximum forced regression.

Seismic reflection data show that the UY1 valley-fills are expressed by dim seismic facies (Figure 7G, Figure 10E), suggesting that subsequent transgression resulted in passive onlap and back-filling of valleys by fine-grained material, similar to many Pleistocene valleys (Blum et al., 2013; Wang et al., 2020).

As regression of UY2 commenced, previous incisions on the subaerial unconformity were filled to the extent that shorelines, slope channels and later incised canyons occupied new positions along the outer shelf. In contrast to UY1 valleys, UY2 canyons did not erode as far back across the shelf (maximum 1.5 km). Because these canyons eroded into the youngest clinoforms deposited at maximum regression, it is clear that UY2 canyon formation must have occurred during transgression. The FDC maps in Figures 7F,G show toe-of-slope channels linking up with the two canyons at the same time as there is no evidence of fluvial channels at the canyon heads. The absence of channels at the canyon heads, together with the northward prograding clinoforms, suggest that longshore transport from littoral drift was the main source of sediment to these northern canyons (Figure 8).

Collectively, the observations from UY1 and UY2 suggest that sediment was supplied by rivers directly to the shelf break along a wide section of the shelf. Incised valleys were not the main conveyers of sediment to the slope during forced regression. The different canyons and incision morphologies observed along the shelf are interpreted to reflect the proximity to the area of high sediment input. In areas of direct sediment delivery, the slope gradient was lower, and the system was graded, preventing deep canyon incision (Figure 12). Toward the north, where shelf-slope progradation mainly was maintained by littoral transport, steeper, out-of-grade slopes developed. These slopes were prone to mass-wasting, gully and canyon incision (Micallief et al., 2014; Prêlat et al., 2015).

Discussion

Stratigraphic expression of regional uplift—A North Atlantic comparison

The North Atlantic margins experienced regional uplift in the Paleocene and early Eocene (e.g., Parker, 1975). Although controversy still exists regarding the driving mechanisms behind the uplift, recent studies have documented several short- and long-lived phases of uplift along the Faroe-Shetland Basin (FSB) and East Shetland Basin (ESB) during the Paleocene and earliest Eocene (Nadin and Kuszniir, 1995; Underhill, 2001; Mudge and Jones, 2004; Mackay et al., 2005; Champion et al., 2008; Hartley et al., 2011a; Stucky de Quay et al., 2017; Soutter et al., 2018; Conway-Jones and White, 2022). Comparison of surfaces and events between the FSB, ESB and the Froan Basin (Figure 13) is relevant because the regions are located with variable distance from the outer rifted margin, areas of active volcanism and the center of the Iceland mantle plume. Similarities and differences between the areas may help to understand the nature

of allogenic controls and their stratigraphic expressions in different basin settings.

The Selandian to Ypresian shelf-slope wedge in the Froan Basin represents ~8 Myr of progradation in response to hinterland uplift. The base Paleogene unconformity below the unit documents a ~30 Myr hiatus in the proximal part of the basin where the entire Upper Cretaceous succession is missing. The hiatus is marked by an angular unconformity, similar to what has been documented in the British Isles (Gale and Lovell, 2018) and East Greenland (Hovikoski et al., 2021). Along the Norwegian margin, this initial phase of uplift resulted in hinterland erosion and increased sediment supply during the Danian (Martinsen et al., 1999; Gjelberg et al., 2001; Sømme et al., 2019).

Although regional uplift started in the Danian, the first arrival of coarse sediment to the Froan Basin occurred in the early Thanetian (59–58 Ma), expressed by remobilized submarine channel complexes and lobes within UT2 (Figure 7A, Figure 9). This phase of initial coarse clastic input was followed by erosion of the Thanetian T2 unconformity (~57–58 Ma). The T2 unconformity is time equivalent with the Upper Paleocene Unconformity in the Faroe-Shetland Basin (Jolley et al., 2021; Walker et al., 2022), where Jolley et al. (2021) suggested a causal link to a phase of increased volcanism.

