



Crustal structure of the North Svalbard margin: Continental breakup and Eurasia Basin opening

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ABSTRACT

Continental breakup at the North Svalbard margin and the northern Barents Sea during the Paleogene led to the opening of the oceanic Eurasia Basin. However, the mechanisms behind this rift-to-drift evolution remain unclear. Here, we present seven crustal transects (CT1–7) integrating 2D forward modelling of potential field data and constrained by structural interpretation of multichannel seismic reflection profiles. These transects extend from the continental regions of North Svalbard to the oceanic domain in southwest Eurasia Basin. The Yermak Plateau comprises two juxtaposed tectono-magmatic blocks: the Northeast Yermak Plateau, linked to pre- or early Eureka settings, and the Southwest Yermak Plateau, formed by shearing and oblique extension along the West Svalbard margin. Broadly, the North Svalbard margin is divided into three segments: (1) Northwest Spitsbergen-Southwest Yermak Plateau, (2) Northeast Spitsbergen-Northwest Kvitøya, and (3) Northeast Kvitøya-North Barents. Segmentation follows inherited north-northwest trending weak zones, guiding Paleogene multi-phase oblique rifting and formation of the Sophia Basin, underlain by thinned, high-density metamorphic crust. The results indicate a rift-shear breakup mode with exhumation of continental lower crust and/or serpentinized subcontinental mantle blocks within the continent-ocean transition (COT) that precedes seafloor spreading which started at ca. 53 Ma during magnetic chron C24. The continental breakup processes that led to margin segmentation reflect the interplay between inherited basement fabrics and plate opening directions of the Eurasia Basin region.

1. Introduction

Structural inheritance plays a key role in shaping the dynamics and morphology of continental breakup, influencing the type and evolution of passive margins (e.g. Schiffer et al., 2020). Structural inheritance refers to the role of the pre-existing geology, tectonics and related anisotropies within the crust and mantle, often derived from ancient orogenic belts, on the direction and mechanics of later tectonic events (e.g., Vauchez et al., 1998; Tommasi and Vauchez, 2001; Brune, 2014; Zwaan et al., 2024). The interaction between these inherited structures and tectonic forces leads to diverse breakup mechanisms and complex

opening modes, including oblique extension and strike-slip deformation (e.g., Basile, 2015; Nemčok et al., 2016; Le Pourhiet et al., 2017; Peace et al., 2018; Balázs et al., 2023) that often result in segmented rift patterns as the crust accommodates various stress regimes (e.g., Brune, 2014; Zwaan et al., 2024). In the context of the Northeast Atlantic towards the Barents Sea and Svalbard continental margins, inherited Caledonian and Timanian basement fabrics, as inferred by previous studies (e.g., Breivik et al., 2005; Marello et al., 2013; Klitzke et al., 2019; Shulgin et al., 2020; Gabrielsen et al., 2022), has guided the rifting and subsequent continental breakup process. The geodynamic setting prior to continental breakup at the northwest corner of the Barents Shelf

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possibly consisted of a triple junction between the Eurasian, Greenland, and North American plates during Late Cretaceous–Paleocene (66–56 million year ago, Ma, e.g., Feden et al., 1979; Brozena et al., 2003; Gaina et al., 2009; Kristoffersen et al., 2020). This setting indicates that continental breakup should have occurred under conditions of oblique extension and shearing (e.g., Minakov et al., 2012a; Minakov et al., 2013). However, the role of pre-existing structures within the vicinity of the northern Svalbard margin and their controls on the final mode of opening of the active the Eurasia Basin are not well understood.

The Svalbard and northern Barents Sea regions north of $\sim 80^\circ\text{N}$ have been the target for extensive geological and geophysical studies during the last decades (e.g. Jokat and Micksch, 2004; Geissler and Jokat, 2004; Engen et al., 2009; Minakov et al., 2018). One of the least understood events concerning the North Svalbard region is the Paleogene continental breakup between the Barents–Kara shelves and the Lomonosov Ridge (Minakov et al., 2012a; Faleide et al., 2024a; Abdelmalak et al., 2024a; Abdelmalak et al., 2024b) occurring at ~ 56 –53 Ma, an age frame constrained by the oldest unambiguous seafloor spreading anomaly (i.e., Chron C24; Brozena et al., 2003; Glebovsky et al., 2006). Continental breakup was followed by slow seafloor spreading at the mid-ocean Gakkel Ridge (e.g., Nikishin et al., 2018). A region underlain by serpentinized mantle north of Barents Sea shelf has been inferred from multichannel seismic reflection data (e.g., Lutz et al., 2018). The present-day ultra-slow Gakkel Ridge exhibits distinct segmentation along the spreading axis into magmatic (basaltic) and amagmatic

(peridotite-dominated) crust areas (Michael et al., 2003). Magnetic anomaly distribution and amplitude variations characterize the Eurasia Basin and might indicate that this segmentation started earlier. However, it is not known what controls the initial segmentation and how stable it was through time.

In this contribution, we study the continental and oceanic basin architecture and crustal structure of the offshore region of North Svalbard using an integrated workflow with seismic, gravity, and magnetic data. Geophysical data is combined with geological observations to model a series of crustal-scale transects covering the study area. We identify different margin segments and explore pre-existing structures within the continental crystalline basement for understanding their possible role in the dynamics of breakup and subsequent opening of the Eurasia Basin.

2. Regional setting

The North Svalbard margin is confined between the Nansen Basin to the north and the northern prolongation of the Northeast Atlantic to the west (Fig. 1). Geological and geophysical information offshore north and northeast of Svalbard is rather limited because of the challenging conditions to collect data in this polar region. Studies conducted in more accessible areas onshore Svalbard and in the ice-free waters to the south have provided some insights into the evolution of the northern Barents shelf. Several integrated geophysical studies have indicated that the crystalline basement fabric is oriented in E–W direction in the east

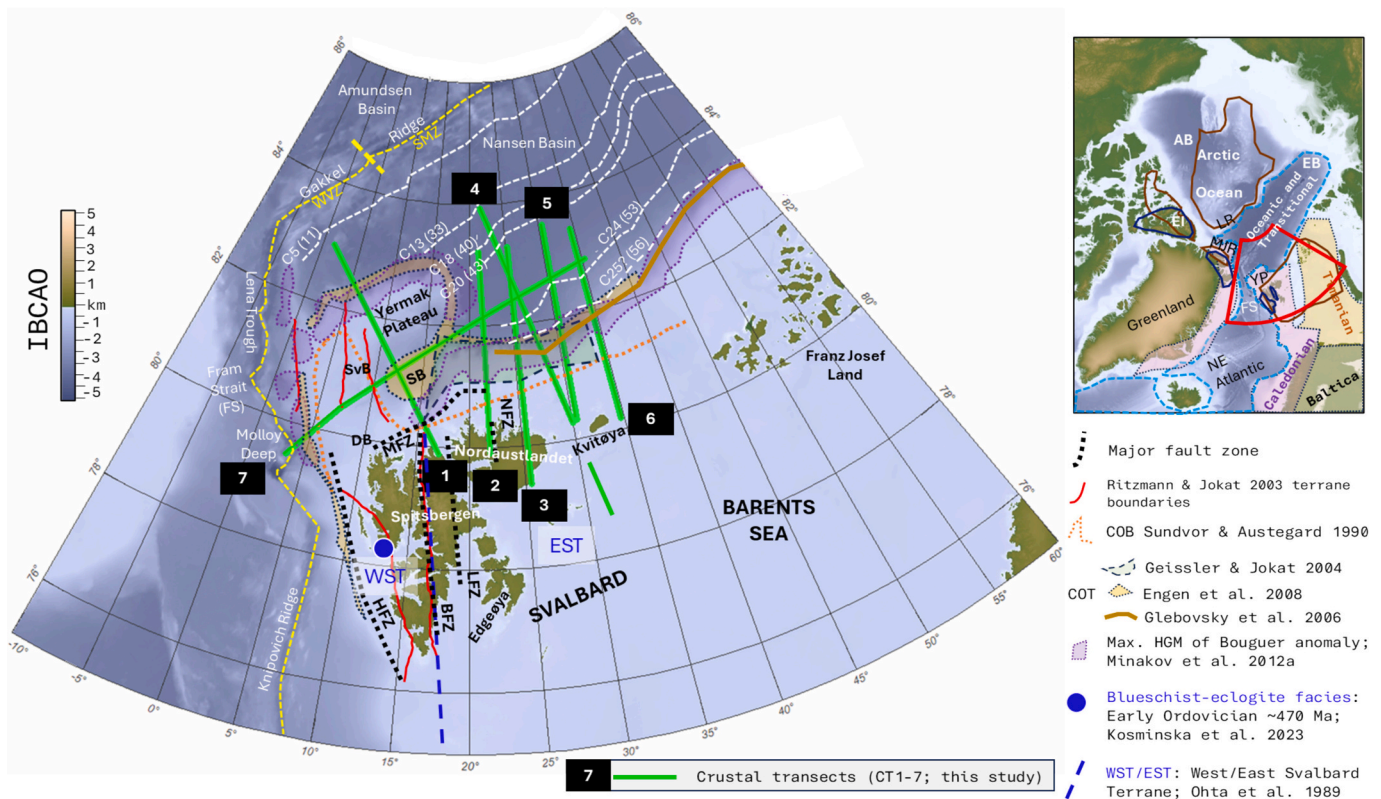


Fig. 1. Overview map of the study area in the North Svalbard continental margin. The modelled crustal transects (CT1 to CT7) are indicated by thick light green lines. Main structural elements are indicated. Seafloor isochrons (with age in million years within brackets) in the southwest Eurasia Basin are interpolated after (Glebovsky et al., 2006), including a lineation interpreted as C25 by (Brozena et al., 2003). Gakkel Ridge western volcanic/sparsely magmatic zones (WVZ/SMZ) after (Michael et al., 2003). The base map is the IBCAO bathymetry (v. 4.0; Jakobsson et al., 2020). Note that the island of Nordaustlandet has a (horizontal) length of ~ 170 km. BFZ, Billefjorden Fault Zone. COB, continent–ocean boundary. COT, continent–ocean transition. DB, Danskøya Basin. HFZ, Hornsund/De Geer Fault Zone. HGM, high gradient magnitude of the Bouguer gravity anomaly. LFZ, Lomfjorden Fault Zone. NFZ, Nordaustlandet Fault Zone. SB, Sophia Basin. SvB, Sverdrup Bank. Inset map in the top right displays the NE Atlantic–Arctic regional hypsometry, with main tectonic Caledonian and Timanian trends (Breivik et al., 2005; Faleide et al., 2018). The dark-brown and dark-blue envelopes indicate areas affected by the High-Arctic Large Igneous Province and Eureka deformation, respectively (Estrada et al., 2024; Meier et al., 2024). The red polygon indicates the location of our study area. AB, Amerasia Basin. EB, Eurasia Basin. EI, Ellesmere Island. FS, Fram Strait. LR, Lomonosov Ridge. MJR, Morris Jesup Rise. YP, Yermak Plateau. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

Barents Sea, and NNE-SSW in the north and west Barents Sea (Breivik et al., 2005; Breivik et al., 2002; Ritzmann and Faleide, 2007; Marelllo et al., 2010; Marelllo et al., 2013; Klitzke et al., 2019; Shulgin et al., 2020; Schiffer et al., 2020; Gabrielsen et al., 2022) (Fig. 1). These trends have been linked to the Timanian (~530–510 Ma) and Caledonian (~490–390 Ma) orogenies, respectively (Fig. 1). This classification is more diffuse towards the north of these regions because of possible overlap of these two deformation events in the Svalbard and Barents Sea regions (e.g., Abdelmalak et al., 2024a). For instance, a major N-S-trending Caledonian deformation is postulated across Spitsbergen through the Billefjorden Fault Zone separating a strongly metamorphosed and deformed western terrane from a less deformed one to the east (i.e., West/East Svalbard Terrain, Fig. 1; Ohta et al., 1989; Koglin et al., 2022; Kościńska et al., 2023). These terrains behaved as independent mobile tectonic belts until they collided to each other during the Caledonian orogeny (~425 Ma; Cocks and Torsvik, 2011). The West Svalbard Terrain shows signs of greater degrees of deformation with amphibolite and blue-schist metamorphic facies as remnants of Iapetus Ocean subduction (e.g., Ohta et al., 1989) (Fig. 1). The East Svalbard Terrain includes the areas of Northeast Spitsbergen (i.e., Ny Friesland), Nordaustlandet, and Kvitøya, and consists of metamorphic facies of lesser deformation intensities (e.g., meta-sediments and greenschists; Johansson et al., 2005; Michalski et al., 2023).

The area east of Kvitøya towards Franz Josef Land, instead, represents another major weak zone that delimits Caledonian domain from a Caledonian-overprinted Timanian basement (Cocks and Torsvik, 2011; Marelllo et al., 2013; Faleide et al., 2018; Abdelmalak et al., 2024a; Fig. 1). These geological structures (Caledonian and Timanian) existed before the initiation of the Arctic Ocean basin (including the older Amerasia Basin) in mid-late Mesozoic times (Faleide et al., 2018) (Fig. 1). During the Cretaceous, widespread and multi-phase regional volcanism and magmatism occurred (i.e., High-Arctic Large Igneous Plateau, HALIP; Heyn et al., 2024 and references therein). In particular, magma emplaced within Ellesmere Island, Northeast Greenland, Svalbard, and northern Barents Shelf, partly followed pre-existing shear zones, but also intruded across the existing structural grain (Minakov et al., 2018; Minakov et al., 2012b) (Fig. 1). Uplifted areas and compressional domes formed across the Barents Shelf throughout the Early Cretaceous (Kairanov et al., 2018) and later during the Paleogene Eureka deformation (Piepjohn et al., 2016; Gac et al., 2020; Fig. 1), which were often nucleated at older structures within the crystalline basement (Kairanov et al., 2018).

The Yermak Plateau is a prominent bathymetric structure that most likely represents the continuation of the North Svalbard continental margin into the Eurasia Basin east of ~15°E, including the Sophia Basin in the west (Fig. 1). These structures are difficult to kinematically restore in time-dependent rigid plate reconstruction of the western Eurasia Basin (e.g., Kristoffersen et al., 2025), as their origin is debated. The Yermak Plateau is suggested to be the result or overprinted by Late Cretaceous-Paleogene magmatism, rifting in the Eurasia Basin, and alternating transpression/transension between the Greenland, North American, and Eurasian lithospheric plates near their triple junction north of the study area (Brozena et al., 2003; Riefstahl et al., 2013; Kristoffersen et al., 2020). The Sophia Basin (Fig. 1), a semi-enclosed bathymetric depression, contains a relatively thick pile of sediments (~4 km; e.g., Faleide et al., 2024a; Geissler et al., 2011; Geissler and Jokat, 2004), and its origin might be related to some of the tectonic event(s) that formed and modified the Yermak Plateau (Geissler and Jokat, 2004; Engen et al., 2008; Abdelmalak et al., 2024b). The North Svalbard margin to the east faces directly against the Cenozoic Eurasia Basin (Fig. 1).

Earlier studies on the basis of seismic reflection data initially defined structural constraints delimiting the continental, transitional (i.e., the region between the shelf break and continental slope, within bathymetric regions of ~1500–2500 m below sea level), and oceanic crust domains (i.e. at the continent-ocean boundary, COB) as the difference

between the elevated region bordering the southwestern parts of the Yermak Plateau and the deeper Eurasia Basin (Sundvor and Austegard, 1990) (Fig. 1). The COB was later redefined and extended as a continent-ocean transition (COT) region, marked by either its landward limit at the Moffen Fault Zone or as an area projected northwards from that fault feature (Geissler and Jokat, 2004; Glebovsky et al., 2006; Engen et al., 2008) (Fig. 1). These and subsequent observations on the basin and crustal structure character using seismic reflection data defined the North Svalbard facing the Eurasia Basin as a magma-poor margin (Geissler and Jokat, 2004; Minakov et al., 2012a; Lutz et al., 2018). The COT zone extent was later refined as a narrow ~50–100 km-wide region (Fig. 1) on the basis of crustal and geodynamic modelling (Minakov et al., 2012a, 2013).

The two subbasins of the Eurasia Basin, the Nansen and Amundsen oceanic basins, flank each side of the mid-ocean Gakkel Ridge (Fig. 1). The two sub-basins of Eurasia Basin formed by seafloor spreading following continental breakup between the Lomonosov Ridge and the northern margin of Barents and Kara seas. Crustal thickness estimates based on refraction data are sparse in the Eurasia Basin, with some studies postulating that the oceanic crust is probably no thicker than ~5 km in the intermediate to deep ocean regions (~2000 to 4000 m below sea level) of the Nansen Basin (Jackson et al., 1982; Duckworth and Baggeroer, 1985; Geissler and Jokat, 2004; Schmidt-Aursch and Jokat, 2016). The interpretation of magnetic isochrons suggests variable spreading rates at the Gakkel Ridge both spatially and through time but generally range between ~0.5–2.7 cm/yr full spreading rates (Glebovsky et al., 2006; Nikishin et al., 2018).

3. Data and methods

3.1. Input datasets and models

We used seismic reflection and potential field data, together with available seismic velocity and crustal models combined with stratigraphic constraints from existent boreholes to build crustal transects. The utilized compilation of two-dimensional (2D) multichannel seismic reflection data covers the North Svalbard margin, including both the Yermak Plateau and Sophia Basin, as well as the transitional and deeper basin regions into the Nansen Basin (Fig. 2). The seismic reflection database is a collection of marine expeditions into the Arctic Ocean over the last few decades (Faleide et al., 2024a). Newly released marine seismic data collected in this region includes data acquired by the Norwegian Offshore Directorate (Sodir) for the UN Law of the Sea program, seismic lines by Federal Institute for Geosciences and Natural Resources (BGR), Alfred Wegener Institute (AWI), and results of the GoNorth-2022 expedition (collaborative efforts of several Norwegian research institutions and Geological Survey of Denmark and Greenland, GEUS; <https://gonortharctic.no/>) (Fig. 2).

