Bathymetry, controlled source seismic and gravity observations of the Mendeleev ridge; implications for ridge structure, origin, and regional tectonics

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SUMMARY
Multichannel seismic (MCS), seismic refraction, and gravity data collected down the flank of the Chukchi Plateau, but predominantly over the Mendeleev Ridge have been processed and interpreted to describe the crustal style of the ridge, as well as the structural history. These results provide constraints on the origin of the ridge, and the tectonic evolution of the Amerasian Basin. MCS images reveal two primary sediment sequences separated by an unconformity that persists across the entire Mendeleev Ridge. The basement and lower sediment sequence exhibit pervasive normal faulting. The upper sequence is laterally conformable and not effected by faulting, thus the regional unconformity dividing the two sequences is interpreted to mark the end of extensional deformation. Modeling of sonobuoy seismic refraction data reveals upper crustal P-wave velocities ranging from 3.5 to 6.4 km s\(^{-1}\) approximately 5 km into the basement. The velocity structure of the Mendeleev Ridge is consistent with either a volcanic rifted continental margin, or an oceanic plateau origin. Observed gravity anomalies over the ridge are reproduced by a model consisting of bathymetry, sediment and basement horizons from the MCS data and a single crustal layer of 2.86 g cm\(^{-3}\). This result is consistent with homogeneous, mafic crust. The similar velocity and density structures of the Mendeleev and Alpha ridges is consistent with a model where the two ridges are contiguous and share a common geological origin. Gravity modelling over the transition between the Chukchi Plateau and the Mendeleev Ridge suggests the two features have differing compositions and distinct emplacement histories. Three tectonic models are presented for the origin of the Alpha Mendeleev Ridge (AMR) that satisfy constraints set by this and previous studies: (1) a rifted volcanic continental margin, (2) an oceanic plateau formed at a spreading centre—perpendicular to the AMR and (3) an oceanic plateau formed at a spreading centre—parallel to the AMR.

Key words: Gravity anomalies and Earth structure; Controlled source seismology; Oceanic plateaus and microcontinents; Submarine tectonics and volcanism; Large igneous provinces; Arctic region.

1 INTRODUCTION
The Arctic Ocean can be divided into two distinct regions: the Eurasian, and Amerasian basins (Fig. 1). The Eurasian Basin was formed by the propagation of the very slow spreading Gakkel Ridge (6–12.7 mm yr\(^{-1}\)), concurrent with the opening of the N. Atlantic Ocean at \(\sim 56\) Ma (Ostenso & Wold 1973; Coles et al. 1978; Vogt et al. 1979; Kristoffersen 1990; Cochrane et al. 2003). The Amerasian and Eurasian basins are separated by the Lomonosov Ridge,
Figure 1. Top panel: regional bathymetry map of the Arctic Ocean taken from the IBCAO grid with ship track (Jakobsson et al. 2000). Bottom inset: Mendeleev Ridge study area. HLY0503 ship track (black). Segmented purple lines are MCS lines. Red dots indicate sonobuoy drop locations and modelled sonobuoys (34, 35, 37, 44, 58) are identified. Yellow lines A–A’ and B–B’ represent projected gravity lines. Gravity models for these profiles are shown in Figs 12 and 13. Gravity data projection and mapping utilise Generic Mapping Tools (GMT; Wessel & Smith 1998).

A continental fragment rifted off the Barents shelf by spreading along the Gakkel Ridge. This hypothesis, first proposed by Heezen & Ewing (1961) became generally accepted through seismic reflection imaging of the crust (Jokat et al. 1992; Jokat 2005), and was confirmed by the recent ACEX drilling of the ridge (Moran et al. 2006).

In contrast to the Eurasian Basin, the geological history of the Amerasian Basin is less understood and presumed to be more
of anomalies is tentatively observed in the southern Canada Basin by Taylor et al. (1981) and Coles & Taylor (1990), and correlated with magnetic anomalies (M25–M12) to indicate initial orthogonal, followed by rotational seafloor spreading at 153–127 Ma (Fig. 2).

The diffuse Canada Basin magnetic anomalies are coincident with a north–south lineated gravity low which has been interpreted as a fossil spreading centre (Fig. 3; Laxon & McAdoo 1994). Seismic refraction and reflection experiments reveal horst and graben basement structures coincident with the potential field observations (Grantz et al. 1998). Acknowledging the diffuse fan-shaped anomalies of the Canada Basin, Grantz et al. (1998) describe cores containing contemporaneous Phanerozoic stratigraphy in the Northwind Ridge (Chukchi Borderland) and the Sverdrup Basin (Arctic Canada), providing further evidence to support the rotational model.