Rapid uplift and transition from normal to forced regression occurred just before the PETM. Although this phase coincides with a period of eustatic sea-level fall of ~40 m (Figure 13), the observed relative sea-level fall of at least ~300 m in the Froan Basin is an order of magnitude larger, suggesting a tectonic control. Continued uplift culminated in the formation of subaerial unconformities (Y1 and Y3) in the Froan Basin between 55.5–55 Ma. These are coeval with the “R” and the “Bressay” landscapes observed along the ESB (Stucky de Quay et al., 2017; Conway-Jones and White, 2022), and the Flett unconformity in the FSB (Hartley et al., 2011b; Jolley et al., 2021; Walker et al., 2022), suggesting a common mechanism for their formation across the North Atlantic region.

Another similarity between the Froan Basin and FSB is the stratigraphic position of the PETM within the prograding shelf-slope wedge. Jolley et al. (2021) demonstrated that the PETM predated subaerial exposure and the formation of the Flett unconformity in the FSB in the same way as the PETM predates the Y1 subaerial unconformity in the Froan Basin. Since the PETM is considered a regional datum (Denison, 2021), this comparison shows that there was no delayed transient uplift in the Froan Basin compared to the FSB. Within the resolution of biostratigraphic data, it appears that maximum uplift and subaerial exposure was broadly time equivalent across the region.

The Y2 flooding event (between UY1 and UY2) formed between 55.5–55 Ma at the time of peak transient uplift in the Froan Basin. A similarly aged flooding above the Forties sandstone member is also present in the Moray Firth area (Thomas and Hartley, 2014). If dynamic topography resulted in vertical motion of several hundred meters between 55.5 and 55 Ma, how can an ~85 m relative sea-level rise between maximum regression of UY1 and UY2 be explained? One possibility is a combination of eustasy, compaction, and tectonic- and load-induced subsidence. Isostatic considerations suggest that the total load-induced subsidence of the UT3-UY1 wedge amounts to ~200 m (which equals to an average rate of 0.1 mm/yr for the ~2 Myr unit). Such a subsidence rate superimposed on a ~40 m eustatic sea level rise between 56 and 55 Ma (Figure 13) might be sufficient to explain the rapid transgression and flooding observed between the two units during long-lived regional uplift. Increasing load-induced subsidence in

combination with post Cretaceous tectonic subsidence could also explain the gradual shift from overall forced regression during UY1 towards higher accommodation during UY2, to aggradation and final backstepping at the end of UY3.

Flooding of UY1 and higher accommodation during deposition of UY2 demonstrates that regional subsidence commenced between 55.5 and 54.5 Ma. This subsidence coincides with transgression and backstepping of Ypresian shorelines in the FSB, which commenced at 54.8 Ma (Hardman et al., 2018), suggesting that increasing accommodation likely was related to regional subsidence.

After deposition of UY2, accommodation increased significantly on the shelf, preventing large volumes of sediment reaching the toe-of-slope. The data show three phases of regression within the *Dracodinium simile* zone (between 54.5 Ma and prior to final flooding at 51 Ma), suggesting that UY3 is largely time equivalent with the Halibut Delta in the Moray Firth area (Zimmer et al., 2019). The eustatic curve shows only one sea-level fall during this period (between 53–52 Ma) but Conway-Jones and White (2022) also recognize two phases of uplift in the ESB (“S” and “T” landscapes), suggesting a possible common control across this area.

Mudge et al. (2014) showed that the Ypresian transgression occurred at ~54 Ma along the East Shetland Basin, which coincides with abandonment and condensation in the distal part of the system in the Gimsan and Froan Basin (Figure 13). In the more proximal area represented by well 66407/12-2, however, sedimentation continued until ~51 Ma, when the last sand was deposited, and the shoreline finally stepped back (onlap of Y7) (Figure 3). Final flooding in the proximal part of the Froan basin at ~51 Ma correlates with transgression of the Halibut Delta, which drowned around 50.4 Ma (Zimmer et al., 2019), suggesting that flooding of shelf-slope wedges occurred concurrently across larger areas of the North Atlantic.