The stratigraphic constraints for the offshore component of the North Svalbard margin are represented by three scientific boreholes drilled on the southern Yermak Plateau that yielded sediment records back to the late Miocene (~6 Ma; Ocean Drilling Program, ODP, sites 910C, 911 A, 912 A; Mattingsdal et al., 2014; and see also ~11 Ma, ODP site 909 within Molloy Deep region; Gruetzner et al., 2022 (Fig. 2). Magnetic anomaly and sonobuoy data in combination with seismic reflection profiles and known ODP stratigraphy provide maximum age relationships and velocities for the deeper, more consolidated units, and basement (e.g., Eldholm et al., 1984a; Myhre, 1984; Geissler and Jokat, 2004; Engen et al., 2009; Geissler et al., 2011) (Fig. 2). Gridded gravity and magnetic data (Fig. 3) covering the study area (MAGNOW, Glebovsky et al., 2006; DTU17, Andersen and Knudsen, 2020; US Naval Research Laboratory Arctic surveys 1998/99, NCEI/NOAA, 2024 and Brozena et al., 2003; Circum-Arctic mapping project, CAMP, Gaina et al., 2011), in addition to previous models for Moho topography and depth to lithosphere-asthenosphere boundary (ArcCRUST, Lebedeva-Ivanova et al., 2019; LAB, Klitzke et al., 2015, respectively) were used in

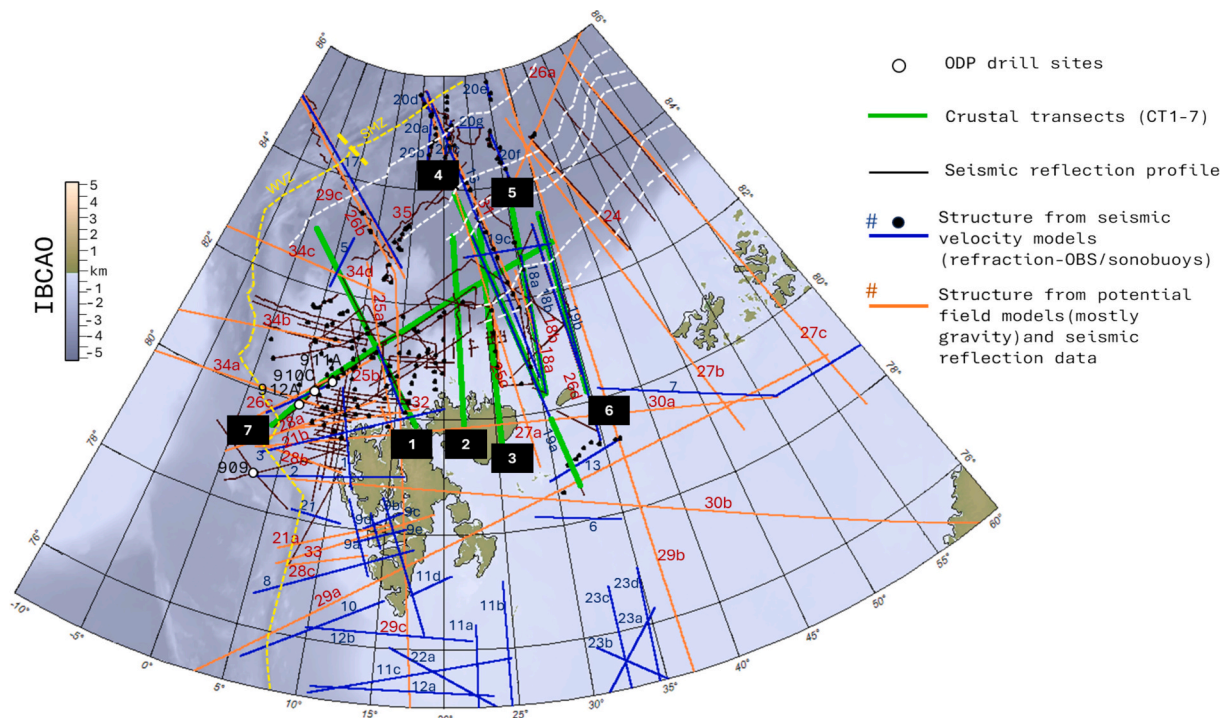


Fig. 2. Seismic reflection database used in this study. The location of 2D models of the crustal structure of the northern Barents and Svalbard are indicated in the map by thick light green lines. Crustal structure from seismic velocity models include refraction measurements from ocean bottom seismometers (OBS) and land stations, as well as sonobuoys (in black circles) deployed offshore during various marine expeditions into the Arctic (e.g., Faleide et al., 2024a). Crustal structure from potential-field models and seismic reflection data are also indicated. Numbers in lines for profile locations indicate the reference to the existing work (i.e. the “#” in legend; see Table 2 for complete list). Scientific shallow boreholes from Ocean Drilling Program (ODP) sites 912 A, 910C, and 911 A at the southern Yermak Plateau and site 909 at the Molloy Deep are indicated with white circles. Other abbreviations and map features as in Fig. 1. See Methods section for details. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

this work (Fig. 4). Constraints on crustal structure within the Svalbard, Barents Sea, and Eurasia Basin regions from previous work were also compiled in this study (see Fig. 2 and Table 2 for details). These include models derived from seismic velocities (i.e., travel-time modelling results from deep seismic refraction profiles and sonobuoys, and interval velocities used in reflection seismic studies), as well as models derived from integrated studies using potential field data (Fig. 2 and Table 2). This compilation served to establish velocity-depth curves required to translate seismic reflection profiles into crustal models (Fig. 5).

3.2. Construction and modelling of crustal transects

Seven crustal transects (CT1 to CT7; Fig. 1) have been constructed across (traversing the margin, six profiles) and along (parallel to margin, one profile) the North Svalbard margin coincident with the locations of existing seismic reflection profiles and available refraction data and regional models (Fig. 2 and Tables 1–2). Six of the crustal transects (CT1 to CT6) document the structure of the North Svalbard margin between 15°E and 35°E across 350–400 km-long and roughly N-S-trending sections, whereas one of the transects (CT7) was constructed in a direction subparallel to the COT over a 600 km-long crustal profile and approximately spanning from 5°E to 35°E and 80°N to 83°N. The utilized seismic reflection data provided structural information of the basin architecture, specifically, of key seismic reflectors: 1) the seabed, 2) the intra-sedimentary, and 3) the top acoustic basement (see also Supplementary figures).

3.2.1. Sediments and basement

Local differences between IBCAO bathymetry version 4 (Jakobsson et al., 2020) (Fig. 1) and the mapped seabed seismic reflector of up to ~500 m vertically exists due to uneven resolution of the bathymetry grid

on a scale of <5–10 km. The intra-sedimentary reflector is a boundary correlated based on regional seismic-stratigraphy and ODP constraints to a base-/intra-Miocene reflector in southwestern Eurasia Basin and North Svalbard margin (~23–10 Ma in the Nansen Basin to ~10?–6 Ma in the Yermak Plateau; Geissler and Jokat, 2004; Engen et al., 2009; Geissler et al., 2011; Mattingsdal et al., 2014; Faleide et al., 2024a; Meza-Cala et al., 2024; Fig. 2). This reflector is located at a depth where the average interval velocity of the upper sedimentary successions increases from ~2.00–2.35 km/s to ~2.50–2.70 km/s (Fig. 5). The top acoustic basement picked from the reflection profiles has a variable and partly discontinuous morphology, and a high-impedance character in the shelf. The top acoustic basement reflector dips below the prograding glacial sedimentary units infilling the Nansen Basin (Lasabuda et al., 2018; Faleide et al., 2024a), and coincides with an interface where seismic velocities represent an increase of ~2.70 km/s to more than ~4.50 km/s (Fig. 5). The seismic signatures of the top acoustic basement on the shelf are related to an erosional unconformity character linked to glacial/sedimentary and breakup processes (Geissler and Jokat, 2004; Faleide et al., 2024a).

For the deeper structure of the crust, we mapped the seismic reflectivity patterns (e.g., orientations, position, and intensities) to create a structural model representing possible internal rheological or structural boundaries (e.g., Moho) across the continent-ocean transition. The mapped key seismic reflectors, together with gridded datasets and models (Figs. 3 and 4) and compiled velocity information (Fig. 5) were used to construct initial crustal models with main model units along the margin transects (see Table 2 and Supplementary figures). The sedimentary layer (i.e., post-breakup/passive margin successions) is defined from the seabed to the top acoustic basement reflectors. The sedimentary layer unit is further sub-divided into an upper (Miocene-present) and lower (Eocene-Oligocene and probably pre-Eocene in the shelf)

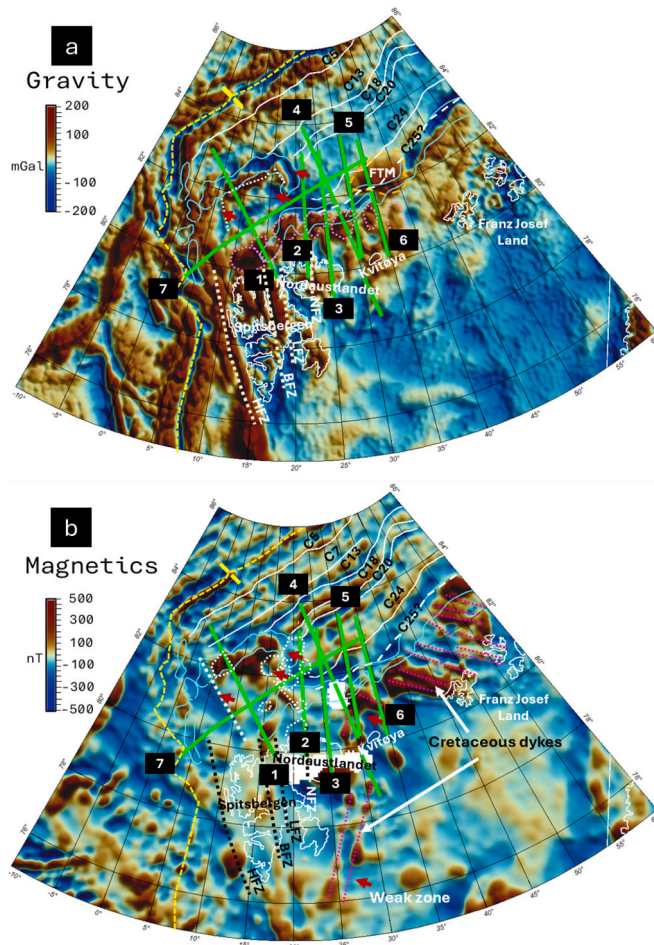


Fig. 3. Input potential field anomalies dataset. (a) Gravity (DTU17, Andersen and Knudsen, 2020). (b) Magnetics (MAG_NOW, Glebovsky et al., 2006). Locations of crustal transects are indicated by thick light green lines. Line drawings on top of maps highlight different tectono-magmatic trends (see text for further explanations). FTM, Franz Josef Trough Mouth Fan. The high gradient magnitude of the Bouguer gravity anomaly is indicated in light-blue polygon. Red arrows indicate amplitude shifts in potential field maps (e.g., see Geissler and Jokat, 2004). Pink dotted lines are axes of magnetic anomalies interpreted as Early Cretaceous dolerite dykes (e.g. see Minakov et al., 2018; Abdelmalak et al., 2024a). Other abbreviations and map features as in Fig. 1. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

units by the intra-sedimentary reflector (Table 1; Faleide et al., 2024a). The mapped internal reflectivity allowed us to determine a continental crust model unit sub-divided into upper, middle, and lower units. In the case of the upper continental crust unit, pre-breakup strata are also partly inferred based on the sub-horizontal and semi-continuous character of the seismic reflectivity (Table 3). The transition from middle to lower continental crust is in some cases marked by a change in seismic reflectivity character expressed by a group of dipping reflections towards a region of more sub-horizontal and subdued amplitudes. The lower continental crust towards its deeper parts presents high-amplitude and chaotic reflectivity, yet is coherent as it seems to be located near a boundary where faults may detach at depth. A similar procedure for the delimitation of the lower crustal continental blocks was implemented towards the continental slope and beyond, north of the seismic profiles. These areas were later divided into transitional (upper and lower) and oceanic (upper and lower) crust main model units (see below for further constraints on delimitation of these model blocks; Table 1).

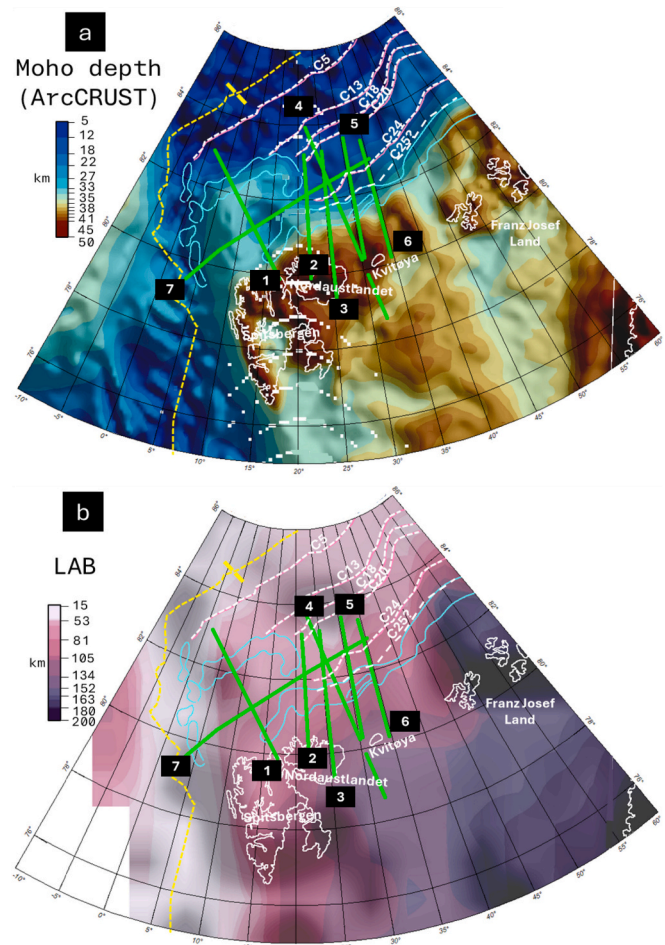


Fig. 4. Input models of crust and lithosphere structure. (a) Depth to Moho (ArcCRUST, Lebedeva-Ivanova et al., 2019). (b) Depth to Lithosphere Asthenosphere Boundary (LAB, Klitzke et al., 2015). Locations of crustal transects are indicated by thick light green lines. Other abbreviations and map features as in Figs. 1 and 3. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

3.2.2. Lithospheric mantle

To account for the regional variation in gravity field, a lithospheric thickness has been assumed based on an available model of the lithosphere-asthenosphere boundary (LAB; Klitzke et al., 2015) (Fig. 4). This workflow resulted in an upper and lower mantle lithosphere and asthenosphere layers, respectively (Table 3). The density heterogeneity due to a thermally differentiated lithospheric mantle is standard for the gravity modelling procedure at rifted continental margins (e.g., Breivik et al., 1999; Breivik et al., 2003; Chappell and Kusznir, 2008; Minakov et al., 2012a; Lebedeva-Ivanova et al., 2019).

3.2.3. Potential field modelling

The 2D gravity and magnetic modelling procedures have been carried out using the GM-SYS/Oasis Montaj software (Seequent, 2024), and its GYMSYS-2D module based on the method by (Talwani et al., 1959). The initial forward modelling of gravity consists of a four-layered density model with water (1040 kg/m³), crust (2800 kg/m³), and upper/lower mantle (3300/3200 kg/m³) blocks anchored in oceanic crust region (i.e. isostatically balanced sections in presumed oceanic crust region, e.g. Tsikalas et al., 2005; Minakov et al., 2012a) and using the IBCAO seabed and ArcCRUST Moho and LAB models (inset in lower parts of crustal transect models; Figs. 1, 4). Potential field data of different sources are included as observed signatures (MAG_NOW, DTU, NOAA NRL, CAMP). The initial models allowed for inspection of the

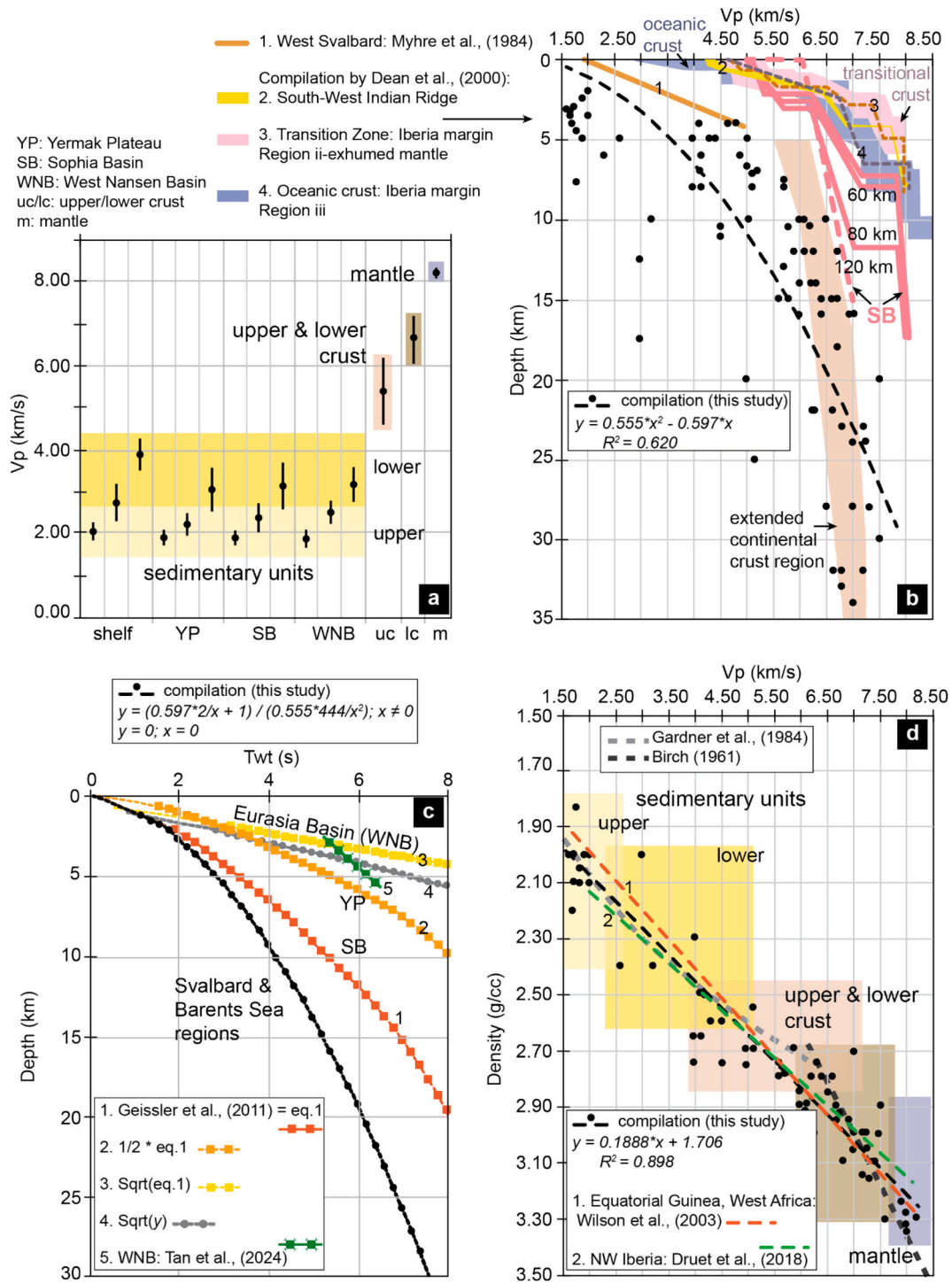


Fig. 5. Velocity, depth, and density relationships used for modelling of crustal transects. The physical parameters and their uncertainty ranges are part of the compilation of available 2D models of the crustal structure based on published seismic and potential field studies (see Fig. 2, Table 2, and Supplementary Material table). (a) Velocity information of the post-rift sedimentary structure, crust, and mantle for the different margin and basin areas within North Svalbard and Eurasia Basin. (b) Velocity-depth functions relevant for the study area. The derived second-order polynomial function used to fit the compiled information on crustal structure is plotted. Velocity fields typical of extended continental, oceanic, and transitional crust (Karner et al., 2021), as well as different crustal regions constrained by seismic experiments from the Iberia margin and South-West Indian Ridge (Dean et al., 2000) are indicated in the plot. Seismic refraction results of profile distances at 60 km, 80 km, and 120 km across the Sophia Basin (SB) are also indicated (Quer, 2015). (c) Time-depth functions for conversion of mapped interfaces (e.g., intra-sedimentary and top acoustic basement reflectors) on seismic reflection data (in seconds, two-way traveltime) to depth (in kilometers). The velocity-depth function in (b) is solved for time-to-depth conversion. (d) Velocity-density functions for the study area, as well as classic functions (Gardner's sedimentary rocks and Birch's diabase, gabbro, and eclogite profiles, retrieved from Marengo et al., 2013) and those from other magma-poor margins are included for comparison (West Africa, Wilson et al., 2003; NW Iberia, Druet et al., 2018).

Table 1
Data used in the construction of the crustal transects (CT1–7).