The rotational model has been tentatively accepted due to a variety of supporting geological and geophysical observations, but also due to the absence of substantial evidence supporting other models. Still, alternative models have been presented. Lane (1997) describes the space problem created by the presence of the Chukchi Borderlands, that is, restoring Alaska–Chukotka to the Canadian Arctic margin results in significant continental overlap. Miller et al. (2006) dated zircon suites from several terrigenous sandstones surrounding the Amerasian Basin testing for common geographic sources. The results suggest that Chukotka is not part of the Arctic–Alaska microplate but rather originated from the east, near Taimyr and Verkhoyansk. The authors go on to infer that the Chukchi Borderland, like the Alpha and Mendeleev Ridge is thinned continental crust that was transported to its present position from the direction of the Barents Shelf, not from the Sverdrup Basin as predicted by the rotational model. These results do not preclude the rotational model for the southern Canada Basin, but do argue against this explanation for the whole of the Amerasian Basin.

2.2 Alpha and Mendeleev ridges

Due to the absence of strong magnetic lineations in the Canada Basin the rotational model is speculative, but perhaps most inhibiting to a complete tectonic model of the Amerasian Basin is the unknown origin of the Alpha and Mendeleev Ridge (AMR). The ridges are bathymetrically contiguous, roughly symmetrical about their axes and together encompass an area of approximately 708 000 km² (Fig. 1; Jakobsson et al. 2003). The AMR stretches across the Amerasian Basin from the East Siberian Shelf to northern Canada, separating the Canada Basin from the Makarov Basin.

The Alpha Ridge and Mendeleev Ridge may have formed in a single, continuous event. The working hypothesis proposes that the ridges constitute a single oceanic plateau created by hotspot volcanism during the Late Cretaceous (Forsyth et al. 1986; Weber 1986; Asudeh et al. 1988; Lawver & Müller 1994; Lawver et al. 2002; Jokat 2003), although several of these authors note that they cannot rule out a continental origin for the ridges. Miller et al. (2006) and Lebedeva-Ivanova et al. (2006) prefer an interpretation of attenuated continental crust at least for the Mendeleev Ridge.

In order to develop an accurate model of the geological history of the Amerasian Basin, understanding how the Alpha and Mendeleev ridges relate to one another and to surrounding features is of utmost importance.

2.2.1 Mendeleev ridge

There have been few geological and geophysical projects dedicated to discerning the geological history of the Mendeleev Ridge. Hall
Figure 2. Regional magnetic anomaly map. Smoothed IBCAO bathymetry contours (500 m) overlay magnetic anomalies (Verhoef et al. 1996). Solid black line in the Canada Basin represents spreading axis tentatively identified by Taylor et al. (1981). Dashed black line in the Canada Basin delineates gravity low; See Fig. 3.

(1970) re-examined the seismic, gravity and magnetic data taken from T-3 with the assistance of newer bathymetry data (Beal 1968). Sediment thicknesses were found to vary between several hundred and 1000 m. Gravity modelling revealed that the crust was ~32 km thick with the upper 5 km as East Pacific Rise style crust and the lower 27 km representing anomalous crust with a bulk density of 3.15 g cm$^{-3}$. Hall (1970) preferred spreading centre hypothesis for the AMR, mapping multiple fracture zones along an axial rift valley on both the Alpha and Mendeleev ridges. Several spreading events were invoked: Palaeozoic, Late Mesozoic and Early Tertiary. However, the refutation of the normal spreading centre hypothesis was published soon after Hall’s work and shows the sediment-corrected depth of the AMR is too shallow to have formed at a normal Cretaceous spreading centre (Delaurier 1978).

The Russian Ice station NP-26 crossed the Mendeleev Ridge several times between 1983 and 1986, though seismic reflection data were published only recently along with results from the Arctic 2000 investigation of the ridge (Lebedeva-Ivanova et al. 2006). These seismic images reveal a rugged basement with sediment thicknesses ranging from 0.1 to ~2.3 km on the western slopes off the ridge bordering the Makarov Basin. During the Arctic 2000 survey, seismic refraction, shallow reflection and gravity data were collected along a 500 km, EW oriented profile crossing the northern Mendeleev Ridge at 82°N. The refraction Moho was imaged at 32 km beneath the ridge and 13 km at the ridge flanks. The authors presented a velocity–depth structure and gravity anomalies, from which they prefer the interpretation that the Mendeleev Ridge is ‘highly attenuated continental crust’.