Comparison between surfaces and events along the North Atlantic margin suggests a common forcing for several of the stratigraphic events. Yet, the areas also show significant discrepancies in the age of unconformities, stratigraphic stacking patterns and phases of sand delivery to the shelf margins. Lovell (2010) and Conway-Jones and White (2022) suggested that repeated uplift and subsidence along the ESB reflect fluctuation in mantle plume activity causing regressions and transgressions on the scale of hundreds of meters. Although several transgressive-regressive phases of similar age are present in the Froan Basin, the presence of local flooding surfaces indicates that the preservation of tectonic signals like the 2–3 Myrs fluctuations suggested here may be modulated by local controls. These discrepancies could reflect variable distance to the area of dynamic uplift or basin configuration (Friedrich et al., 2018), but could also reflect sampling and data bias, or local controls such as variations in subsidence and sediment supply. Thus, even if the margins were exposed to the same tectonic forcing, a non-unique stratigraphic response may have resulted across the region (Burgess and Prince, 2015).

Stratigraphic expression of the PETM in the Froan Basin

The short-lived PETM climate event lasting 100–200 kyr is known to have been associated with significant global warming (Sluijs et al.,

2007), changes in the hydrological cycle and increasing sediment supply to North Atlantic basins (e.g., Dypvik et al., 2011; Kender et al., 2012; Pogge von Strandmann et al., 2021; Rush et al., 2021; Stokke et al., 2021; Jin et al., 2022).

In the Froan Basin the PETM is constrained by the presence of *A. augustum* in a ~54 m thick succession in 6407/12-2 (Figure 3), a ~20 m thick succession in 6407/11-1 (Figure 6), decreasing to 10–20 m in more distal wells (Figure 6). The lithology of the PETM succession in the Froan Basin is typically non-calcareous, grey, green to brown claystone, locally grading to siltstone with streaks of sandstone (Supplementary Table S1). In log data, the PETM is expressed by an abrupt increase in GR values followed by a gradual decline (Figure 6), similar to what has been described elsewhere in the North Sea (Vieira and Jolley, 2020).

Figure 14 shows that the ~20 m thick PETM section in 6407/11-1 correlates with the two low-sinuuous channel complexes at the lower part of UY1, NW and SE of the well location. The FDC map demonstrates that these channel complexes are connected to the basin-floor fan in the Gimsan Basin and that they can be traced landward across the toe-of-slope aprons towards the shelf break where they terminate. The map thus documents the shallow to deep-marine depositional expression of the PETM in the Froan Basin.

Although the PETM often is linked to a period of eustatic sea-level rise (Sluijs et al., 2008; Harding et al., 2011), the climate event coincides with a period of forced regression in the Froan Basin and the only indication of base-level rise is at the top of the shallow-marine PETM succession (Figure 7C, section B). One possible explanation for this apparent discrepancy from other global PETM locations is that the amplitude of eustatic sea-level rise was too low to be expressed by onlap in seismic data (i.e., below seismic resolution). Another, partly complementary explanation, is that the rate of dynamic uplift balanced the rate of eustatic sea-level rise during the main phase of the PETM. We prefer a model where rapid dynamic uplift combined with high sediment supply outpaced eustatic sea-level rise during the main phase of the PETM, resulting in net forced regression. However, it is possible that the indication of base-level rise observed in Figure 7C, section B marks the waning phase of the PETM when the sediment supply rate decreased. Interestingly, a maximum flooding surface has also been interpreted above a sand-rich PETM succession in the deep-marine part of the North Sea (Vieira and Jolley, 2020; Jolley et al., 2022).

The seismic data show that the shoreline was at the shelf break and that the fluvial system delivered sediment directly to the upper slope during the PETM. The sediments delivered to the shelf-edge delta were primarily transported down-slope in up to 400 m wide and erosive channel complexes (Figure 7D, section A). When reaching the toe-of-slope, the erosive PETM channel complexes started to deposit sediment locally, but parts of the sediment load bypassed the toe-of-slope and continued toward the Gimsan Basin in two depositional channel complexes. The development of such wide and long channel complexes is in stark contrast to the pre- and post-PETM situation when the slope was dominated by narrow (up to 100 m wide) channels terminating near the toe-of-slope.