Crustal transect	Seismic reflection profiles	Seismic refraction data	References and comments
CT1	NPD-MF-4-93; AWI-20130200; AWI-20040005; AWI-1991–21/24/25/26	Profiles AWI-20130-199/200; Sea-ice drift stations of FRAM III and IV experiments	Seismic reflection data is from ARCTIC’91 expedition, Jokat et al. (1995) ; AWI-2004, Geissler and Jokat (2004) , Jokat et al. (2008) ; AWI-2013 of MSM31 expedition, Quer (2015) ; and NPD-MF-4-93(in-house data). Seismic refraction data in the Sophia Basin in Quer (2015) and in the northern parts of the Yermak Plateau and Nansen Basin in Jackson et al. (1984) ; Duckworth and Baggeroer (1985) Seismic reflection data in the shelf northwest of Nordaustlandet (NPD; in-house) and a recently acquired profile (L3-GoNorth22) in the southwestern Nansen Basin during the Go-North-2022 marine scientific expedition (Simon and Paasche, 2024) The model is supported by existing sonobuoy and seismic reflection data only in the shelf northeast of Nordaustlandet (100–200 km offset length of the model). Dataset with an earlier crustal model across the same section using the gravity anomaly in Geissler and Jokat (2004) Seismic reflection in Meza-Cala et al. (2024) . Nansen Basin’s architecture also derived from the adjacent (~30 km to the east) seismic reflection profile of AWI-20010100 in Jokat and Micksch (2004) . The oceanic crust domain at the location of C20 to C18 is just ~30 km south of the FRAM-IV experiment (~4–5 km-thick oceanic crust; Kristoffersen and Husebye, 1985). Ymer expedition results in Eldholm et al. (1984) Seismic profiles in Lutz et al. (2018) , who also presented crustal models based on gravity data (see also Weigelt, 1998) Seismic profile as in Kristoffersen et al. (2021) , and Faleide et al. (2024a)
CT2	NPD-HAP-04-84; L3	None	
CT3	AWI-990170	Sonobuoy stations SB99–32/33/34/35	
CT4	NPD-1204-402	Sonobuoy stations of “Ymer” expedition south of Kvitøya	
CT5	BGR-13-203/204	None	
CT6	NPD-1204-422	None	

Table 1 (continued)

Crustal transect	Seismic reflection profiles	Seismic refraction data	References and comments
CT7	AWI-20130-112/130/165	Sonobuoy data from ARKTIS-XV/2 (ARK99) expedition	Seismic profiles in Quer (2015) , MSM31 expedition. Stratigraphic information from Ocean Drilling Program (ODP) drilled at the southern Yermak Plateau (late Miocene, ~6 Ma; Mattingdsal et al., 2014). ARK99 expedition results in Geissler and Jokat (2004)

behavior of the gravity field along the profile due to the combined influence of starting input parameters and helped to identify areas for consideration during subsequent adjustments of local discrepancies. The changes in densities and magnetic susceptibilities obtained from the modelling of potential fields, and a list of the crustal model parameters is provided in [Table 3](#). The goal of the modelling was to create coherent geological models that address local variations in the simplified crustal transects by making continuous adjustments to the crustal geometries based on mapped interfaces from the seismic interpretation procedure, as well as the densities and magnetic susceptibilities of the input bodies (e.g., [Tsikalas et al., 2005](#); [Antobreh et al., 2009](#); [Blaich et al., 2011](#); [Marello et al., 2013](#); [Zastrozhnov et al., 2018](#)). The densities used for the gravity modelling and related uncertainties were constrained from the crustal structure compilation and were defined by empirical velocity-density-depth functions ([Fig. 5](#) and [Table 3](#)) (e.g. [Wilson et al., 2003](#); [Peron-Pinvidic et al., 2016](#); [Ramos et al., 2017](#); [Druet et al., 2018](#); [Abdelmalak et al., 2017](#); [Karner et al., 2021](#); [Pérez-Gussinyé et al., 2023](#)). The acceptable error limit was kept under a narrow margin of ~7–16 % between observed and calculated gravity anomalies, representing less than ~20 mGal difference. For the case of the magnetic modelling a wider error of up to ~30–60 % between observed and calculated responses has been obtained. The magnetic anomaly of Svalbard and the Barents Sea regions have a complex pattern reflecting the amalgamation of different tectonic blocks and modified fabrics overprinted by dyke/sill intrusions and making magnetic modelling a difficult procedure ([Marello et al., 2010](#); [Marello et al., 2013](#); [Minakov et al., 2012b](#)). Nonetheless, some first-order observations can be deduced from the magnetic modelling exercise in areas where the wavelengths fitted a marked positive spike (> 200 nT) relative to the surroundings, and in the context of the orientations, position, and extent in map view ([Fig. 3b](#)).

4. Results

4.1. Observations and trends in potential field data

Potential field anomalies in the North Svalbard margin are additional indicators of structural trends that may correspond to heterogeneities within the basement and crustal plateaus (e.g., [Marello et al., 2010](#); [Marello et al., 2013](#); [Klitzke et al., 2016](#)) ([Fig. 3](#)). Semi-circular (five to six) local positive/maxima gravity anomalies are located on the outer shelf north of Kvitøya between 10°E and 40°E ([Geissler and Jokat, 2004](#)) ([Fig. 3a](#)). These anomalies are interrupted towards the east at the location of stacked NW-SE-trending positive total intensity magnetic anomalies representing Cretaceous dykes emplaced during the HALIP events ([Minakov et al., 2012b](#); [Minakov et al., 2018](#)), and correlates to a weak zone in the region of southern Kvitøya ([Marello et al., 2013](#)) ([Fig. 3b](#)). A similar spatial character on the magnetic data is also observed on the younger magnetic isochrons C13 to C7 of the Gakkel Ridge truncated onto the northern Yermak Plateau, and later into the Fram Strait around a time of 21 Ma and then into its southern

Table 2

References of published crustal models and information of North Svalbard and vicinity within Eurasia Basin and Barents Sea regions. See also Supplementary table file.

Reference number	Reference	Type of study	Comments
1	Ritzmann and Jokat, 2003	Seismic refraction and gravity models	N-S refraction survey in NW Svalbard
2	Ritzmann et al., 2004	Seismic refraction and gravity models	E-W refraction survey in NW Svalbard
3	Czuba et al., 2005	Seismic refraction and gravity models	E-W refraction survey in NW Svalbard
4	Quer, 2015	Seismic reflection and refraction	AWI-2013 seismic reflection and refraction surveys in Sophia Basin
5	Jackson et al., 1982	Seismic reflection and refraction, gravity models	FRAM-I seismic experiment in West Nansen Basin
6 and 7	Minakov et al., 2012b, 2018	Seismic reflection and refraction, gravity models, dyke emplacement models	ESVA and 4AR seismic surveys in east Spitsbergen and North Barents
8	Ritzmann et al., 2002	Seismic reflection and refraction, gravity models	E-W refraction survey in W Svalbard to Knipovich Ridge
9	Sellevoll et al., 1991	Seismic refraction	N-S and E-W refraction surveys on Svalbard
10	Ljones et al., 2004	Seismic refraction	E-W refraction survey in W Svalbard to Knipovich Ridge
11	Breivik et al., 2005	Seismic reflection and refraction, gravity models	Seismic experiments in south Svalbard
12	Myhre, 1984	Seismic reflection/refraction	Sonobuoys in southern Svalbard
13	Eldholm et al., 1984	Seismic reflection/refraction	YMER-expedition; Sonobuoys in south Greenland
14	Crane et al., 1982	Seismic reflection/refraction, heat flow modelling	Sonobuoys and heat flow measurements in Yermak Plateau
15	Nikishin et al., 2018	Seismic reflection and sonobuoys	Profiles in Eurasia Basin
16	Jokat and Micksch, 2004	Seismic reflection and sonobuoys	Profiles in North Barents and Eurasia Basin
17	Jokat et al., 1995	Seismic reflection and sonobuoys	ARCTIC'91-expedition; West Eurasia Basin, Yermak Plateau, Morris Jesup Rise, Gakkel Ridge
18	Lutz et al., 2018	Seismic reflection, gravity models	Profiles in North Barents and Eurasia Basin
19	Meza-Cala et al., 2024; Kristoffersen et al., 2021	Seismic reflection	UNCLOS-Norway; Profiles in North Barents and Eurasia Basin
20	Engen et al., 2009	Seismic reflection/refraction	Seismic profiles and sonobuoys in West Nansen Basin
21	Sundvor and Austegard, 1990	Seismic reflection/refraction, gravity models	Seismic profiles in west and northwest Spitsbergen
22	Shulgin et al., 2020	Seismic reflection/refraction, gravity models	Seismic surveys in SW Barents
23	Breivik et al., 2002	Seismic reflection/refraction, gravity models	Seismic surveys in eastern Bjørnøya
24	Sokolov et al., 2023	Seismic reflection and gravity models	West Nansen Basin
25	Geissler and Jokat, 2004;	Seismic reflection/refraction, gravity models	Seismic profiles and sonobuoys in northern Svalbard

Table 2 (continued)

Reference number	Reference	Type of study	Comments
26	Geissler et al., 2011; Weigelt, 1998; Weigelt and Jokat, 2001	Seismic reflection and gravity models	North Svalbard and West Eurasia Basin
27	Minakov et al., 2012a	Gravity models	North Svalbard, northern Barents-Kara, Eurasia Basin
28	Baturin et al., 1994	Seismic reflection and gravity models	NW Spitsbergen, north Barents, southwest Nansen Basin
29	Faleide et al., 2018	Gravity models	Barents-Kara regions
30	Marello et al., 2010, 2013	Potential field models	Barents Sea
31	Glebovsky et al., 2013	Gravity models	Eurasia Basin and Arctic Ocean
32	Eiken, 1994	Seismic reflection	Svalbard
33	Faleide et al., 1991	Seismic reflection and gravity models	West Barents Sea and Svalbard
34	Boebel, 2000	Seismic reflection and gravity models	Svalbard, Fram Strait, northeast Greenland
35	Kristoffersen and Husebye, 1985	Seismic reflection	FRAM-IV seismic experiment in northern Yermak Plateau and West Nansen Basin

Table 3

Summary of density and magnetic susceptibility values used for modelling in crustal transects CT1–7.

Main model units	Density ($\times 10^3$ kg/m ³)	Magnetic susceptibility (SI units)
Water	1.04	–
Upper sedimentary (post-breakup, Miocene-Quaternary)	2.00–2.10	–
Lower sediments (post-breakup, Eocene-Oligocene)	2.30–2.45	–
Lower sediments (shelf and/or syn-rift, pre-Eocene)	2.45–2.55	–
Evaporite (salt/halite)	2.20	–
Upper continental crust (and pre-rift in shelf)	2.54–2.70	–
Middle continental crust	2.74–2.88	0.002–0.0004
Lower continental crust	2.80–2.96	0.001–0.0008
Northeast Yermak Plateau	2.64–2.80	0.002–0.0003
Southwest Yermak Plateau	2.60–2.76	0.002–0.0001
Upper transitional basement (COT North Svalbard-Nansen Basin)	2.74–2.90	0.001–0.004
Lower transitional basement (COT North Svalbard-Nansen Basin)	2.97–3.18	0.01–0.008
Sophia Basin (upper-lower blocks)	2.70/2.76–3.00	0.0007
Molloy Deep	3.15	0.001
Oceanic crust (upper-lower blocks)	2.75/2.80–2.95	0.006
Mantle (upper-lower)	3.30–3.20	0.0006

termination 11 Myr after (Fig. 3b; Jokat et al., 2016).

Potential field anomaly datasets (e.g., Feden et al., 1979; Kristoffersen and Husebye, 1985; Brozena et al., 2003; Glebovsky et al., 2006; Jokat et al., 2008; Jokat et al., 2016; Gaina et al., 2011) indicate two broadly different regions within the Yermak Plateau, one to the southwest with low amplitude anomalies, and another to the northeast that shows large positive anomaly values (> 200 mGal and 300 nT) (Fig. 3). In the northeast part of the plateau and parts of the Sophia Basin, several semi-elongated trends (e.g., line-ridge features) are visible and oriented in a northeast-southwest direction, with one of them covering about ~ 130 km of the plateau's length (Fig. 3). This feature appears to be slightly segmented in half across a narrower ~ 15 km wide

zone of less intense magnetic anomaly responses (Fig. 3b). The areas where segmentation occurs are characterized by shifts in the potential field anomaly maps (red arrows in Fig. 3) and are oblique to structural lineaments of major fault zones mapped onshore Svalbard (Fig. 1). The line-ridge structures observed on the magnetic data are observed to decrease in size and anomaly intensities following a southwards trend and within the Sophia Basin (Fig. 3b).

4.2. Crustal structure along modelled transects

4.2.1. CT1 model

The CT1 model is a 400 km-long and N-S trending crustal transect that crosses the shelf northeast of Spitsbergen, the Sophia Basin, the northern Yermak Plateau, and the northwestern Nansen Basin (Fig. 6 and Table 1). The top acoustic basement reflector, identified beneath a thick sequence of prograding sediments of the upper sedimentary unit, is discontinuous and observed over the first ~30 km of the profile towards the south. Structural interpretation of the internal seismic reflectivity in the shelf north of Spitsbergen might suggest a multi-layered continental crust consisting of a ~50 km-wide continental block with one or more smaller down-faulted zones (Fig. 6). The top acoustic basement and Moho relief define a relatively narrow ~50 km taper geometry towards the Sophia Basin crustal units (Fig. 6). The top acoustic basement is deformed in the Sophia Basin with continent-facing fault surfaces. The top acoustic basement overlies a body at deeper crustal levels (~6–8 s twt) with P-wave seismic velocity greater than 7.0 km/s (Quer, 2015). We modelled this body with a high-density (3000 kg/m³) and thin (~5–7 km-thick) crust (Fig. 6) in line with previous studies regarding the structure of the Sophia Basin (Boebel, 2000; Geissler and Jokat, 2004).

There is little indication of internal reflectivity on the northern part of this transect beneath ~3–4 s twt except in some small parts underneath the Mosby Seamount, as seafloor multiples and the occurrence of opaque reflectors interpreted as lava flows or sills (e.g., Geissler and Jokat, 2004) primarily obscure the seismic image (Fig. 6). The northern part of the southwest Yermak Plateau within the Sverdrup Bank (Fig. 1) area has an upper sedimentary unit composed by contourite deposits (< 20 Ma units) (Fig. 6). The upper continental crust unit in the north of the plateau consists of an area with highly deformed or folded units beneath the top acoustic basement reflector (Fig. 6). In this area, the high-amplitude (> 500 nT) magnetic anomalies gradually decay from the northern tip of the plateau to nearly zero-to-negative in the northwestern Nansen Basin (i.e. at Chron C7; Fig. 3b). To fit gravity anomalies, we adjusted the initial model to a shallower Moho boundary between the plateau and basin region (Fig. 6).

4.2.2. CT2 model

The CT2 model is a ~350 km-long N-S-trending crustal transect that crosses the northwestern part of Nordaustlandet and continues along the western border of the southwestern Nansen Basin (Fig. 7 and Table 1). The shelf is characterized by two local depocenters within the lower sedimentary unit, no deeper than ~3 s twt (> ~5 km-depth) and separated in the middle by a basement high, thus configuring a horst-and-graben system (Fig. 7). These grabens are filled with probably syn-rift sediments with densities of 2500 kg/m³. This area of the Svalbard margin is characterized by segmented/local gravity highs in the shelf (e.g., Geissler and Jokat, 2004) (pink dotted polygons in Fig. 3a). At the continental slope, a high-density lower crust (2950–3000 kg/m³) is required to fit the gravity signature. The recently acquired GoNorth22 seismic reflection profile L3 shows in that region an upper crust structure with deformation accommodated by continent-dipping faults that may detach beneath 6–8 s twt (~10 km depth) (Fig. 7).

4.2.3. CT3 model

The CT3 model is a ~400 km-long, N-S-trending crustal transect that crosses the eastern part of Nordaustlandet and ends towards the

southwestern Nansen Basin (Fig. 8 and Table 1). The onshore continental crust units depicted by the model are not well constrained due to sparse data coverage. Based on the onshore geology, the Precambrian crystalline basement of Nordaustlandet is overlaid by upper Paleozoic and Mesozoic sediments (e.g., Grogan et al., 1999). The continental block units exhibit a structural depression with intense normal faulting, and a prominent horst structure is imaged as a marginal high at the seaward limit of the shelf break (Fig. 8). Internal seismic reflectivity within the upper crust unit can be traced down to ~5–6 s twt (~7 km depth) towards the middle crust unit. The gravity anomaly values shift from 40 mGal to –20 mGal over a width of less than ~50 km and modelled with high-density blocks (>2980 kg/m³) (Fig. 8).

4.2.4. CT4 model

The N-S trending CT4 model consists of two sections totaling ~500 km in length across the shelf of Kvitøya and into the Nansen Basin (Fig. 9 and Table 1). The first section is 100 km-long south of Kvitøya and is separated by a spatial gap of 110 km from the second section which is ~400 km-long and crosses the shelf northwest of Kvitøya towards the southwestern Nansen Basin (Fig. 9). The shelf south of Kvitøya consists of reflections truncated towards the north against the seafloor and that are likely of pre-Cenozoic origin due to the range of average seismic velocities (~2.83 to 4.67 km/s) and densities (2500–2650 kg/m³) (Fig. 5, e.g., Eldholm et al., 1984; Minakov et al., 2012a; Faleide et al., 2018). A salt diapir structure with relatively low-density (2200 kg/m³) is observed in the central part of CT4's first section within interpreted pre-breakup sedimentary units (Fig. 9).

On the shelf northwest of Kvitøya, several parallel and gently dipping reflections are observed underneath the top acoustic basement within the upper continental crust unit at ~4 s twt (0 km profile distance, second section; Fig. 9). An internal crustal reflectivity with variable dips within the same area indicates a heterogeneous middle and lower crust at ~6–8 s twt (~8–10 km depth; Fig. 9). Igneous intrusions, such as NE-SW-trending Cretaceous HALIP dikes (e.g., Minakov et al., 2012b; Minakov et al., 2018) could cause the high-amplitude seismic reflections and corresponding intense magnetic anomalies reaching 400 nT (Figs. 3 and 9). The horst-and-graben structural features on the shelf spread over ~50 km and are accompanied with a gradual increase of the gravity anomalies with magnitudes reaching up to ~60 mGal. In this location, the top acoustic basement is downfaulted with throws of ~1.5 s twt (< 3 km) (Fig. 9).

The shelf break-proximal region presents a set of reflections that are found from ~6 to ~4 s twt in the upper and middle continental crust units and delineate a well-defined rotated fault-block (~150 km profile length) (Fig. 9). The orientation of this structural feature indicates rotation guided by low-angle fault planes dipping towards the ocean (Fig. 9). Moving further seaward, there is another sudden down-step of the basement topography from ~6 to ~8 s twt, and internal reflectivity agglomerates over a distance of less than ~15 km in width within the units of the continental slope. This structure can be interpreted as a region with double-sided fault plane orientations that may detach at 8 s twt (~10–15 km depth). High-density blocks are required to fit in this area the observed gravity anomaly (2800–3150 kg/m³).

4.2.5. CT5 model

The CT5 model is a ~350 km-long and N-S-trending crustal transect that crosses the shelf north of Kvitøya and terminates into the southwestern Nansen Basin (Fig. 10 and Table 1). The shelf region is characterized by top-to-the-north dipping reflections that are visible at no deeper than ~2 s twt within the upper crust unit and underneath the top acoustic basement (Fig. 10). Moving north along the shelf, an abrupt down-stepping of the top acoustic basement can be observed and is associated with a V-shaped valley. This topographic depression imposed by the top acoustic basement marker appears to be infilled by syn-rift reflection packages with wedge-shaped geometries within the lower sedimentary unit that grows towards the south (Fig. 10). Moving north

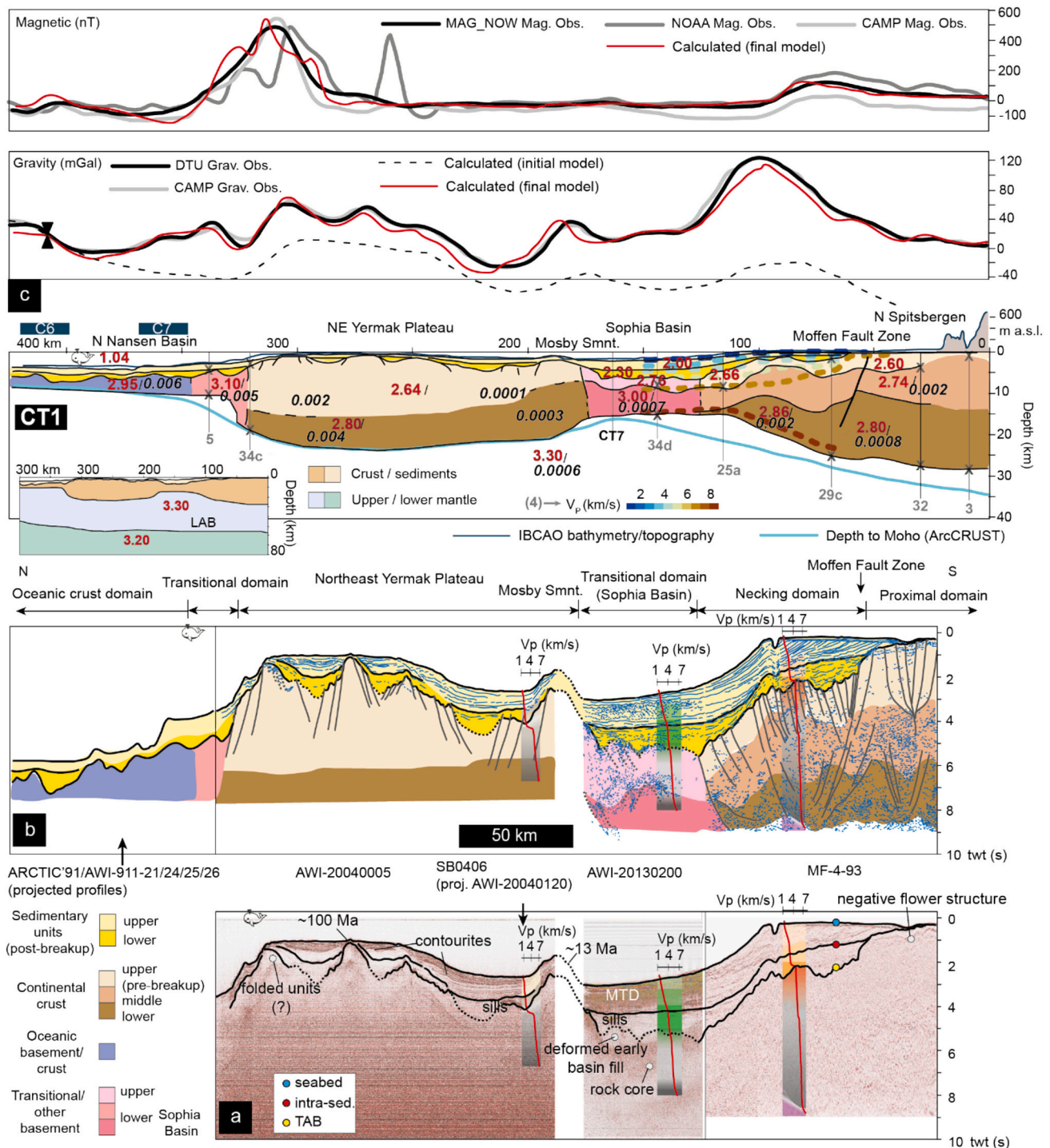


Fig. 6. Crustal transect 1 (CT1). (a) Composite seismic reflection profile from the shelf northeast of Spitsbergen and across the Sophia Basin, the northern Yermak Plateau, and until the northwestern Nansen Basin. The main mapped seismic horizons, seabed, intra-sedimentary (post-breakup), and top acoustic basement (TAB) are indicated. Velocity profiles are based on refraction model results (sonobuoy and OBS; Fig. 5). Age dating constraints are shown for dredged samples at Mosby Smnt. (~13 Ma; Geissler et al., 2019) and Northeast Yermak Plateau (~100 Ma; Estrada et al., 2024). Deformed units and negative flower structure below TAB reflector, as well as sill intrusions, MTD complex and contourites within post-breakup sedimentary structure are indicated. MTD, mass-transport deposits. (b) Geoseismic section interpretation with main units and structures. Rifted margin and crustal domains are indicated. (c) Crustal model with fitted gravity and magnetic anomalies. For the gravity model an initial and a final model are presented. References to used seismic reflection profiles, as well as the crossings with published 2D crustal models (vertical lines with numbers in gray), together with density (in red numbers, 10^3 kg/m^3) and magnetic susceptibility values (in gray numbers, SI units) are indicated and summarized in Tables 1 to 3, respectively. The vertical scale of the onshore topography is exaggerated compared to the subsurface depth scale. Note that the upper continental crust unit may comprise pre-breakup sedimentary strata (see text and Table 3 for further details). Interpreted seafloor spreading anomalies (as Fig. 1) are projected on top. Seismic P-wave velocity gradients based on refraction model results (Quer, 2015; Table 2) is displayed as colour-coded thick dashed lines. See Methods for details. See Supplementary material Figs. 1 and 2 for close-ups to seismic. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

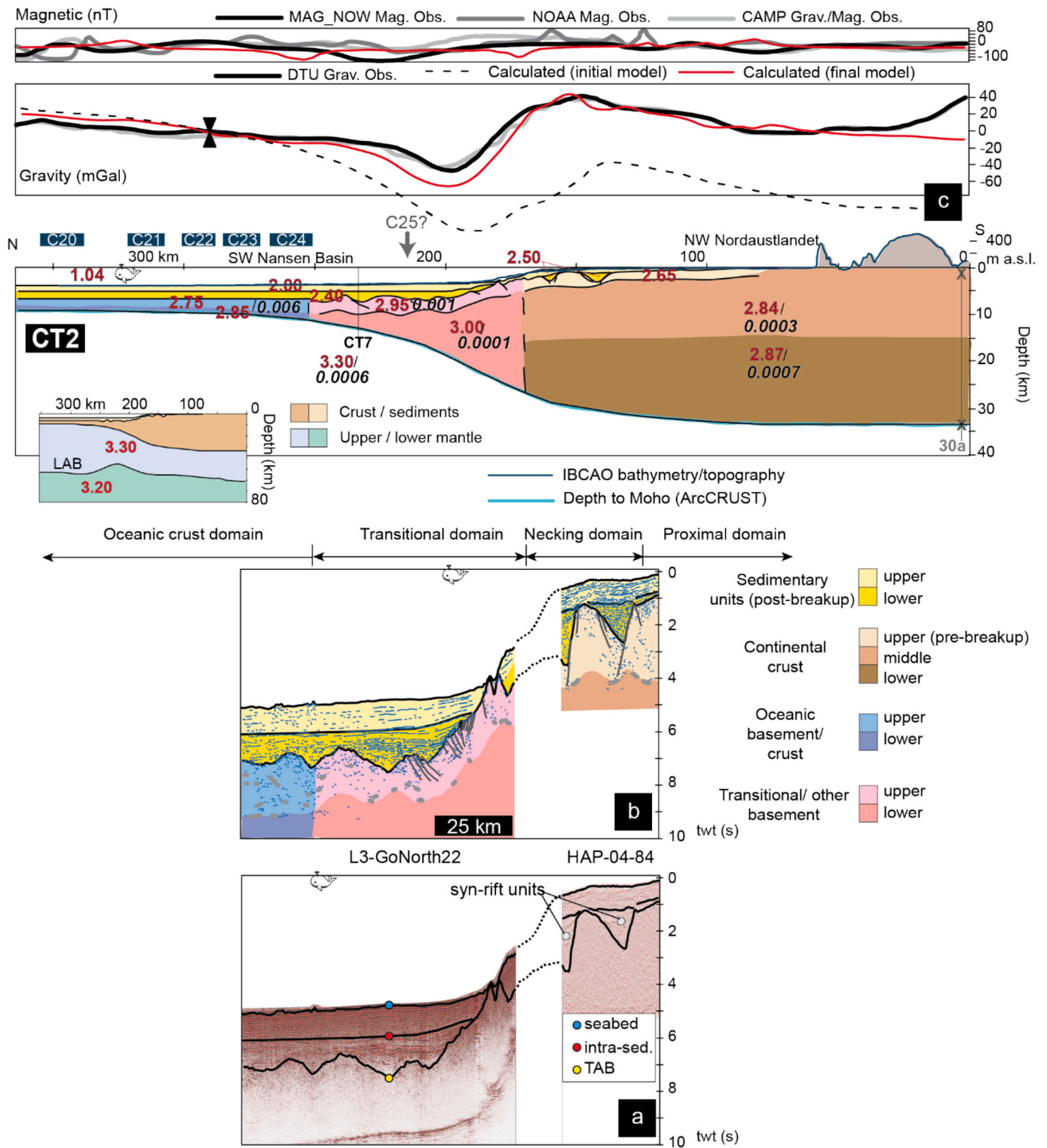


Fig. 7. Crustal transect 2 (CT2). (a) Composite seismic reflection profile crossing the shelf northwest of Nordaustlandet into the southwest Nansen Basin. Note the seismic profile in the Nansen Basin (L3-GoNorth22) is part of the results obtained during the GoNorth 2022 marine expedition in the Arctic Ocean. (b) Geoseismic section interpretation with main units and structures. Rifted margin and crustal domains are indicated. Note the coherent reflective area in the shelf indicating large half-graben systems. (c) Crustal model with fitted gravity and magnetic anomalies. Other explanations as in Fig. 6. See Supplementary material Fig. 3 for close-up to seismic.

along the transect, these syn-rift units are observed resting atop the southern flank of a marginal high (Fig. 10). The marginal high constitutes a ~ 50 km-wide concave ridge that occupies the shelf break. This structural feature is downfaulted (~1.2 s twt) towards its northern flank, where a different ~20 km-wide rotated fault-block is juxtaposed and

separates the shelf break from the deep oceanic Nansen Basin part through a seaward dipping fault (throw greater than ~2 s twt) (Fig. 10). The intracrustal reflectivity is chaotic between the area of the concave structural high and the latter rotated fault-block but configure a more coherent zone at depths deeper than 8 s twt (~10 km). These areas

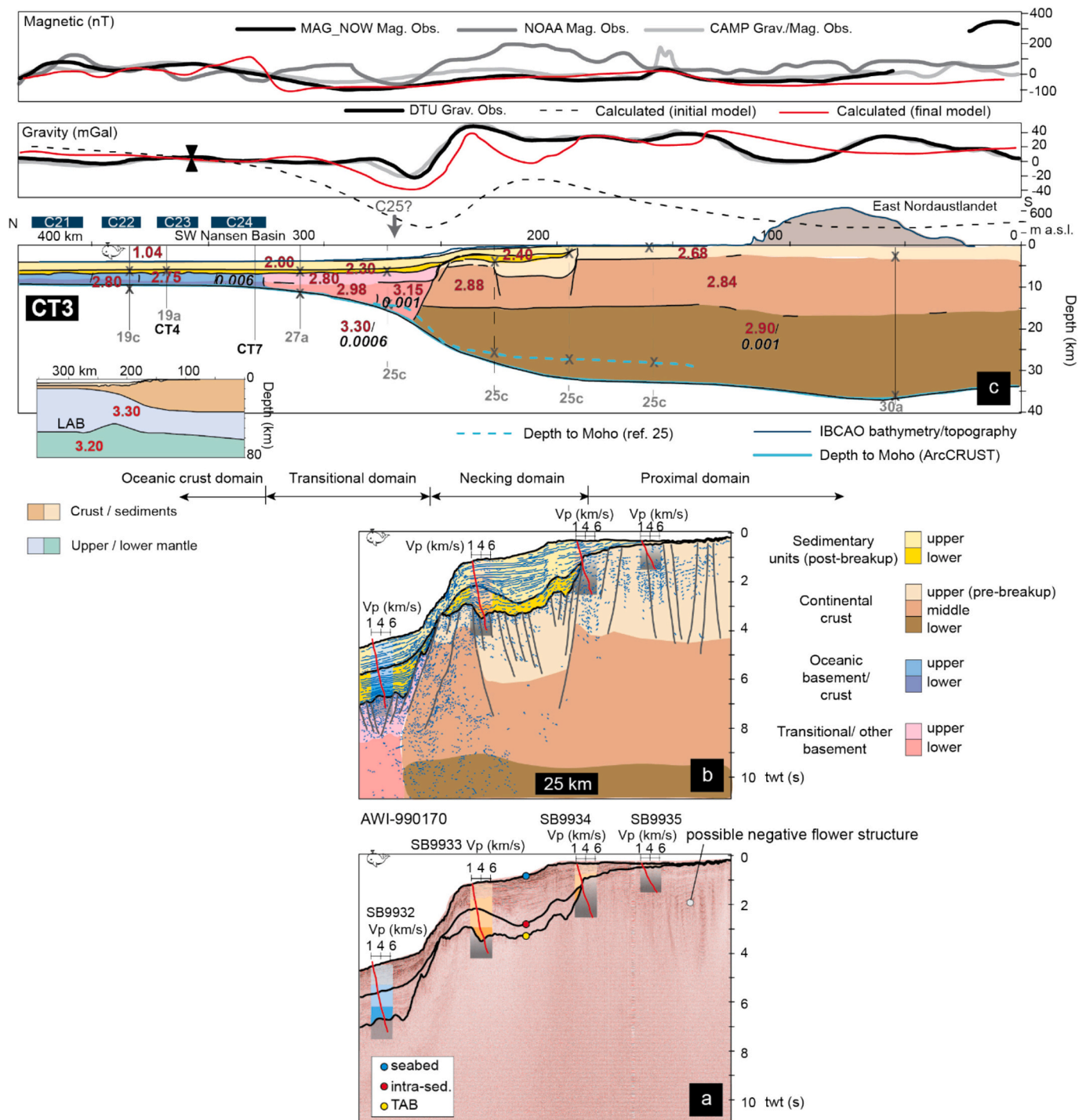


Fig. 8. Crustal transect 3 (CT3). (a) Seismic reflection profile crossing the shelf northeast of Nordaustlandet and parts of the southwest Nansen Basin. Seismic profile and velocity are those as in (Geissler and Jokat, 2004). (b) Geoseismic section interpretation with main units and structures. Rifted margin and crustal domains are indicated. Note the coherent reflective area in the southern parts of the shelf indicating planar fault systems with likely strike-slip components. (c) Crustal model with fitted gravity and magnetic anomalies. The model is extrapolated to the south and north across Nordaustlandet and the Nansen Basin, respectively. Other explanations as in Fig. 6.

within the middle and lower crust units may correspond to shear zones that are accompanied with shallower Moho depths seawards (~10 km deep) (Figs. 4a and 10).

In the deeper parts of the Nansen Basin which are imaged by the reflection data north of the above-mentioned border fault, previous studies have suggested deep (< ~11 km depth) continent-dipping detachment surfaces (Lutz et al., 2018). In this study we observe a similar structuring regarding the presence and orientation of probable

detachment surfaces. However, we test a different interpretation by suggesting wedge-shape units within the large rotated crustal blocks in the upper transitional crust units (Fig. 10). These wedge-like units are found draped at the top by a top acoustic basement marker with high-impedance and gentle dips (<5–10°) (Fig. 10). A sharp relief of the Moho is required to fit the gravity and a region of high-density (> 3000 kg/m³) blocks are required to match the potential field response as compared to alternative models with deeper and broader reliefs of the

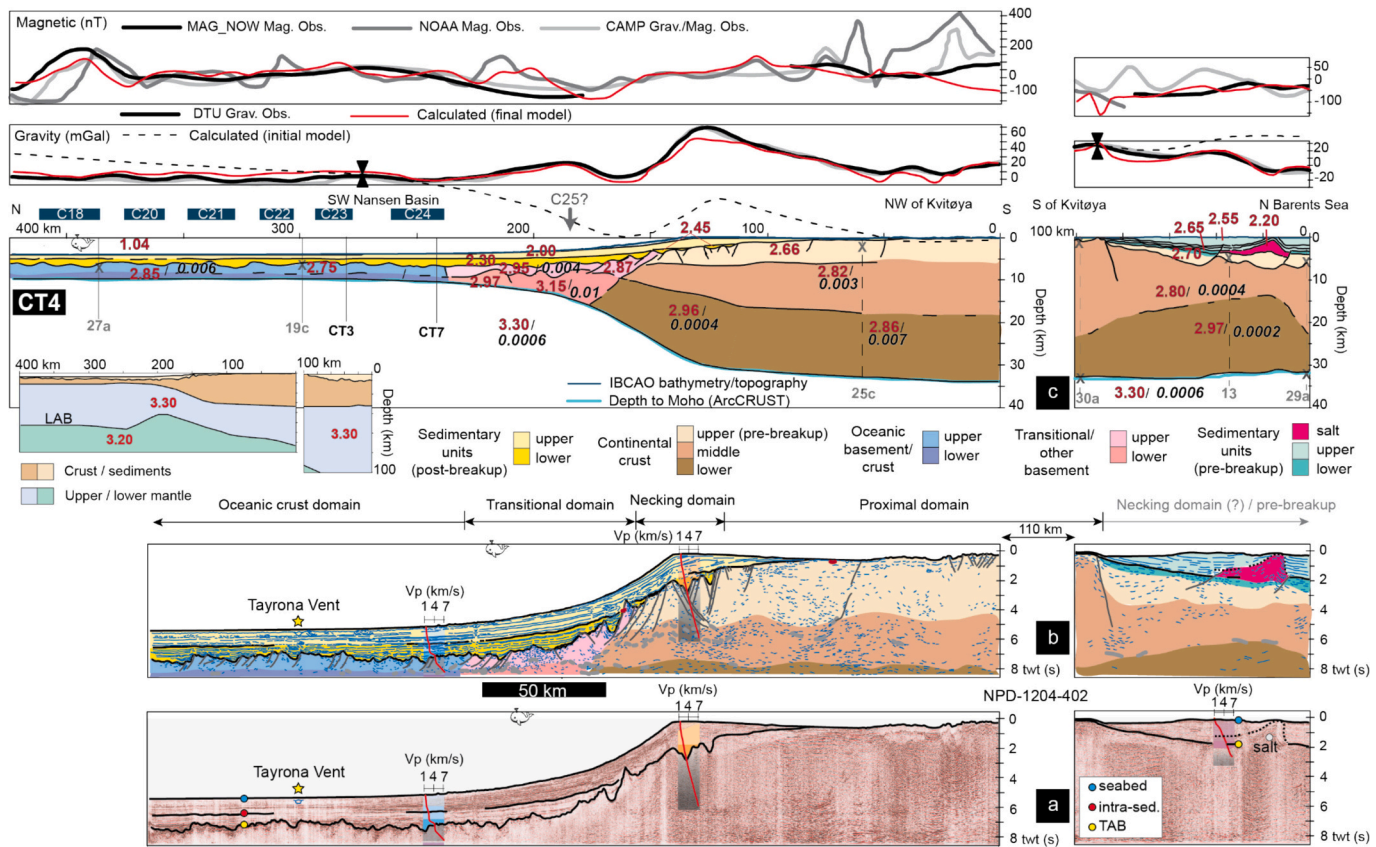


Fig. 9. Crustal transect 4 (CT4). (a) Seismic reflection profile crossing the shelf southwest of Kvitøya, the shelf northwest of Kvitøya, and into the southwest Nansen Basin (profile NPD-1204-402; Meza-Cala et al., 2024). The profile consists of a single 2D multichannel seismic line, and the blank area of ~110 km between the panels has not been modelled in the current study. The velocity profiles are estimated from interval velocities used for migration. Location of a young (~2.6 Ma) hydrothermal feature, Tayrona Vent (Meza-Cala et al., 2024), is indicated in profile. (b) Geoseismic section interpretation with main units and structures. Rifted margin and crustal domains are indicated. Note the presence of pre-breakup Paleozoic and Lower Mesozoic rift structuring with evaporitic deposits (salt diapir) south of Kvitøya. The shelf break and continental slope areas towards the Nansen Basin indicates an extensional mode with lower crust and mantle unroofing structural type. (c) Crustal model with fitted gravity and magnetic anomalies. Other explanations as in Fig. 6. See Supplementary material Fig. 4 for close-ups to seismic.

depth to Moho (note difference between Model A and Model B, as well as the 7 km/s velocity gradient in this area in Fig. 10).

4.2.6. CT6 model

The CT6 model is a ~350 km-long and N-S-trending crustal transect that crosses the shelf east of Kvitøya into the Nansen Basin (Fig. 11 and Table 1). The top acoustic basement marker is a well-defined reflector underlying the prograding margin infill of the upper sedimentary unit across the imaged portions from the shelf towards the shelf break. Semi-horizontal reflections are still visible with relatively small-scale (<~10 km-width) rotated-fault blocks below the top acoustic basement in the upper continental crust unit over a broad area (~150 km) (Fig. 11). As this seismic marker approaches the narrow (~20 km-wide) shelf break region, increased intracrustal reflectivity can be observed at different levels at ~4 and ~6 s twt (Fig. 11). In the same area, seismic interpretations suggest a symmetric antiformal structure with faults mirroring each other and configuring a triangular-shape complex coinciding with a local gravity maximum and a dull magnetic response (Figs. 3 and 11). The shelf break is marked north of the seaward dipping surface of the fault complex, with upper/middle crustal unit structures that include down-faulted geometries dipping towards the south and which, gradually become larger at depth and towards the Nansen Basin. Two large continent-dipping reflections at the foot of the continental slope within the Nansen Basin are interpreted as possible detachment surfaces (Fig. 11), similar to those observed in the CT5 model and previous studies (Lutz et al., 2018). The top acoustic basement along this seismic profile does not have a draping character over deeper reflection

packages but is rather found delineating the rough topography of the oceanic basin.