Cross-sections through these channel complexes show that they change morphology basinward. As they exit the toe-of-slope aprons (Figure 14, section A) they are typically 0.7 km

wide, up to 80 m thick and show only minor erosion at the base. They have no pronounced levees and show considerable differential compaction, suggesting that they are sand-prone (Jennette et al., 2000; Bryn and Ackers, 2014; Edwards et al., 2018). Internal reflectivity and lack of deep incision at the bases show that they are overall aggradational. Farther basinward (Figure 14, section B), the channel complexes are wider (up to 1 km) and thinner (up to ~50 m). They show less internal geometries and less differential compaction compared to the more proximal location, indicating that they are less sand prone. Toward the Bremsein Fault Complex (Figure 14, section C), the channel complexes are up to ~1.8 km wide, ~30 m thick and have no internal seismic character. They show subtle differential compaction, but significantly less than in the more landward parts, indicating lower sand content. The basin gradient decreases basinward of the Bremstein Fault Complex and this transition marks the channel-lobe transition zone where the channel complexes become distributary as they deliver sediment to the basin-floor fan in the Gimsan Basin. The basin floor fan is expressed by a single seismic reflection and has no measurable thickness and show no indications of differential compaction. Wells penetrating the distal part of the basin-floor fan (e.g., 6407/8-1, Figure 6) show that the PETM succession typically is <20 m thick and dominated by claystone and minor siltstone (Supplementary Table S1).

Based on these observations it seems that most of the coarse material supplied to the deep-water during the PETM was deposited in the toe-of-slope aprons and in the proximal part of channel complexes landward of the Bremstein Fault Complex. Only the fine-grained fraction was transported >50 km into the Gimsan Basin. However, there was not sufficient time and/or volume of sediment to develop thick levees along the channel complexes, pointing to an abrupt and short-lived pulse of sediment delivery to the distal basin during the PETM.

A transition from narrow (<100 m wide) channels terminating at the toe-of-slope before the PETM, to wide and erosive slope channel complexes feeding sediment more than 50 km into the basin during the PETM support a significant change in flow properties during the climate excursion. With time, deep-water channels tend to develop towards a state of equilibrium characterized by sinuous channel morphology and decreasing width and depth down-stream (Kneller, 2003; Pirmez and Imran, 2003; Shumaker et al., 2018). Wider channel complexes during the PETM suggest that gravity flows were larger and more erosive (Konsoer et al., 2013; Ferguson et al., 2020), and it is likely that they were denser and had more mud entrained in the flow in order to travel longer distances (Baas et al., 2004; Traer et al., 2018; Pope et al., 2022). A fundamental shift in depositional style to larger channels and lobes, combined with increased transport distance of clastic material observed in the Froan and Gimsan basins, is thus in line with recent work from other locations in the North Atlantic where the PETM was associated with higher onshore weathering rates, increased erosion and delivery of sand and clay to the marine realm (Pogge von Strandmann et al., 2021; Rush et al., 2021; Jin et al., 2022). As the fluvial system delivered sediment directly to the shelf edge during the PETM, more seasonal variations in discharge (Eldrett et al., 2014; Jolley et al., 2022) could have resulted in rapid deposition and slope failure on the upper slope together with larger and more frequent hyperpycnal flows

(Petter and Steel, 2006; Henriksen et al., 2011), triggering the inferred change in gravity-flow size and composition inferred during the PETM. However, in contrast to Jin et al. (2022) who demonstrated increased sand delivery to the deep-water North Sea basin, the Froan example documents increased mud-delivery to the distal deep-water Gimsan basin during the PETM. A similar transition toward more mud-rich deposits during the PETM has also been described from deep-water settings in the Pyrenees (Pujalte et al., 2015) and Gulf of Mexico (Sharman et al., 2022).