4.2.7. CT7 model

The CT7 model is a more than 600 km-long W-E-trending crustal transect that crosses the Molloy Deep, the southern Yermak Plateau, the Sophia Basin, and the southwestern Nansen Basin (Fig. 12 and Table 1). The Molloy Deep is a sediment-filled ~20 km-wide U-shape rift valley (Fig. 12). The modelled sedimentary units are represented by ~2 km-thick reflection packages above a patchy top acoustic basement horizon at a tentative depth of ~5 km. The paired gravity anomalies delineate the valley and the terraces on the flanks of the Molloy Deep (Fig. 3a). Immediately to the east of the Molloy Deep valley, an abrupt basement up-stepping reflection is observed at depth of ~4 s twt on the southern Yermak Plateau (Fig. 12). We modelled the Molloy Deep crust unit assuming a basement density of 3150 kg/m³. In contrast, the southern Yermak Plateau located ~30 km and marked by ODP-912 site (Fig. 12) has presumably thinned continental crust (~20 km) with density of 2800 kg/m³. Fault-block rotation is evident along the entire southern part of the Yermak Plateau with mostly planar faults soling out at ~8 s twt (~15 km depth) (Czuba et al., 2005). Sedimentary wedges within the lower sedimentary unit, probably older than Miocene, are observed on the crests of some of these rotated, and probably intruded, fault blocks at the margins of the Yermak Plateau and Sophia Basin (Fig. 12). The Moho boundary along the southern Yermak Plateau is flat and is shallower compared to the plateau's northern area (CT1 transect, Fig. 6). The differences in the Moho depth between these two transects

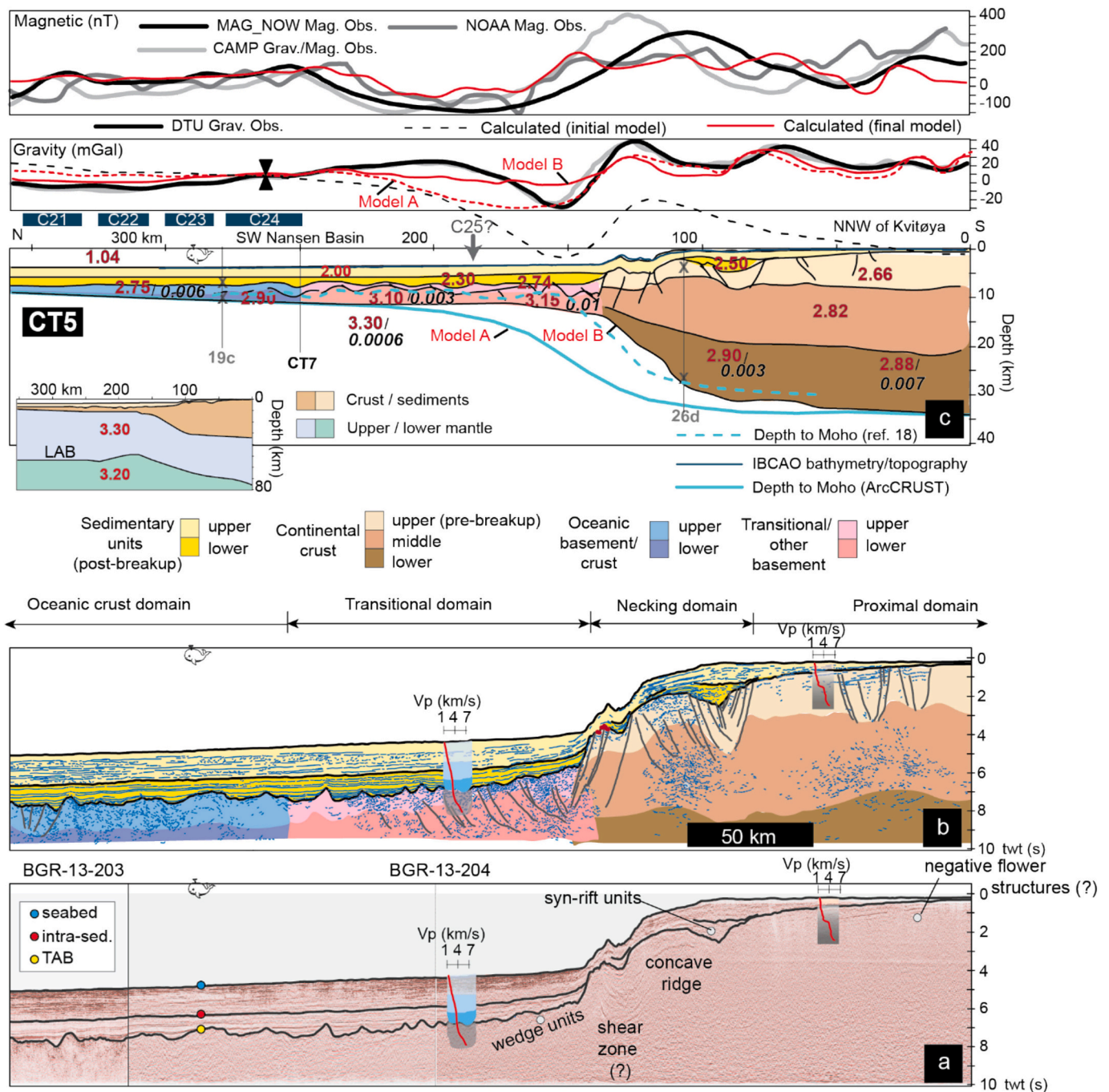


Fig. 10. Crustal transect 5 (CT5). (a) Composite seismic reflection profile crossing the shelf north of Kvitøya into the southwest Nansen Basin (profiles BGR-13-203/204; Lutz et al., 2018). The velocity profiles are estimated from interval velocities used for migration and depth conversion (see Lutz et al., 2018). (b) Geoseismic section interpretation with main units and structures. Rifted margin and crustal domains are indicated. Note the half-graben and concave structural high in the necking domain, and the wedge units within the continental slope at the transitional domain. (c) Crustal model with fitted gravity and magnetic anomalies. Note that two different gravity models that change only the depth to Moho are presented (Models A and B). Other explanations as in Fig. 6. See Supplementary material Fig. 5 for close-up to seismic.

are ~5–7 km. In the southern part of the Yermak Plateau, the crust is thinner than in the northern part by ~5 km and less dense (Figs. 6 and 12, Table 3).

The transition area between the southern Yermak Plateau into the Sophia Basin, as well as further east along the CT7 transect, shows larger acoustic energy reflected at the top acoustic basement (Fig. 12). This seismic marker exhibits opaque signatures, probably masking deeper reflections/units, and corresponds to an interface with P-wave seismic velocities of ~4–5 km/s (Geissler et al., 2011). High-impedance

reflections at a profile distance of ~300 km along CT7 are interpreted as an intruded basement ridge (Fig. 12). To the east of this structure, a sudden basement down-step morphology is observed (from ~4 to ~6 s twt), and it is accompanied with thickness variations of the sedimentary units above. In this area, the basin configuration consists of a westward tilted margin over a high-density crust ($>2850 \text{ kg/m}^3$) (Fig. 12).

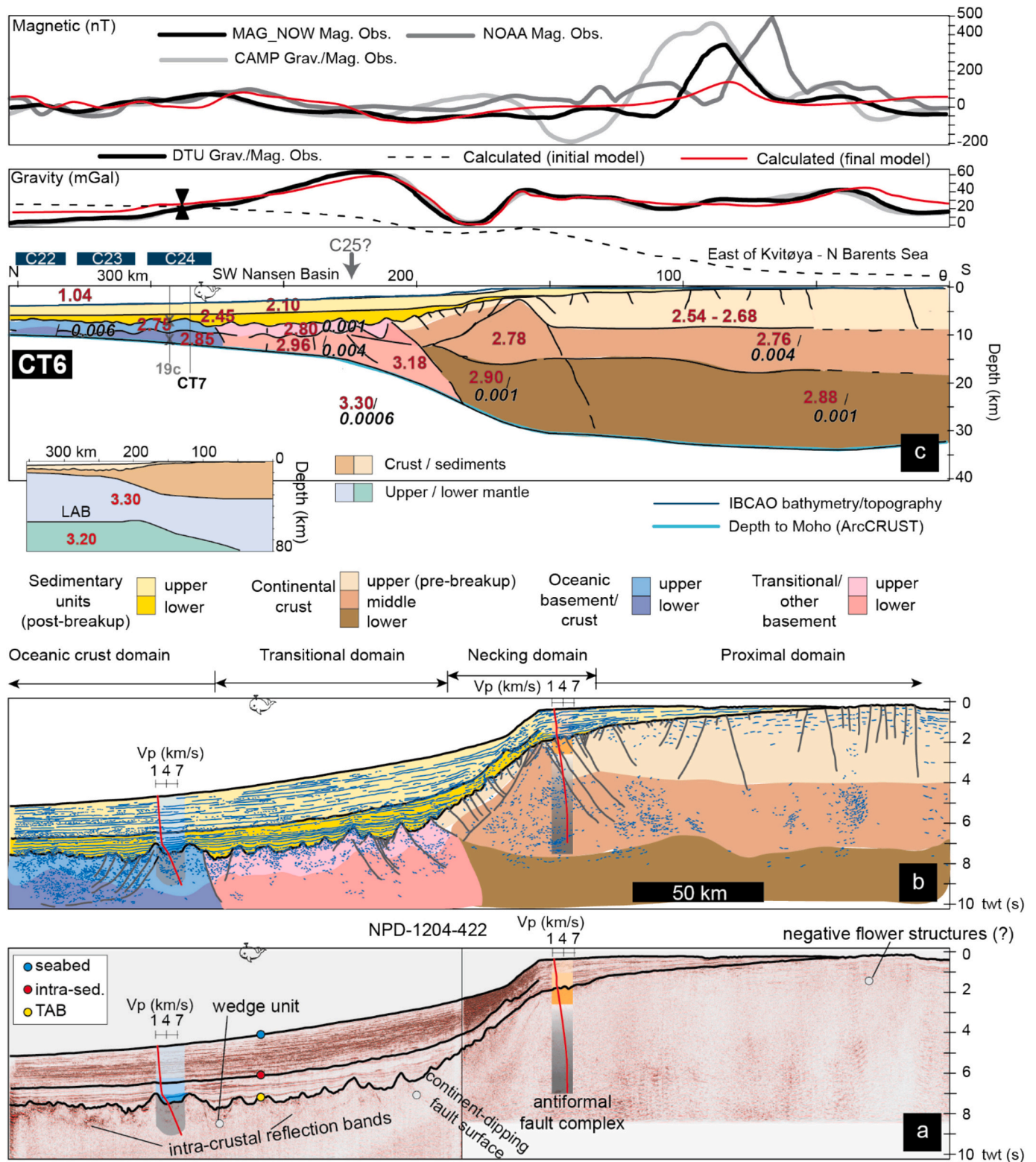


Fig. 11. Crustal transect 6 (CT6). (a) Seismic reflection profile across the shelf northeast of Kvitøya into the southwest Nansen Basin (profile NPD-1204-422; Kristoffersen et al., 2021; Faleide et al., 2024a). The velocity profiles are estimated from interval velocities used for migration. (b) Geoseismic section interpretation with main units and structures. Rifted margin and crustal domains are indicated. Note the coherent reflective area in the southern parts of the shelf within the proximal domain indicating planar fault systems with likely strike-slip components. The necking domain is characterized by a nearly symmetric antiformal structure, whereas large-scale continent-dipping detachment surfaces can be observed in the foot of the slope and within the transitional domain. Further deformation, possibly low-angle faulting is present in the oceanic crust domain. (c) Crustal model with fitted gravity and magnetic anomalies. Other explanations as in Fig. 6. See Supplementary material Fig. 6 for close-ups to seismic.

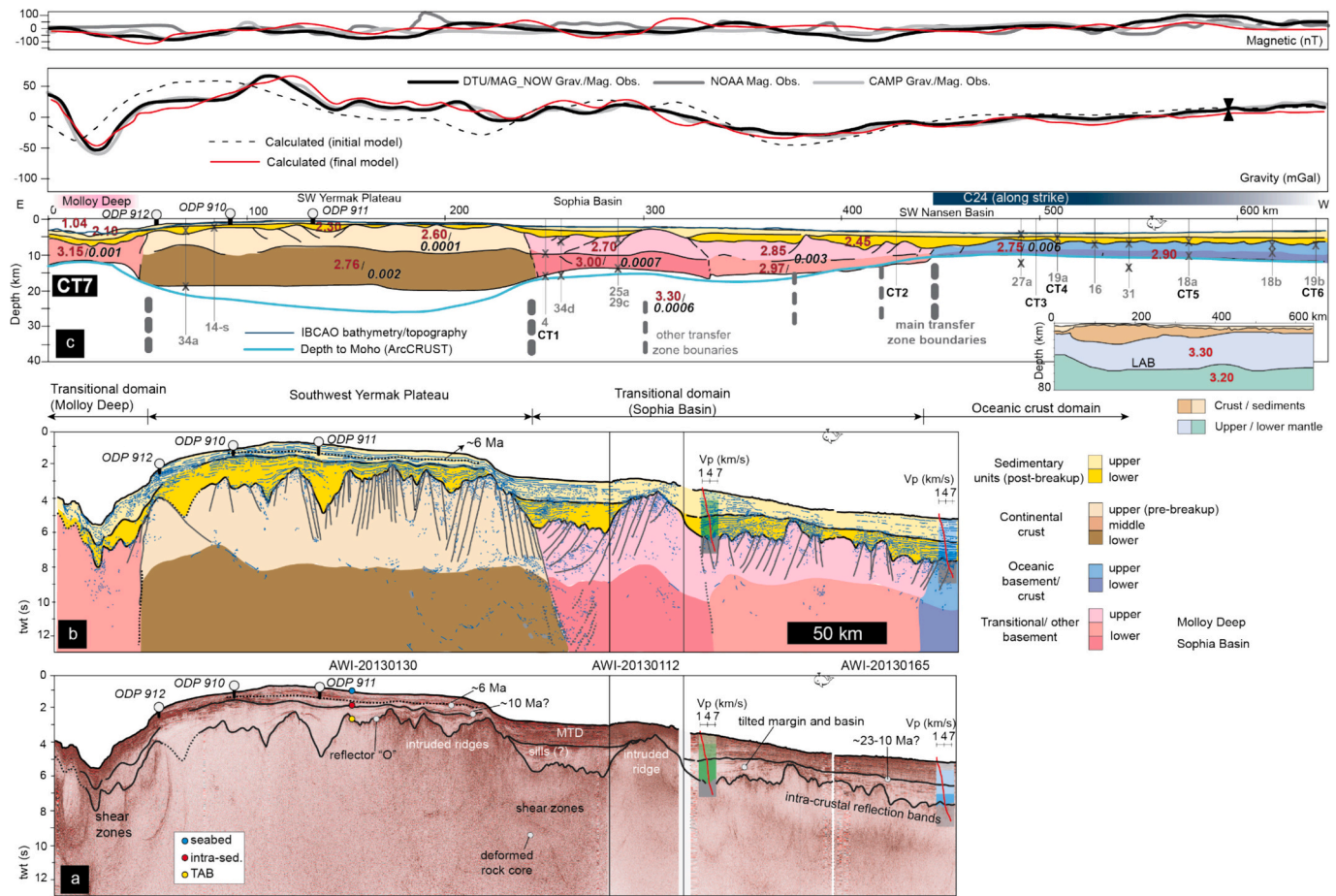


Fig. 12. Crustal transect 7 (CT7). (a) Composite seismic reflection profile crossing the Molloy Deep, the southern part of the Yermak Plateau, the Sophia Basin, and along the southwest Nansen Basin (seismic data of the AWI-2013 survey, [Quer, 2015](#)). CT7 model crosses the area (and the other crustal transects) in an E-W trend. The ODP drill sites 912, 910, 911 penetrate down to ~6 Ma sediments ([Mattingsdal et al., 2014](#)). See also ([Meza-Cala et al., 2024](#)) for a seismic-stratigraphic correlation of slightly older sediments retrieved in the Molloy Deep area (~10 Ma, ODP site 909; [Gruetzner et al., 2022](#)) in the west with respect to the seismic stratigraphy of the Nansen Basin ([Engen et al., 2009](#)) in the east (~23–10 Ma). (b) Geoseismic section interpretation with main units and structures. Rifted margin and crustal domains are indicated. Note the coherent reflective area in the crust between the Yermak Plateau and the Sophia Basin. (c) Crustal model with fitted gravity and magnetic anomalies. Projection of main and secondary possible transfer zones are indicated in gray-dashed lines. Other explanations as in [Fig. 6](#).

5. Discussion

5.1. Rifted margin and crustal domains

The mapping of the rifted and crustal domains is summarized in [Figs. 13 and 14](#). The geometries and structure of the architecture depicted in the seismic reflection data, together with the integration with potential field models reflect how and to what extent rifting processes affected the study area. In this study, we have adapted the terminology from the most recent study regarding the rifted margin evolution of the North Svalbard margin facing the Eurasia Basin, and thus have defined three main rifted (continental crust) domains across the area as ([Abdelmalak et al., 2024b](#)): 1) proximal; 2) necking; and 3) transitional domains. These domains have been mapped considering the general three sequential extensional modes of magma-poor systems, respectively ([Lavie and Manatschal, 2006](#)), but other relevant considerations from previous studies on archetypal magma-poor systems (e.g. [Boillot and Froitzheim, 2001](#); [Reston, 2009](#); [Mohn et al., 2012](#); [Péron-Pinvidic et al., 2017](#); [Loparev et al., 2021](#); [Gillard et al., 2017](#)) have been adapted to account for the specific characteristics of the study area. Therefore, the transitional domain defined here corresponds to a composite domain that includes the hyper-extended and magmatically-intruded continental crust and exhumed mantle of the outer and COT domains, respectively. An additional oceanic (crust) domain has been

included to describe the region characterized by seafloor spreading (e.g. [Whitmarsh et al., 2001](#)). The magnetic anomaly data helped to initially constrain a COB location at the oldest and regionally continuous chron C24 in the Eurasia Basin ([Brozena et al., 2003](#); [Glebovsky et al., 2006](#); [Minakov et al., 2012a](#)) and served as a starting test scenario for crustal modelling.

5.1.1. Proximal domain: The North Svalbard shelf

The proximal domain was mapped where deformation is accommodated by high-angle normal faults beneath a continuous, gently dipping (<10°), high-amplitude reflection character of the top acoustic basement. The continental crust typically has a thickness of >25 km in this domain based on interpretations of seismic reflection data (e.g. [Mohn et al., 2012](#)). The proximal domain is characterized by broadly distributed brittle deformation, and is followed by strain localization and crustal thinning along upper crustal faults (e.g., [Lavie and Manatschal, 2006](#)). The seaward limit of the proximal domain occurs where the top acoustic basement on the shelf presents an offset due to down-faulting with throws of more than ~1 s twt (< 2 km).

We define first the approximate landward limit affected by pronounced rifting leading to continental breakup at about ~80°N towards the eastern part of the study area, and thus the boundary of the proximal domain ([Fig. 14](#)). The definition of this landward limit is proposed based on geological observations from the surrounding regions east of

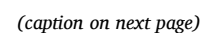


Fig. 13. Summary of the crustal models along the North Svalbard continental margin. The crustal models indicate the first-order rifted margin and crustal domains and are aligned with respect to the maximum horizontal gradient magnitude of the Bouguer gravity anomaly (Minakov et al., 2012a). An additional crustal transect model is presented for comparison (i.e., Cii; Cii adapted after Ritzmann and Jokat, 2003; and Ciib using seismic reflection survey AWI-20213, Quer, 2015; location in Fig. 14). Other abbreviations/explanations as in Figs. 1 and 14.

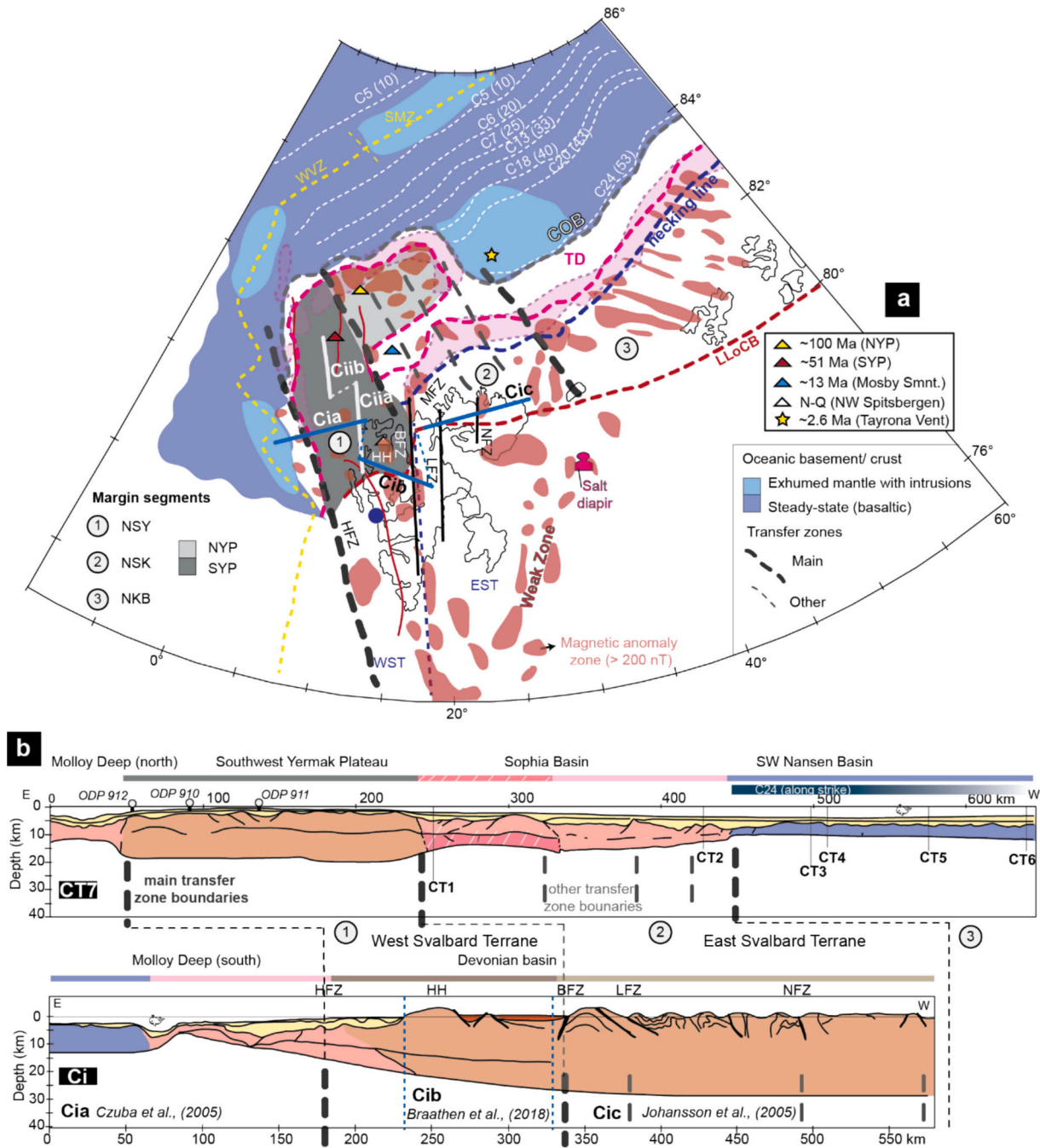


Fig. 14. Crustal and rifted margin domains map. (a) The map presents the summary of the structural interpretation of seismic reflection data and the integrated analysis of the modelling of crustal transects with gravity and magnetic data. An approximate Cenozoic landward limit of continental breakup (LLOCB) and thus of the proximal domain are indicated. Pink dashed line marks the landward limit of the transitional domain (TD). The continent-ocean boundary (COB) is indicated landward of the C24 anomaly in western Nansen Basin (see also text for discussion of COB location along the northern Yermak Plateau). The north-northwest trending possible transfer zone boundaries are also indicated and separate the proposed three main margin segments across the North Svalbard shelf, namely, (1) NW Spitsbergen-Southwest Yermak Plateau (NSY), (2) NE Spitsbergen-NW Kvitøya (NSK), and (3) NE Kvitøya-N Barents (NKB). HH, Hecla Hoek. Age constraints for volcanic products is indicated (Northeast/Southwest Yermak Plateau, NYP/SYP, Estrada et al., 2024; Mosby Smnt., Geissler et al., 2019; NW Spitsbergen and Tayrona Vent, Meza-Cala et al., 2024). Other map features as Fig. 1. (b) Simplified version of CT7 model (see location in Fig. 1) and a composite crustal transect (i.e., Ci) located south of and roughly parallel to that transect model (i.e. segmented blue lines Cia, Cib, Cic in the north of Svalbard; adapted after Czuba et al., 2005; Braathen et al., 2018; Johansson et al., 2005). The margin segmentation proposed in the offshore area is correlated to the mapped geology of North Svalbard. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

Nordautlandet and around Kvitøya that indicate today pre-Devonian (Hecla Hoek) basement exposures (Elverhøi and Lauritzen, 1984; Smelror et al., 2025; Fig. 14). This basement acts as a barrier for thick (up to ~5 km) and conformable late Paleozoic/Mesozoic basins truncated to the south of the proposed limit (e.g., CT4, Figs. 9 and 13). Seismic refraction data within the same area (Eldholm et al., 1984), gravity models (Minakov et al., 2012a), potential field data, as well as results from the compilation of the margin's crystalline basement and Moho structure (Figs. 2 and 4) also support a geophysical expression for the landward limit affected by strong deformation related to the time of continental breakup. Towards the western part of the North Svalbard margin, the landward extent of the proximal domain is diffuse because there is a lack of constraints of re-activation during breakup time. This is due to the highly deformed crust onshore North Spitsbergen caused by Eureka deformation and the interaction with Precambrian and Caledonian metamorphic basement core rocks (Braathén et al., 2018) (Fig. 14). Thus, the landward limit of the proximal domain may be extended further south near ~78°N and across the Devonian basin of Andrée Land Fault Block within pre-Devonian basement of North Spitsbergen (Grogan et al., 1999; Johansson et al., 2005; Smelror et al., 2025). These Paleozoic detachments may have been re-activated during the Oligocene in the offshore area at the southwestern Yermak Plateau, as constrained by the plate tectonic setting of West Svalbard (Ritzmann and Jokat, 2003).

An east-northeast trending fault system is observed in the shelf north of Spitsbergen (CT1 model; Fig. 6), and it is later flanked to the west by a north-northeast trending fault system associated to the sheared western margin of Svalbard (e.g., Eldholm et al., 1987; Faleide et al., 2024b). The apparent dominant fault orientations of the shelf northeast of Spitsbergen gradually become more north-northeast towards the northwestern part of Nordautlandet, locally shifting the main fault systems orientation into a west-northwest trending basin architecture (Geissler and Jokat, 2004). This geometry is also reflected in the sharp triangle-like shaped region observed on both the bathymetry and the potential field data (Figs. 3 and 14). In these regions, we indicate both transpression and transtension events characterized by positive and negative flower structures within pre-breakup seismic units (i.e., reflection packages underneath the top acoustic basement) (CT3–6, Fig. 13).

We also indicate the presence of evaporite-like features (i.e., salt diapir) within the shelves south of Nordautlandet and Kvitøya as they correlate well with the negative gravity field there (–20 mGal; CT4 model; Figs. 9 and 13). Available Paleozoic plate tectonic reconstruction models and seismic-stratigraphic frameworks of Svalbard and North Barents adjacent to the eastern shelves of our study area (e.g., Cocks and Torsvik, 2011; Lundschién et al., 2025) indicate the plausible occurrence of sedimentary environments prone to form salt deposits during the late Paleozoic (Fig. 3a). Seismic interpretation indicates salt extrusive features can be found distorting the seismic reflection image within the greater platform region in offshore Svalbard in addition to Cretaceous HALIP sill intrusions (Minakov et al., 2012b; Fig. 3b). However, the detailed analysis regarding the formation of such units or constrained timing of associated deformation remains yet to be determined. In the southern Barents Sea, salt deposits are of Carboniferous-early Permian age, but can act as kinematic indicators until Mesozoic and Cenozoic times via the reactivation of major NW-SE magnetic and basement trends within late Paleozoic grabens (Gernigon et al., 2018; Hassaan et al., 2020).

5.1.2. Necking domain: The shelf break region

Approximately, north of ~82°N an area defined by rapid thinning (from ~30 km down to ~25–15 km thick crust) corresponds to a narrow zone of 50–80 km-width associated with the maximum horizontal gradient magnitude of the Bouguer gravity anomaly (Minakov et al., 2012a) (Figs. 1, 13 and 14). The criteria for mapping the necking domain includes a down-step relief of the top acoustic basement at the continental shelf associated with the first large fault (i.e., throws larger

than ~1 s twt at the Moffen Fault Zone, Eiken, 1994; Fig. 1) and a reduction of crustal thickness to less than ~25 km (e.g. crustal taper geometry; Osmundsen and Redfield, 2011; Mohn et al., 2012) (Figs. 13 and 14). The crustal thickness is progressively reduced seawards as the Moho boundary progressively shallows and tends to converge with a deepening continental basement top (Fig. 13). The basin architecture observed in the shelf break north of Nordautlandet consists of half-graben depocenters oriented parallel to the direction of the ridges in a nearly east-west trend as depicted by our crustal models (CT2–6; Figs. 7–11). The infill age is speculative: no Upper Cretaceous units are present on Svalbard, but based on regional considerations farther east towards the North Barents Sea, the depocenters may contain Upper Cretaceous and Paleogene sequences probably related to the early opening of the Eurasia Basin (syn-rift units in Figs. 7 and 10) (e.g., Baturin et al., 1994; Geissler and Jokat, 2004; Engen et al., 2008; Kristoffersen et al., 2020; Abdelmalak et al., 2024b).

The structure of the shelf break northeast of Kvitøya suggests both a decrease in fault throws bounding half-graben basins (from more than 1 down to 0.5 s twt) and a low inclination of the top acoustic basement (Fig. 13). This character is interpreted as an erosional unconformity (e.g., Causer et al., 2020), signaling a gradual transition between the proximal and the necking domains (Figs. 14 and 15). The along-margin transition between the shelf northeast of Nordautlandet into northern Kvitøya is also marked by the presence of the prominent concave marginal high depicted in the CT5 model (Figs. 10 and 14). The top acoustic basement on the shelf has an opaque character masking reflectivity underneath (~1 s twt-thick units), and with modelled P-wave seismic velocities that reach up to ~6.0 km/s towards the shelf northeast of Nordautlandet (CT2 and CT3 models, Figs. 7 and 8; Czuba et al., 2005; Geissler et al., 2011). Our potential field modelling and interpretation of this domal feature suggests it is of volcanic origin (Fig. 10). The seismic character of this marginal high, as well as that of the offshore top acoustic basement towards some parts of the shelf north of Spitsbergen and Nordautlandet resemble to the volcanic layers located at the outer limit of the necking domain that formed before/during breakup times on the NE Atlantic region, including the southwest Barents Sea (Berndt et al., 2001; Tsikalas et al., 2012; Zastrozhnov et al., 2018; Gernigon et al., 2015; Gernigon et al., 2020; Meza-Cala et al., 2021). On the other hand, the shelf northeast of Kvitøya (CT6, Fig. 11) has the antiformal symmetric fault complex below the top acoustic basement (i.e., breakup unconformity). Based on the morphology and location of this antiformal feature, we suggest is related to brittle-ductile necking associated with exhumation processes in the crust and continental core complex formation (e.g., Brun et al., 2018). We anticipate that these processes occurred at the shelf break of the North Svalbard margin (e.g., CT4 and CT6; Figs. 9, 11 and 13), and reflect also different volcanic budgets on the marginal highs at the seaward limit of the necking zones (e.g., Fig. 3, CT3 and CT5 models; Figs. 8, 10).

5.1.3. Transitional domain and nature of the continent-ocean transition

The transitional domain is characterized by highly thinned continental crust (<15–10 km), formation of crustal-scale detachments along downward-concave faults, and high-density blocks (> 2900 kg/m³) (Fig. 13). In general, the transitional domain encompasses an area of negative amplitudes observed from the potential field anomalies which are oriented parallel to the outer shelves along the continental slope of the North Svalbard continental margin and the Yermak Plateau (Fig. 3) (Abdelmalak et al., 2024b). The Molloy Deep region and southwestern edge of the southwestern Yermak Plateau as modelled in this study (CT7 model; Fig. 12) is dominated by transcurrent faulting: strike-slip faulting was followed by oblique extension causing rapid/strong thinning of the crust along the western Svalbard and southwestern Yermak Plateau margins (e.g., Faleide et al., 1991; Faleide et al., 2024b) (Fig. 1). This area exhibits features that correspond to mantle exhumation processes, amagmatic crust production, and inside corner structures producing serpentinized mantle occurrences, like the areas in the vicinity at the

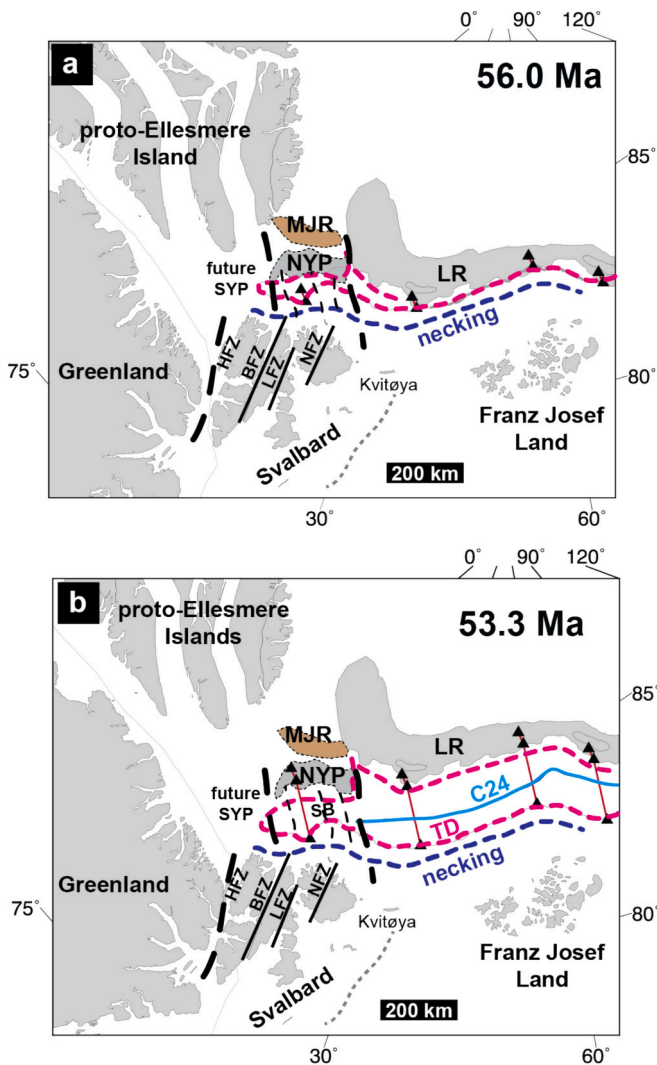


Fig. 15. Plate reconstruction of the Paleogene setting. (a) Late Paleocene at ~56 Ma. (b) Early Eocene at ~53 Ma (C24) with Gakkel Ridge's slow seafloor spreading initiated. The area occupied by the Sophia Basin (SB) and surroundings may involve multi-phase transtension/transpression. Sets of black triangles with connecting red lines indicate initial Eurasia Basin rift direction during breakup. The identified transfer zones must have acted as dextral shear accommodation regions throughout these plate kinematic models. Reconstructions show Europe (Eurasian Plate and Svalbard) fixed, and Lomonosov Ridge (LR) is moving as part of North America, while Northeast Yermak Plateau (NYP) and Morris Jesup Rise (MJR) are moving relative to Europe. Rotations derived by plate model of (Gaina et al., 2009; Gaina et al., 2002). The TD location along LR is approximately parallel to the lineament interpreted as C25 (Brozena et al., 2003; Glebovsky et al., 2006). Gray dashed line beneath Kvitøya tentatively indicates a weak zone between Caledonian (Svalbard) and (Caledonian-modified) Timanian basement fabrics to the west and east, respectively (e.g. Marelllo et al., 2013; Faleide et al., 2018). Other explanations of map features as Figs. 1 and 14. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

Lena Trough and northernmost Knipovich Ridge (e.g., Ljones et al., 2004; Ritzmann et al., 2004; Ritzmann et al., 2002), or the Southwest Indian Ridge (e.g., Sauter et al., 2013). We also expect a similar composition of the crust within the transitional domains at the flanks along the Yermak Plateau (Czuba et al., 2005; ~50–80 km-wide zone) based on the structural setting interpretation and modelling results (CT1 and CT7 models; Figs. 6, 12–14).

The transitional domain mapped in the area from northwest Nordaustlandet to north of Kvitøya facing directly into the southwestern

Nansen Basin corresponds to an up to ~100 km wide zone up to the C24 anomaly (Fig. 14). In magma-poor margins, the COT region, part of the transitional domain in this study, is characterized by exhumation of sub-continental mantle (e.g. Mohn et al., 2012). Decoupling and hyper-extension of the continental crust relative to the mantle lithosphere is one of the mechanisms to explain this exhumation (e.g., Huismans and Beaumont, 2011). The crust in the transitional domain gradually may become intruded with magmatic sills as it reaches a more seaward location in proximity to a shallow asthenosphere indicating the onset of seafloor spreading (i.e., at the COT, e.g., Gillard et al., 2017). In addition, available models for the lithosphere-asthenosphere boundary, LAB, (e.g., Klitzke et al., 2015) indicate strong undulations with a shallowing of the lower mantle lithosphere structure (up to ~50 km depth) below the transitional domain region in the vicinity (Fig. 4).

An initial opening phase of the Eurasia Basin at ~58 Ma has been proposed by (Brozena et al., 2003) who identified a linear magnetic anomaly formerly interpreted as chron 25 (C25) (Figs. 1 and 3) (e.g. Døssing et al., 2014). However, this anomaly is patchy, discontinuous, and only better preserved on the Amundsen Basin (Glebovsky et al., 2006; Nikishin et al., 2018). In the western Amundsen Basin, recent seismic velocity and gravity models indicates that the first seafloor spreading anomaly is C24, linked to the breakup time in NE Atlantic, and that the formerly interpreted magnetic lineation C25 may be rather associated to exhumed and serpentinized mantle (Funck et al., 2022). We therefore argue that the identification of “C25” in the Nansen Basin belongs to the transitional domain indicating unroofing of mantle as our models also suggest (Figs. 13 and 14). Within the transitional domain and the narrow rift mode between the Lomonosov Ridge and Barents-Kara shelves, an upwelling divergent flow model predicts exhumation of 50 km of the continental mantle lithosphere and the lower crust (Minakov et al., 2012a). These mechanisms can explain the presence of margin parallel ridges imaged by seismic reflection data on both western Nansen (Lutz et al., 2018; this study; Fig. 11) and Amundsen (Funck et al., 2022) basins, resembling the breakup processes of other magma-poor margin systems (e.g., Iberia-Newfoundland, Shillington et al., 2006; Australia-Antarctica, Gillard et al., 2015).