Jin et al. (2022) also documented an order of magnitude increase in sedimentation rates during the PETM in the North Sea from well data. When calculating sedimentation rates for wells 6407/12-2 and 6407/11-1 (Supplementary Table S2), both wells similarly show an increase in sedimentation rate from 8–75 m/Myr before the PETM, to 270–350 m/Myr during the PETM, and back to <140 m/Myr after the event. This variability supports the idea that the climate excursion was associated with an order-of-magnitude increase in sedimentation rates when measured in one-dimensional well data. However, when comparing volumetrics from seismic units it becomes clear that the change in sediment supply across the PETM was less dramatic (Supplementary Table S3). When we use the PETM thickness maps in Figure 5 and assume a PETM thickness of 10–20 m in the distal basin (where the interval is below seismic resolution), and a duration of 100–200 kyr, we can calculate high and low sediment supply scenarios before, during and after the PETM (Supplementary Table S3). The results show that the transition from UT3 to the PETM was associated with up to an order of magnitude increase in sediment supply, from 46 km³/Myr during deposition of UT3 to 185–470 km³/Myr during the PETM. After the PETM, sediment supply decreased to 111 and 153 km³/Myr during UY1 and UY2, respectively. There are significant uncertainties associated with these numbers since the units have internal unconformities and as parts of the sediment budget may not be accounted for, but the overall trend is that although the transition into the PETM was associated with up to an order of magnitude increase in sediment supply, the sediment supply immediately after the hyperthermal was only deduced by 50%–75%. One possibility is that the order-of-magnitude increase in sedimentation rate observed in well data might be biased towards local areas of active deposition and/or a result from averaging over short time intervals (Schumer and Jerolmack, 2009). Another key factor is that the PETM in the Froan Basin occurred during forced regression, when shelf accommodation was lower compared to the preceding UT3 unit. This transition from normal to forced regression shifted the depocenter basinward and the full volumetric extent of the unit is thus better preserved in the rock record. We suspect that the real increase in sediment delivery to the Froan Basin during the PETM was more moderate (2–4 times the pre- and post PETM supply), and that the order-of-magnitude values identified in individual wells may not be indicative of changes in sediment supply for an entire source-to-sink system during such rapid climate events.

From the discussion above it is clear the PETM response in well data is highly dependent on the sample location along the sediment routing system and the local interplay between accommodation and sediment supply. In the North Atlantic, PETM eustatic sea-level rise appears to have been balanced by rapid dynamic uplift in many locations, resulting in overall regression and deposition of a sand-

rich unit in areas that were already dominated by basin-floor fan deposition (Vieira and Jolley, 2020; Jin et al., 2022). The expression of the PETM climate signal is further dependent on the sampling location along the routing system. The sand-dominated PETM succession sampled by Jin et al. (2022) was in a medial to distal position of an older sand-rich basin floor fan system and it is possible that a well in a more distal position will capture a more mud-rich PETM signal as observed in well 6407/8-1 in the Froan Basin (Figure 14). Similarly, a more proximal sampling location along the channel complexes and toe-of-slope deposits in the Froan Basin would likely have recorded a thicker and more sand-rich PETM succession. The proximal, but off-axis position of well 6407/12-2 captures a relatively thick, silty, and mud-rich PETM succession. The integration of seismic geomorphology is therefore key to understand the system-wide context of one-dimensional well data and the response of an entire source-to-sink system to rapid climate change.

Sediment partitioning—External controls

Sediment partitioning, stratigraphic architecture and the spatial-temporal link between shallow and deep-water deposits reflect the interplay between external controls on accommodation and sediment supply, autogenic processes, and local environmental processes (Gawthorpe et al., 1994; Madof et al., 2016; Straub, 2019; Paumard et al., 2020). In supply-dominated settings, sediment dispersal to the basin floor can occur throughout the sea-level cycle (e.g., Carvajal and Steel, 2006), especially in greenhouse systems where shelves may have been relatively narrow compared to icehouse systems (Burgess et al., 2022).

Our data analysis from the Froan Basin demonstrate that long-term transient uplift and subsidence drove the large-scale partitioning of sediments along the margin, whereas short-lived eustatic fluctuations, isostasy and environmental processes controlled spatial changes in sedimentation. During the initial phase of normal regression (Figure 4), sediments were largely trapped locally in topsets and proximal foresets (Figure 15), expressed by high proximal sedimentation rates (up to 350 m/Myr) and rapid basinward thinning of foresets during UT3. Sediment partitioning changed at the onset of forced regression when the shelf-break trajectory fell and when there was no accommodation landward of the advancing shoreline (UY1). Lack of accommodation resulted in thick foresets and bottomsets, aggradation and progradation of toe-of-slope aprons across a larger area and with high sedimentation rates (up to 200 m/Myr). The observed shift in sediment partitioning is similar to previous studies (Porębski and Steel, 2003; Johannessen and Steel, 2005), showing that basin floor fans (or toe-of-slope aprons) can develop in front of supply-dominated shelves even with rising shelf-break trajectories (Helland-Hansen and Martinsen, 1995), but that those fans tend to be thinner and have shorter run-out lengths compared to the ones developed when the trajectory is flat or falling (Carvajal and Steel, 2006; Koo et al., 2016). As accommodation increased during UY2 and UY3, sediments were trapped on the shelf. UY3 covers around 3–4 Myrs of deposition and the resulting sedimentation rates on the shelf are thus low (up to 50 m/Myr), suggesting that sediment accumulated landward of the shelf, possibly combined by an overall lower sediment supply.