The transitional domain on the northeast part of Kvitøya presents a morphology of the top acoustic basement reflection with a rather gradual relief configuration transitioning from the relatively flat character across the shelf as it drapes pre-breakup units towards the slope (Fig. 11). Once at the deep portions of the Nansen Basin, large continent-dipping surfaces are observed near the foot of the slope within the transitional domain and reaching deep into the lower crust and possibly upper mantle (> ~8 s twt) (Lutz et al., 2018) (Figs. 10 and 11). These features are similar to the so-called “flip-flop” structures (or detachment faults; Boillot and Froitzheim, 2001; Reston and Ranero, 2011) of exhumed mantle blocks documented in the COT region of the magma-poor setting of the Australian-Antarctic conjugate margin (e.g., Gillard et al., 2015). The shift from the necking into the transitional domain north off Kvitøya is seen as an abrupt one (i.e., sharp basement down-step, Fig. 10), likely associated to shearing processes operating during breakup (Minakov et al., 2013). In addition, there is a larger positive response of the gravity anomaly (~40–60 mGal) partly across the western edges of the Franz Josef Trough Mouth Fan (e.g., Lasabuda et al., 2018) (Fig. 3a), coincident with the area of probable mantle ridges having intra-crustal reflection bands associated to density and structure typical of serpentinized rocks (e.g., “L-reflector” in the transitional domain of Newfoundland, Lau et al., 2006, or West Iberia, Boillot and Froitzheim, 2001; CT5–7 models, Figs. 10–12). Our results thus support earlier studies that described in the vicinity tectonic regimes with serpentinized exhumed mantle (Minakov et al., 2012a; Minakov et al., 2013; Lutz et al., 2018).

5.1.4. Oceanic domain: The Nansen Basin and delineation of the continent-ocean boundary

The identification of the continent-ocean boundary (COB) within the

Nansen Basin and in the presumed magma-poor rift setting of the North Svalbard margin (Geissler and Jokat, 2004; Minakov et al., 2012a; Lutz et al., 2018) is controversial considering it is not necessarily a distinct boundary but instead a diffuse zone. In other magma-poor settings, the COB may correspond to an area that sits beyond in a seaward location away of (1) the plausible exhumed mantle ridges, (2) the magnetic isochrons with an exhumed mantle related-signal adjacent to the transitional domain regions, or (3) the magmatic underplating of the continental lithosphere before and during breakup (e.g., Boillot and Froitzheim, 2001; Lau et al., 2006; Bronner et al., 2011; Dean et al., 2015; Druet et al., 2018; Geoffroy et al., 2022). Other possible complications for the location of the COB can also include (1) a ridge jump or re-location (Shillington et al., 2006), (2) un-recognized complete tectono-stratigraphic evolution of rift sediments and the character of the top acoustic basement (i.e., breakup unconformity; Péron-Pinvidic et al., 2007; Causer et al., 2020), (3) the presence of magmatically-reworked extensional allochthons of continental origin in the proximal oceanic basin edges (i.e., “H-blocks”, Lavier and Manatschal, 2006; “C-blocks”, Geoffroy et al., 2022), and (4) old continental mantle lithosphere underneath oceanic-like crust (Gillard et al., 2017).

Here we consider the COB as a linear landward limit of the oceanic crust where a clear and continuous magnetic anomaly can be identified, or the outer limit of the transitional domain (Minakov et al., 2012a; Faleide et al., 2024a; Abdelmalak et al., 2024b; Fig. 14). This oceanic crust may be anomalously thin and deformed, like in southwestern Nansen Basin (Fig. 13), with characteristics of the so-called proto-oceanic domain such as in other magma-poor regions (e.g., Wilson et al., 2003; Lau et al., 2006; Gillard et al., 2015; Gillard et al., 2017; Péron-Pinvidic et al., 2017; Druet et al., 2018). Our modelling approach for the oceanic crust captures the first-order gravity field given the available datasets (Figs. 6–12). In this sense, the oceanic crust of the Nansen Basin is seen as a region of a broad stable gravity responses of less than 40 mGal from anomaly C24 and oceanward (e.g., Weigelt, 1998; Weigelt and Jokat, 2001; Glebovsky et al., 2013) (Fig. 3a). Thus, we define the COB in the southwestern Nansen Basin landward of the Eurasia Basin's oldest unambiguous seafloor spreading anomaly C24 (Brozena et al., 2003; Glebovsky et al., 2006; Funck et al., 2022; Fig. 3b). The COB continues to the west following the Northeast Yermak Plateau morphology. North of this plateau, the COB may be located just south of Chron C7 (the oldest continuous isochron identified there, Figs. 3b and 14) within a slightly denser ($>2900 \text{ kg/m}^3$; CT1; Fig. 6) and adjacent to a probably thicker $\sim 7 \text{ km}$ crust interpreted as a transitional region in our modelled CT1 transect (Jackson et al., 1984; Jokat et al., 1995; Engen et al., 2009). Oceanic crust no older than C13 may be present just $\sim 70 \text{ km}$ east from the location of the CT1 transect at the crossing area of the Northeast Yermak Plateau (Jokat et al., 2016), and near the location of the FRAM-IV seismic experiment in the $\sim 4\text{--}5 \text{ km}$ -thick oceanic crust of the northwest Nansen Basin (Kristoffersen and Husebye, 1985; Duckworth and Baggeroer, 1985; Figs. 2–3 and Table 2).

The top acoustic basement on the deep-water setting ($> 2500 \text{ m b.s.l.}$) defines a clear interface imaged by seismic reflection data in the oceanic crust domain of the Nansen Basin (Fig. 13). The oceanic top acoustic basement reflector in the northwestern and southwestern parts of the Nansen Basin indicates a seismic interval velocity exceeding 4.60 km/s (Jokat et al., 1995; Jokat and Micksch, 2004; Geissler and Jokat, 2004; Engen et al., 2009; Fig. 5 and Table 2). The top acoustic basement interface in the Nansen Basin corresponds to the mapped top of the oceanic layer that consists of $\sim 3\text{--}5 \text{ km}$ thickness modelled for all crustal transects (Fig. 5). In general, the internal seismic reflectivity of the oceanic crust within the southwestern Nansen Basin (Figs. 6–12) may indicate local structural or geochemical layering with possibly a two-layered crust (e.g., interface between oceanic Layer 2 and Layer 3 or the base of serpentinization reactions and top of fresh peridotite) within the apparent oceanic basement (e.g., Newfoundland, Shillington et al., 2006; Labrador Sea, Chalmers and Laursen, 1995; Australian-Antarctic conjugate margin, Gillard et al., 2015). Such oceanic crust layering

may relate to the emplacement of serpentinized exhumed mantle inferred from gravity modelling (Minakov et al., 2012a) and the slow/ultra-slow seafloor spreading mode of the Gakkel Ridge (Lutz et al., 2018; Sokolov et al., 2023) (Figs. 13 and 14). Furthermore, the seismic interval velocity model used for depth migration within the southwestern Nansen Basin may indicate values exceeding 6.80 km/s below the top acoustic basement reflector and a structure dominated by fault-block rotation and exhumed mantle ridges at the western parts of seafloor spreading anomalies C24 to C21 (Berglar et al., 2016; Lutz et al., 2018; Meza-Cala et al., 2024; this study; Figs. 8–12). Recent studies regarding the formation of oceanic crust within the conjugate Amundsen Basin looked at seismic velocity models and oceanic basement roughness. These studies indicate the presence of one, or a combination of both, oceanic Layer 2 and Layer 3, suggesting an asymmetric spatio-temporal evolution of magma extraction along the Gakkel Ridge within the oceanic crust in Eurasia Basin (Nikishin et al., 2018; Funck et al., 2022; Castro et al., 2024; Tan et al., 2024).

We have identified internal reflectivity within the oceanic basement in parts of the southwestern Nansen Basin (Figs. 13 and 14) that may indicate probable proto-oceanic regions (i.e., deformed crust and exhumed mantle) rather than the so called steady-state oceanic crust (i.e., basaltic ridges) (see also Berglar et al., 2016; Lutz et al., 2018; Nikishin et al., 2018; Funck et al., 2022; Meza-Cala et al., 2024). The density for the oceanic crust modelled in the crustal transects ($2750\text{--}2900 \text{ kg/m}^3$; Table 3) and the upper and lower units in the oceanic domain is expected to represent oceanic Layer 2 and Layer 3, respectively, with dominance of exhumed mantle towards the southwestern corner of the Nansen Basin as indicated by seismic reflection and magnetic data (Fig. 14). Locally, wedges below the top acoustic basement can be observed resting at the landward edge of C24 in the eastern crustal profiles (CT5 and CT6 models) and may represent volcanic rocks deposited by the time of early oceanification in this area (Figs. 10 and 11). The magnetic anomaly responses appear better differentiated and with greater amplitudes following a northward and eastward trend within the Nansen Basin (Figs. 6–12), possibly indicating strong focused hydrothermal activity in the opposite southwestern parts of the basin throughout the Cenozoic (Meza-Cala et al., 2024). Alternatively, considering the presence of wedge geometries in the eastern part of the studied Nansen Basin area, an explanation for the magnetic anomaly responses could be due to an increased magma-budget that was able to surface following the same orientations from breakup time (e.g., similar to the associated process for the formation of the “J-anomaly” in the Iberia margin; Bronner et al., 2011) and during early seafloor spreading processes (Figs. 10 and 11).

5.2. Margin segmentation

We observe segmentation locally at the North Svalbard margin, specifically across the Yermak Plateau, and the necking and transitional domains in the shelf north of Spitsbergen, Nordaustlandet, and Kvitøya (Fig. 14). The segmented architecture of the North Svalbard margin is inferred from the trends in gravity-magnetic data (Fig. 3), the orientation and distribution of half-graben basins (Figs. 13–15), the top acoustic basement seismic character and its relationship to shift in polarities of the largest fault systems ($> 0.5\text{--}1 \text{ s}$ twt fault throws) on the shelf and transitional crust regions (CT1–7 models; Figs. 6–12), as well as from other crustal models depicting the general margin configuration (Figs. 2 and 4). Therefore, we have defined three main margin segments within the study area, namely the (1) Northwest Spitsbergen-Southwest Yermak Plateau (NSY), (2) Northeast Spitsbergen-Northwest Kvitøya (NSK), and (3) Northeast Kvitøya-North Barents (NKB) (Fig. 14). These margin segments are separated from each other by proposed north-northwest trending transfer zone boundaries which likely represent inherited structures of Caledonian and Timanian origin from west to east, respectively (Fig. 14). Each of these elements are described and discussed below, including their correlation to the onshore geology of

Svalbard and other geophysical trends further south.

5.2.1. Northeast and Southwest Yermak Plateau

The Yermak Plateau is segmented into two juxtaposed crustal blocks (Fig. 14) based on our modelling results (Figs. 6 and 12) – Northeast Yermak Plateau (NYP) and Southwest Yermak Plateau (SYP). The dual origin of the Yermak Plateau has been conceived also in previous work as the plateau segments reflect different tectono-magmatic histories (e.g., Crane et al., 1982; Jackson et al., 1984; Kristoffersen et al., 2020; Faleide et al., 2024b). Earlier studies have explored the evolution of the Yermak Plateau and have concluded that both the seismic and thermal structures differ on the two segments delimited in this study (Fig. 14), as the SYP is at least ~20 Myr younger than the NYP (Crane et al., 1982; Sellevoll et al., 1991). Heatflow data and thermal modelling of the crust within the transition area between the SYP and the Sophia Basin have also predicted a heating episode not older than ~16–13 Ma (Crane et al., 1982). Interestingly, this model agrees with the reported late Cenozoic volcanic activity at the Mosby Seamount of the Sophia Basin in the vicinity (also ~13 Ma intrusive event; Geissler et al., 2019; Fig. 6). A new compilation of geochemical data within the Yermak Plateau supports segmentation of this structure through dating of dredged volcanic rocks (Estrada et al., 2024), as the NYP samples are older (~100 Ma; Fig. 6) than the SYP ones (~51 Ma; Riefstahl et al., 2013).

The eastern part of the NYP has been suggested to have formed synchronously with the seafloor spreading anomalies C22 to C18 in the Nansen Basin as there is a correspondence between the trends of the plateau and the magnetic chrons (~mid-Eocene times; Kristoffersen et al., 2020). Certainly, the magnetic data indicates the presence of several line-ridge features aligned with the same trend of the Gakkel Ridge's axis (Fig. 3b). Although the transect CT1 is at the western edge of the NYP (Fig. 14), the presence of folded and deformed structures beneath a flat geometry character of the top acoustic basement and the gravity modelling suggest a continental origin in this area (Fig. 6). Portions of the NYP may record Eocene Eurekan deformation (53–34 Ma; Dörr et al., 2012), which is an important conclusion because it suggest that the Eurekan event likely affected a larger area than previously anticipated (Berglar et al., 2016). According to the age of ~100 Ma and the geochemical signature of the volcanic sample and other magmatic additions inferred from geophysical data that are present in the NYP (Figs. 3, 6), this pulse might be related to a late or delayed (~30–20 Myr) expression of HALIP-style magmatism (Estrada et al., 2024), considering the area of North Svalbard is prone to rejuvenation of melt phases beneath thinned crust in association with plume-like activity (Heyn et al., 2024). Furthermore, the Yermak Hotspot model (Fedene et al., 1979) suggests a thermal anomaly in the mantle since at least ~60 Ma with strong magmatic pulses across the NYP during the time intervals of seafloor spreading anomalies C18 to C13, and when the Morris Jesup Rise is proposed to have been attached to the Yermak Plateau, specifically the NYP segment, until the end of Oligocene times (~33–26 Ma) (e.g., Kristoffersen and Husebye, 1985; Kristoffersen et al., 2021; Brotzer et al., 2022; Estrada et al., 2024; Fig. 15). A comparison with numerical models of the evolution of localized ascending thermal anomalies in the mantle within continental rifting and breakup in ultra-slow extensions settings (i.e., Koptev et al., 2018) might provide some insights for the present-day sharp bathymetric and geophysical expressions of the Morris Jesup Rise and NYP (e.g., Fedene et al., 1979; Jokat et al., 1995; Weigelt and Jokat, 2001; Jokat et al., 2008; Engen et al., 2008) (Figs. 1 and 15).

Some of the first geophysical studies conducted within the SYP segment indicated a top acoustic basement with an opaque character across the Sverdrup Bank (Fig. 1), which was seen occasionally masking reflectivity underneath sharp or gentle relief ridge topography (i.e., the seismic reflector “O”; Sundvor and Eldholm, 1979; Jackson et al., 1984; Geissler and Jokat, 2004) (CT7 model; Fig. 12). Thus, the seismic reflection data was used to suggest a maximum landward extent of the oceanic crust (i.e., COB contour in SYP by Sundvor and Eldholm, 1979;

Sundvor and Austegard, 1990; Fig. 1). The top acoustic basement in the SYP marks a boundary for P-wave seismic velocities that do not exceed ~4.5 km/s (Jackson et al., 1984), whereas it can be up to ~6.7 km/s for the NYP in the transition towards the Nansen Basin (Jackson et al., 1982) (Figs. 5 and 6). Therefore, based on the seismic character and velocities, as well as the position within the distal parts of the margin of the top acoustic basement in the SYP, the seismic reflector “O” was associated with Paleogene breakup processes (Sundvor and Eldholm, 1979).

Dredged fragments of quartzites and gneisses (Jackson et al., 1984), together with the geophysical signatures from seismic reflection and potential field anomaly responses (Jokat et al., 2008) within the SYP, hinted at a connection between this part of the Yermak Plateau and Svalbard margin to the Devonian sediments and pre-Devonian basement onshore NW Spitsbergen (Figs. 1, 3, and 14). In principle, our results indicate thinned crust in the SYP (Fig. 13), also thinner with respect to the NYP (CT7 model, Fig. 12). Similarly, seismic refraction studies along the southern parts of the SYP (Ritzmann and Jokat, 2003; Czuba et al., 2005; Figs. 2 and 14 and Table 2) supported the notion of continental crust thinned with contemporaneous shearing and volcanism during Paleogene. The structure likely followed a north-northeast trend with horst-and-graben systems (e.g., Danskøya Basin and Sverdrup Bank regions; Geissler and Jokat, 2004; Fig. 14) linked to the southwest with the Hornsund/De Geer Fault Zone (Kristoffersen et al., 2020; Faleide et al., 2024b; Fig. 3), whilst in the southeast the trend likely followed the fault system onshore Northwest Spitsbergen and west of the Billefjorden Fault Zone (Braathen et al., 2018) (Fig. 14). In this sense, thinning within the SYP mainly developed through post-Eocene oblique extension (Fig. 13). Thus, the SYP is interpreted as a crustal segment related to the northernmost continuation of the sheared margin west of Svalbard (Faleide et al., 2024b; Estrada et al., 2024) and to the re-activation of Devonian detachment structures in Northwest Spitsbergen (Ritzmann and Jokat, 2003; Jokat et al., 2008) (Fig. 13).

5.2.2. Rift-shear architecture of the North Svalbard margin

The first main margin segment is identified as the Northwest Spitsbergen-Southwest Yermak Plateau (NSY), and it is mapped across the area between the northward projection of the Hornsund/De Geer Fault Zone and the Billefjorden Fault Zone (Fig. 14). In the earlier discussion of the Southwest Yermak Plateau (SYP; Fig. 13), we established the tectono-magmatic constraints for the segmentation of this part of the North Svalbard margin, and we suggest that the above-mentioned fault zones act as possible transfer zone regions. The half-graben basins along the western edge of the NSY segment (i.e., west of Danskøya Basin region; Fig. 14) reflect the latest development of depocenters associated to the shearing and separation phases between Svalbard and Greenland, following an old-to-young (late Oligocene-to-Miocene) direction from South to North over mostly Caledonian basement fabrics (e.g., Faleide et al., 2024b; Abdelmalak et al., 2024a).

The second main margin segment is identified as the Northeast Spitsbergen-Northwest Kvitøya (NSK), and it consist of a composite region, including north of Nordaustlandet (Figs. 13–14). The NSK segment is defined here as a composite segment because of (1) the good correlation between oblique (north-northwest) shifts in potential field data and onshore (northerly trending) fault zones (Fig. 3), and (2) the similar orientation and architecture of the main fault-bounded basins mapped in this study (e.g., crustal transects CT1 to CT4; Figs. 6–9). Although one may differentiate additional segments within the NSK segment, we acknowledge the greater complexity of the shelf north of Nordaustlandet representing probably an amalgamated and deformed Caledonian terrane (Fig. 14). Our results indicate, in general, a similar crustal structure across the necking domain in the NSK segment (Fig. 13).

The necking region of Spitsbergen (western NSK segment) may represent an extensional collapse-like architecture of pre-breakup topography (Fig. 6) that accommodates deformation east of the sheared margin of West Svalbard and the NSY segment. Moreover, the

northward projected location of the Lomfjorden Fault Zone in eastern Spitsbergen has a segmented potential field anomaly pattern on the necking domain (i.e., the second ridge from west to east in the gravity data; Geissler and Jokat, 2004) (Figs. 3 and 14). Moving east, a trace projected towards the north of a major fault zone across the central part of Nordaustlandet (i.e., Nordaustlandet Fault Zone, Fig. 1; Mitchell et al., 1990; Dörr et al., 2012) marks the eastern termination of the Moflen Fault Zone. A northerly continued trace of the Nordaustlandet Fault Zone may establish an additional transfer boundary at the narrowest areal extent of the Sophia Basin at a magnetic anomaly shift (Fig. 3b) in-between prominent line-ridges (Fig. 14). The along-margin prolongation of the outer necking domain consists of a marginal ridge (CT2–3; Figs. 7 and 8) that progressively decreases in size on the shelf north of Kvitøya (CT4; Figs. 9 and 13).

The area between northeast of Nordaustlandet and northwest of Kvitøya juxtaposes the eastern edge of the Yermak Plateau's NYP against the transitional domain and oceanic crust of the Nansen Basin (CT7; Figs. 12 and 14). Such behavior of the potential field data occurs over a zone less than ~100 km-wide (Figs. 3 and 15), and is indicative of a narrow rift mode with rapid embrittlement of the crust (e.g., Tsikalas et al., 2005; Reston, 2009), possibly related to short duration shearing and oblique extension mechanisms during breakup (Minakov et al., 2013). The NYP and SYP may have been in tectonic contact with the transitional domain of the Sophia Basin region and the oceanic crust of the Nansen Basin along a transform boundary (Figs. 4, 12–14). Within the oceanic crust northeast of the NSK segment, the transfer zone boundary may have evolved as a divergent-transform setting associated with the formation of inside-corner features and lower crust/mantle upwelling during early seafloor spreading (e.g., Reston and Ranero, 2011; Sauter et al., 2013; Druet et al., 2018; Basile, 2015; Le Pourhiet et al., 2017; Balázs et al., 2023), such as in the nearby western vicinity of the Molloy Deep (CT7 model; Fig. 12). The southwest Nansen Basin is likewise interpreted as exhumed mantle with elevated basement relief at the western edges of seafloor spreading anomalies C24 to C21 (see seismic reflection profiles of Berglar et al., 2016), which may serve as a laboratory to further explore the above tectono-magmatic processes (Meza-Cala, 2025).

We indicate a main final transfer zone across Kvitøya, which indicates the presence of the Northeast Kvitøya-North Barents (NKB) margin segment (Fig. 14). In this area, the bathymetry indicates an apparent abrupt hinge morphology at the shelf break north of Kvitøya (Figs. 1 and 15) and at the location of the concave marginal high in the outer necking region (CT5 model; Fig. 10). The hinge area also corresponds to a shift in HALIP dyke trends (Minakov et al., 2018) (Fig. 3b), supporting the role of the Kvitøya area as a major weak zone in the crust and mantle (Marello et al., 2013) and indicating a more complex basement besides of Caledonian origin (Abdelmalak et al., 2024a) (Fig. 14). Although there is a chaotic, down-bending character of intra-crustal reflectivity beneath the concave marginal high in the shelf break of Kvitøya, these geometries appear in a coherent location in-between the downfaulted blocks (CT5; Fig. 10). This supports the presence of major shear tectonic features in this area, and is comparable to the seismic velocity structure seen at West Svalbard margin (Kristoffersen et al., 2020). In the NKB segment, the crustal architecture has a sharp decrease of the Moho depth (Minakov et al., 2012a; CT5 model; Fig. 10), exhibiting similar crustal-scale marginal ridges of oblique rifted margins with super-imposed shearing and associated deformation (e.g., Mohriak et al., 2008; Antobreh et al., 2009; Nemčok et al., 2016; Le Pourhiet et al., 2017; Loparev et al., 2021; Balázs et al., 2023). Finally, the Kvitøya area marks the eastward onset of well-developed continent-dipping detachment fault surfaces related to mantle exhumation in the transitional domain on seismic profiles (Lutz et al., 2018; CT5–6 models; Figs. 13 and 15).

5.3. Basement controls on final margin configuration

5.3.1. On the origin of the Sophia Basin

The Sophia Basin has a top acoustic basement that consists of a seismic marker draping down-faulted sequences of deformed early basin fill in the upper crustal unit with top-to-south orientations, which are possible to observe in the vicinity of the Mosby Seamount (e.g., 150 km profile distance in sag region of CT1 model; Fig. 6). This structural trend is confirmed along strike via down-faulted top acoustic basement topography with top-to-west orientations (CT7 model; Fig. 12). The ~5–7 km thick crust has a deep seismic velocity structure (>7 km/s below 8 s twt; Fig. 5) modelled with high-density (3000 kg/m³) blocks (Figs. 6 and 12). Previous studies indicated a transitional nature for the basement of the Sophia Basin with intruded and (hyper-)extended continental crust (Geissler and Jokat, 2004; Engen et al., 2008; Kristoffersen et al., 2020; Abdelmalak et al., 2024b). Magnetic data (Fig. 3b), CT7 model interpretations (Fig. 12), and the seismic velocity structure of the lower crustal unit (>7 km/s, Quer, 2015; rock core label in CT1, Figs. 2, 5 and 6, and Table 2) suggest the presence of intruded ridges and volcanic additions within, and in the vicinity of, the Sophia Basin. However, the absence of extensive magmatism (Geissler and Jokat, 2004; Minakov et al., 2012a) makes a fully-magmatic underplating scenario perhaps not totally applicable to this area. Instead, the structure of the Sophia Basin crust indicates that this marginal basin has the geophysical character (i.e., high velocity-density and collapsed-like topography) of basement rocks composed of medium-to-high-grade metamorphic conditions (Figs. 6 and 12). For instance, these rocks may be comparable to modified granitic blocks with some mafic granulites or amphibolites (e.g., see velocity-density charts in Breivik et al., 2002; Shulgin et al., 2020), and/or to serpentinized mantle ribbons formed by exhumation during drastic thinning of the crust (e.g., Iberia margin transition zone, Dean et al., 2000; Bjørnøya Basin in SW Barents Sea, Gernigon et al., 2014; NW Spitsbergen transition zone, Czuba et al., 2005).

The deeper Sophia Basin structure consists of shear zones down to ~6 and 8 s twt (~8–10 km depth), some acting as detachment surfaces within the high-density blocks (CT1 and CT7; Figs. 6 and 12). These structures contain a similar crustal configuration, as well as they are on trend to the regional north-south directed Caledonian-style deformation of Svalbard and the SW Barents Sea (e.g., North Svalbard: Johansson et al., 2005; Dörr et al., 2012; Braathen et al., 2018; Michalski et al., 2023; SW Barents Sea and northern Norway: Breivik et al., 2005; Ritzmann and Faleide, 2007; Gernigon et al., 2014; Shulgin et al., 2020; Gabrielsen et al., 2022; Fig. 14). However, the Sophia Basin crustal units exhibit low magnetic susceptibility (0.0007 SI units; Table 3) with low-to-negative magnetic signatures (< -100 nT) (Figs. 6 and 12), contrasting sharply with the high magnetic susceptibility (>0.02–0.033 SI units) modelled in the deep lower crust of the SW Barents Sea, attributed to remanent Caledonian bodies and associated tectonic sutures (Marello et al., 2010; Marello et al., 2013; Breivik et al., 2002; Breivik et al., 2003; Breivik et al., 2005; Klitzke et al., 2019; Shulgin et al., 2020; Gabrielsen et al., 2022). Nonetheless, non-magnetic Barentsia pre-Carboniferous blocks and Caledonian allochthons that formed along the Baltic margin may explain these characteristics (Marello et al., 2013; Gernigon et al., 2014). The high magnetic susceptibility region associated with the Caledonian suture in the SW Barents Sea extends from northern Norway across the entire West Barents Sea to Svalbard, delineating regions in western Spitsbergen along the west of the Billefjorden Fault Zone, and further to the east across the weak zone in Kvitøya coming from the south (Breivik et al., 2005; Marello et al., 2013) (Figs. 1 and 14).

Moreover, pre-Eurekan (~80–75 Ma) to Eurekan (~56–35 Ma) phases could have exploited the north-south structural grain of West and NW Svalbard (Meier et al., 2024). These would have ultimately created an intra-continental region in the Sophia Basin area that later collapsed and widened as both shearing along the Svalbard-Greenland conjugate margin led to the NE Atlantic-Arctic connection in the west, while

opening of the Eurasia Basin evolved in the east (e.g., Engen et al., 2008; Faleide et al., 2024a; Faleide et al., 2024b) (Figs. 14 and 15). This latter hypothesis may be interesting since parts of the Northeastern Yermak Plateau (NYP) contains candidate Eurekan-style deformation features (Fig. 6), and the trend of volcanism appears to become younger from the NYP southwards towards the NW Spitsbergen, including parts of the SW Nansen Basin (Fig. 14). This may also explain why the aligned line-ridge features in the magnetic anomaly data decrease in intensity within the Sophia Basin following the same southwards trend (Fig. 3b). Our results favor the Sophia Basin region to be interpreted as a felsic basement block in the upper crustal unit with few volcanic additions and pre-breakup sequences (e.g., meta-sediments; like in onshore North Svalbard, Fig. 14), and a combination of granulite-amphibolite rock facies and/or serpentinized sub-continental mantle within the modelled lower crustal unit.

5.3.2. Oblique opening setting

Typically, rifting initiates along inherited structures in passive margins and transform settings (Balázs et al., 2023). Additionally, the presence of rigid/stiff crustal blocks can deflect rifting to more favorable opening directions (Vauchez et al., 1998; Tommasi and Vauchez, 2001). In Svalbard and central northern Barents Sea regions (i.e., Franz Josef Land; Fig. 1), both Caledonian and Caledonian-overprinted Timanian basement domains, respectively (Abdelmalak et al., 2024a), have had visible effects in terms of distribution and orientations of rift basins (Klitzke et al., 2019; Shulgin et al., 2020), as well as in the initial emplacement conditions of igneous complexes (Marello et al., 2013; Minakov et al., 2018; Fig. 14). The thicker (>25 km) continental crust areas within south of Kvitøya (Figs. 4 and 13), may have influenced Paleozoic to Mesozoic rifting and salt basin development in the area (Fig. 9). Such thick crustal buffers could have controlled rift basin configuration leading to the preferential development of pre-salt basins and evaporitic depocenters at the edge of these blocks (i.e., at the location of the salt diapir; Fig. 14), similar to those in the southern Barents Sea (Gernigon et al., 2018; Gernigon et al., 2014).

Tectonic stress associated with the Eurekan phases on Svalbard is suggested to be mostly concentrated in the West Svalbard Terrain (WST; Gac et al., 2020; Meier et al., 2024; Fig. 1) along presumably supra-detachment structures of Northwest Svalbard (Ritzmann and Jokat, 2003; Braathen et al., 2018; Fig. 14), and possibly including the Sophia Basin region (Fig. 15) (Geissler et al., in preparation). The high horizontal gradient of the Bouguer anomaly borders only the flanks of the Yermak Plateau and the shelves in the East Svalbard Terrain (EST; Figs. 1 and 3), which is accompanied by an oceanward shift from the necking region into parts of the transitional domain towards the east (Minakov et al., 2012a; Fig. 13). In this context, the evolution of the Sophia Basin may consist of two phases: (1) Paleocene oblique rifting towards the basin edges and the North Svalbard shelf (Geissler and Jokat, 2004; Engen et al., 2008) (Fig. 15), and (2) Eocene-Oligocene sag basin development, probably guided by re-activated Devonian structures in the Southwest Yermak Plateau (SYP; Ritzmann and Jokat, 2003). Subsequently, rifting relocation or abortion of the Sophia Basin to a wider zone was likely to occur, in this case, along the Lena Trough (Fig. 1) and during NE Atlantic-Arctic connection in mid-Cenozoic times (< 25–15 Ma; Engen et al., 2008). Therefore, the Yermak Plateau is instead associated with the final line of continental breakup in the northwestern sector of the North Svalbard margin (Figs. 13 and 15). The sag phase may have developed soon after strong Eurekan deformation ceased at ~35 Ma (Meier et al., 2024) (Fig. 15), and it indicates correspondence with models of oblique rifting and migration of deformation towards the latest stages of continental transform faulting with marginal plateau uplift (Gernigon et al., 2014; Balázs et al., 2023).

The north-northwest trend of the transfer zones and the rift basins are moderately oblique (~45–30°; e.g., sensu Brune, 2014) with respect to the north-south onshore fault zone structuring of Svalbard (Fig. 13). It has been suggested that the spatial distribution of crustal thinning, rift

basins, and margin segments width primarily depends on their obliquity to extension direction (e.g., Ghana margin, Antobreh et al., 2009; Demerara Plateau in Central Atlantic, Loparev et al., 2021; Nuussuaq in West Greenland, Guan et al., 2023). Moreover, during initial stretching, widespread rift basins develop in the necking domain as the ductile crust is thinned until whole-crust coupling occurs, and mantle inheritance (e.g., suture) can directly transfer deformation into the upper crust (Zwaan et al., 2024). In this sense, the orientation of the proposed possible transfer zones of this study in the North Svalbard margin (Fig. 14), may be oblique to the Caledonian basement fabric. This is because the transfer zones, or the Caledonian fabric, are aligned to the proposed initial Late Mesozoic-Paleogene rift direction of the Eurasian plate that resulted in the nearly orthogonal final breakup line (Gaina et al., 2009) (Fig. 15).

The outer necking and transitional domains of the Northeast Kvitøya-North Barents (NKB) segment consists of a wide zone with large-scale detachment surfaces (>1 s twt fault throws) and increased content of volcano-clastic features (e.g., wedge units; CT5 and CT6, Figs. 10 and 11). By comparison, the transitional domain located at the eastern part of the Northeast Spitsbergen-Northwest Kvitøya (NSK) composite segment (Fig. 14) is narrower and more oblique with respect to the proposed early extension direction (Figs. 13 and 15). If the initial Eurasian plate rift direction was maintained constant through time, the current width of both eastern parts of the NSK and that of the NKB segment could be reproduced (~50 to ~100 km, respectively; Fig. 14). As the NKB segment likely remained preferentially aligned with respect to the opening direction during mature rifting, prominent lower crust and mantle structures typical of intense exhumation and high degrees of coupling may have formed (Brune, 2014; Zwaan et al., 2024; e.g., the antiformal feature and continent-dipping detachments in CT5 and CT6; Figs. 10 and 11). Furthermore, the presence of wedge units north of the NKB segment suggests that the asymmetric spatio-temporal evolution of magma extraction along the Gakkel Ridge (Figs. 1 and 14) could be related to oblique opening mode of the Eurasia Basin (e.g., Brozena et al., 2003; Glebovsky et al., 2006) (Fig. 15). Therefore, the narrower eastern corridor of transitional crust along the Yermak Plateau and Sophia Basin would have synchronously developed as a pull-apart region across the proposed transfer boundary system juxtaposed against the southwestern Nansen Basin crust (Eocene; Figs. 14 and 15), rather than in a highly oblique (> 45°) Late Cretaceous-Paleocene rift setting without the onset of seafloor spreading (Berglar et al., 2016).

6. Summary and conclusions

We present results of an integrated workflow that consists of seismic and structural interpretations with modelling of potential field data to constrain the crustal structure, rift basin architecture, and continental breakup processes at the North Svalbard continental margin. New insights of the tectonic evolution related to the Paleogene continental rifting, the development of the northern Barents Sea passive margin, Yermak Plateau, Sophia Basin, and the opening of the adjacent Eurasia Basin, are constrained from seven crustal transect models (CT1–7).

We suggest that the southwestern Nansen Basin includes a narrow transitional domain (e.g., COT zone of ~50–80 km wide) with exhumed continental lower crust and mantle guided by detachment structures. The COB marks the inception of the oceanic crust, which is delimited landward by the oldest unambiguous seafloor spreading anomaly C24. Wedge geometries underneath the top acoustic basement seismic reflector are present along the COT and occasionally at the eastern COB regions and may locally indicate an increased magmatic budget before and during early seafloor spreading. Internal layering within the oceanic and transitional domains of the southwestern Nansen Basin can be ascribed to a serpentinization front.

Segmentation of the North Svalbard continental margin is inferred from the crustal and rift basin architecture, and includes basement domains across the Yermak Plateau and along the shelves north of

Spitsbergen to Nordaustlandet (Caledonian fabrics), and Kvitøya (Caledonian/Timanian fabrics). The Yermak Plateau is composed of two juxtaposed and re-worked continental blocks with different tectono-magmatic origins, namely the Northeastern Yermak Plateau (NYP) and the Southwest Yermak Plateau (SYP). The NYP is older, formed prior to or during an early Eureka setting (~60 Ma), whereas the SYP is younger, and related to shearing processes followed by oblique extension of the West Svalbard-NE Greenland conjugate margin until end of Oligocene times (~30 Ma). The North Svalbard shelf consists of three main segments: the (1) Northwest Spitsbergen-Southwest Yermak Plateau (NSY), (2) Northeast Spitsbergen-Northwest Kvitøya (NSK), and (3) Northeast Kvitøya-North Barents (NKB). These segments are separated from each other by north-northwest trending transfer zone systems. The transfer zones impose different distributions and orientations of basin-bounded faults in the necking and transitional domains, focus excess magmatism, and serve as sharp/abrupt juxtaposition boundaries across contrasting crustal/lithospheric domains. Apparent polarity shifts and relative displacement observed within different high-positive amplitudes in the potential field anomaly maps also coincide with the location of the transfer zones.

The Sophia Basin region is considered as a transitional domain with basement underlain by high-density crust (~3000 kg/m³). However, its structure and composition are partly different to the transitional domain regions of the North Svalbard margin facing into the Nansen Basin (i.e., NSK and NKB segments). Based on the presence of down-faulted pre-breakup units and severely sheared crustal structure from seismic reflection data, as well as correlations to the onshore geology mapped in northern Svalbard, we suggest that the Sophia Basin is composed of meta-sedimentary rocks within a deformed granitic upper crustal unit, whereas a combination of granulite-amphibolite facies and/or serpentinized sub-continental mantle within the modelled lower crustal unit section. This implies that the Sophia Basin is a marginal depocenter formed by multi-phase extension during Paleogene. A first main phase focused extension on the outer parts of the Sophia Basin and may relate to Paleocene rifting of the North Svalbard region. A second main phase is characterized by high subsidence rates with establishment of a broader sag area that developed as a collapsed intra-continental region towards the end of Eureka deformation in the Eocene-Oligocene transition (~33 Ma). The thin crust (< 5–7 km) and thick sedimentary pile (up to ~4 km) within the sag region of Sophia Basin formed before the rift axis was relocated to more favorable locations along the northern flanks of the Yermak Plateau, and during the separation of the Morris Jesup Rise from the North Svalbard continental margin also towards the end of Paleogene.

Pre-existing Caledonian (N-S) and Timanian (E-W) basement fabrics in Svalbard and northern Barents Sea are heavily modified across a tectonic weak zone affecting Kvitøya and become progressively more oblique. The interplay between these inherited basement structures and the multi-phase rifting setting in the area resulted in shear-extensional processes and selective areal development of associated sedimentary basins since late Paleozoic. During Cenozoic continental breakup, deformation in North Svalbard margin was focused approximately north of the region situated today at 80°N, with only minor crustal thinning beneath late Paleozoic salt basins in the south of Kvitøya. The north-northwest orientation of the transfer zones that separate the margin segments proposed here may be a consequence of the obliqueness between the N-S Caledonian fabrics and early extension direction of the Eurasian plate, probably exploited also by Eureka deformation during Paleogene. These interactions were key during basin formation of the North Svalbard margin necking domain, the opening and later deformation relocation from the Sophia Basin, and the final configuration of the transitional domain in the wider NKB and the narrower NSK segments. The role of the possible transfer zones during the initial tectono-magmatic evolution of the oceanic crust domain of the Eurasia Basin is yet to be fully determined, as some exhumed basement regions within SW Nansen Basin could be a record of early ridge-transform interactions.

In this sense, the traditional magma-poor rifted margin classification in the study area should be re-evaluated to more complex tectonic models, including the interactions with the Greenland plate.

CRediT authorship contribution statement

Juan Camilo Meza-Cala: Writing – original draft, Visualization, Formal analysis. **Jan Inge Faleide:** Writing – review & editing, Validation, Supervision, Investigation, Funding acquisition. **Alexander Minakov:** Writing – review & editing, Validation, Supervision, Data curation. **Mansour M. Abdelmalak:** Writing – review & editing, Visualization, Supervision, Methodology. **Grace E. Shephard:** Writing – review & editing, Supervision. **Wolfram H. Geissler:** Writing – review & editing, Validation, Investigation, Data curation. **Peter Klitzke:** Writing – review & editing, Investigation, Data curation. **Rune Matingsdal:** Validation, Data curation. **Carmen Gaina:** Writing – review & editing, Supervision, Project administration, Funding acquisition, Conceptualization.

Declaration of competing interest

The authors declare that they have no known competing financial interests or personal relationships that could have appeared to influence the work reported in this paper.

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Appendix A. Supplementary data

Supplementary data to this article can be found online at <https://doi.org/10.1016/j.tecto.2025.230861>.

Data availability

Seismic reflection profiles and potential-field datasets are publicly available or can be provided upon request by the designated authorities (see Methods section and the online data repositories at: <https://www.sodor.no/en/diskos/> and <https://www.marine-data.de/>). Supplementary figures file is attached in this work, and a table of published crustal structure information compiled in this study is archived in a Zenodo repository at: doi: <https://doi.org/10.5281/zenodo.15006024>.

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