Sediment partitioning—Environmental controls vs. autogenic processes

Within such a framework of tectonically controlled sediment partitioning during normal regression, forced regression and aggradation, local autogenic processes also controlled sediment distribution along the margin. Figure 7I, section A shows that each successive phase of clinothem progradation stopped near the shelf break established during maximum regression of UY1. The concept of autoretreat suggests that increasing space in front of the advancing delta eventually will cause backstepping (Muto and Steel, 2002), but the seismic geomorphology also shows that the process regime changed as the delta of UY2 approached the shelf break. Northward sediment transport by littoral currents removed parts of the sediment from the advancing shoreline, preventing further progradation. The littoral drift became stronger as the delta approached the shelf break (Dixon et al., 2012), evident from the change in clinothem progradation direction near the shelf break (Figures 7E,F). Although the point of autoretreat can be determined by a suite of intrinsic variables (Burgess et al., 2008), we suggest the process of sediment removal by longshore drift may have been just as important as increasing accommodation in partitioning sediment across and along the shelf during each phase of clinothem progradation.

Sediment partitioning—Processes regime

Figure 12 shows how the slope gradient changes along strike at the top of UY2 from 1 to 3 degrees in the area of direct fluvial supply in the south to as much as 7 degrees northward where the margin was wave-dominated and where sediment was supplied by longshore drift. Here the slope and toe-of-slope is dominated by mass wasting and local development of small lobes in front of canyons fed by longshore drift (Figure 7D). The toe-of-slope in the northernmost part of the study area shows few signs of gravity deposits (Figure 7G), indicating that most of the sediment transported by longshore drift was captured by canyons and gullies before reaching the northernmost part of the shelf. As documented in several other areas, the change in process regime along strike shows that the fluvial-dominated shelf-edge environment in Froan was more efficient in bypassing sediment to the toe-of-slope than the wave dominated part of the systems (Dixon et al., 2012; Cosgrove et al., 2018; Paumard et al., 2020). This along-strike change in gradient and mechanism of sediment delivery from the shelf to the deep-water is similar to the two modes of shelf margin development described in Fisher et al. (2021): out-of-grade margins are typically associated with sediment delivery through shelf-incised canyons resulting in slope onlap and basin floor aggradation, whereas direct-fed margins often show more unconfined sediment delivery to the slope and basin floor causing progradation along the entire shelf-slope system (Hadler-Jacobsen et al., 2005). In the Froan Basin these two modes of system development occurred along strike within the same system. This along-strike variability was facilitated by a change in process regime in which longshore transport drove shelf progradation northward, resulting in steepening of the slope and initiation of canyon and gully incision capturing parts of the littoral drift and supplying it down-slope. In the case of the Froan Basin, along-strike variation in slope gradient is thus an intrinsic control and not related to large-scale basin or margin configuration.

Canyons as sediment conveyers during highstand and lowstand

Submarine canyons fed by littoral drift in modern systems show that they tend to be most active during sea-level rise or highstand, when vigorous longshore currents develop, and that they often are disconnected from sediment supply during lowstands (Covault et al., 2007; Bernhardt and Schwanghart, 2021). Supply to canyon heads during transgression and highstand conditions may also have occurred in the Froan Basin, but in addition, the change in clinof orm progradation during deposition of UY1 and UY2 shows that longshore transport must have been prominent also during forced regression. Such an interpretation is backed by a simple mass balance comparison showing that the volume of sediment stored in the lobe in front of the northern incision in Figure 11 is $\sim 3.8 \text{ km}^3$, whereas the volume missing within the incision is only 0.008 km^3 (Figure 11F). This calculation shows that the dendritic incision itself cannot have supplied the material to the lobe and that longshore drift must have been the main source of sediment for the toe-of-slope lobe deposits.

The preferred model is one where sediment supplied from longshore currents was trapped and conveyed down slope during forced regression when dendritic incisions and incipient valleys formed along the shelf (Figure 11). Since canyons did not extend far across the shelf, the amount of sediment delivered by longshore drift during transgression was likely relatively low. The key mechanism for highstand feeding of canyons in modern systems is the presence of deep, shelf-indented canyons that can bypass sediment from the highstand shoreline. In the Froan Basin UY1 canyons were largely backfilled during transgression and UY2 canyons did not extend far across the shelf, suggesting that these were not important sediment conveyers during highstands.

Conclusion

Integration of well and chronostratigraphic data together with high-resolution seismic geomorphology have allowed us to investigate the interaction between long-term tectonic uplift, short-lived PETM climate change, and more intrinsic environmental processes in six Selandian-Ypresian seismic units in the Froan Basin in the Norwegian Sea.

Transient uplift resulted in a relative sea-level fall of at least 300 m, accompanied by forced regression, valley incision and the formation of rapidly prograding shelf-edge deltas and toe-of-slope aprons. Despite subaerial exposure and valley incision $\sim 0.5 \text{ Myr}$ after the PETM, deep-water toe-of-slope aprons were deposited within 15 km from the shelf break at maximum regression and deep-water systems did not reach the distal Gimsan Basin. In contrast, the PETM climate event was associated with wide bypass channels on the slope, aggradational sandy channel complexes on the proximal basin floor, and the deposition of a mud-rich basin floor fan in the distal part of the basin $>50 \text{ km}$ from the shelf break. This fundamental shift in depositional style indicates that the hyperthermal event was associated with larger and more mud-rich gravity flows at the shelf edge, probably triggered by higher and more seasonal variable runoff in the fluvial system. This shows that short-lived climate changes can have greater impact on the architecture (width, thickness and run-out distance of channels and lobes) of deep-water sedimentary systems than basinward shift of accommodation during periods of sea-level lowstands. In contrast to many other locations around the world where the PETM occurred

during transgression, there was no flooded shelf to buffer the climate signal in the Froan Basin. The climate signal was therefore efficiently transferred from the fluvial to the deep-marine part of the system.

Our analysis further shows that the morphology of incised valleys formed at the time of maximum regression between ~ 55.5 and 55 Ma reflects the proximity to the sediment source. Where sediment was fed directly to shelf-edge deltas, the slope gradient was lower and no valleys, canyons or gullies formed. Progradation of the northern part of the shelf was driven by longshore sediment transport. Here the slope was steeper, resulting in the formation of erosional networks and incised valleys which funneled longshore-fed sediment down-slope, suggesting that incised valleys and canyons were not the main conveyor of sediment between the terrestrial and deep-marine part of the system.

The studied shelf-slope succession in the Froan Basin is the only example of a Paleogene wedge with well-preserved shorelines and shelf-edge geometries along the Norwegian continental shelf. Timing of subaerial exposure and phases of shoreline advance share many similarities with the East Shetland Basin and Faroe-Shetland Basin, pointing to a regional common driver for key stratigraphic events in these basins. The Froan Basin is thus a key datapoint for understanding the regional stratigraphic response to tectonic uplift in the North Atlantic region.

Data availability statement

The data analyzed in this study is subject to the following licenses/restrictions: The datasets used in this study (PGS18M05 and PGS17M05) is not publicly available and is proprietary to PGS. Requests to access these datasets should be directed to <https://www.pgs.com/>.

Author contributions

TS designed the study, conducted the main analysis and wrote the bulk of the manuscript. PS, SH, and JS were involved in data interpretation, discussion and manuscript writing. OM and LV were involved in discussion and manuscript writing.

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Conflict of interest

Authors TS, SH, QM, PS, JS, were employed by the company Equinor ASA.

The remaining author declares that the research was conducted in the absence of any commercial or financial relationships that could be construed as a potential conflict of interest.

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Supplementary material

The Supplementary Material for this article can be found online at: <https://www.frontiersin.org/articles/10.3389/feart.2023.1082203/full#supplementary-material>

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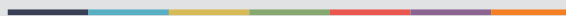
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