The wide-angle refraction data over the Mendeleev Ridge presented by Lebedeva-Ivanova et al. (2006), along with the MCS reflection, refraction, coregistered gravity and multibeam bathymetry data presented in this study, provide new insights into the structure of the Mendeleev Ridge and how it relates to the Alpha Ridge and neighbouring features.

2.2.2 Alpha ridge

Until recently, the Alpha Ridge had been more thoroughly studied than the Mendeleev Ridge. Hunkins (1961) first interpreted seismic reflection and refraction data from the T-3 ice island and favoured a fault block origin for the AMR.

The Canadian expedition to study Alpha Ridge (CESAR) was the first dedicated attempt to discern the geological history of the Alpha Ridge. Seismic refraction experiments revealed a very thick crust of 38 km below the ridge axis (Forsyth et al. 1986). Together with the refraction results, higher densities determined from gravity modelling over the Alpha Ridge than the Lomonosov Ridge, and the homogenous nature of the magnetic anomalies over the
ridge, they hypothesize that the Alpha Ridge ‘may be composed of a large pile of mafic rock, possible unique to this planet (Weber 1986)’. Furthermore, Forsyth et al. (1986), Asudeh et al. (1988) and Weber (1990) cite the thickened crust, the absence of any significant reflector between the top of the basement and the Moho, and the similarities in the velocity gradients of the Alpha Ridge and Iceland, as evidence that the Alpha Ridge is an Oceanic Plateau, possibly a hot spot similar to Iceland.

Sediment cores and dredged basement rocks recovered during the CESAR expedition provided the first direct geological evidence into the origin and formation age of the Alpha Ridge. Mudie & Blasco (1985) found certain fossil assemblages within the cores dating to the Campanian and Maastrichtian, providing an upper limit on the age of the ridge. And vesicular alkali basalts, dredged from graben flanks, are described by Van Wagoner et al. (1986) to be indicative of intraplate volcanism.

Based on plate reconstructions and estimated ages, Lawver & Müller (1994) propose that both the Alpha and Mendeleev ridges were likely produced by the drift of the Amerasian Basin over the Iceland plume.

More recently, Jokat (2003) observed upper basement velocities from the Alpha Ridge that are similar to the T-3 results presented by Hunkins (1961), and consistent with an oceanic plateau origin.

MCS data reveals large scale normal faulting over the Alpha ridge flanks and a regional unconformity possibly related to the opening of the Eurasian Basin. Alkali basalts were also dredged during this survey, and Mühle & Jokat (1999) report a whole rock 40Ar/39Ar plateau age of 82 ± 1 Ma, suggesting a Late Cretaceous age of formation.

2.2.3 Potential field observations of the alpha and Mendeleev ridge

Weber (1986) showed that magnetic anomalies over the Alpha Ridge correlate predominantly with bathymetry and basement topography. Regional gravity models show that the gravity field also correlates with bathymetry, with no observed lateral variation in the crust (Sobzak et al. 1990). Results from these magnetic and gravity studies suggest a homogenous crustal structure, characteristic of oceanic plateaus but not exemplary of the variation expected in continental crust.

Interpreted age of formation from regional magnetic anomalies appears consistent with those derived from geochemical and fossil analysis over both the Alpha and Mendeleev ridges. Weber & Sweeney (1990) suggest ridges formed during a period
of normal polarity, and entirely during the Cretaceous quiet period (120–84 Ma).

Regional gravity (Glebovsky et al. 2000; Kenyon & Forsberg 2001) and magnetic (Verhoef et al. 1996) maps both suggest by the lateral consistency of amplitudes, or anomaly pattern, or both, that the Alpha and Mendeleev ridges are a singular geological province (Figs 2 and 3). Analysis of regional gravity and magnetic anomaly maps also suggests that the crustal material of the AMR extends into the Makarov and Canada basins, beyond its bathymetric expression.

Vogt et al. (2006) synthesize multiple geophysical data sets from the AMR. They also find that magnetic anomalies are generally correlated with bathymetry and ±20 mGal free-air gravity anomalies. In comparison to global oceanic plateaus, the estimated magmatic volume of 10,000 km³ calculated for the AMR is surpassed only by the Ontong Java Plateau. The authors hypothesize that emplacement of the AMR is related to volcanic events in the Sverdrup Basin but do not rule out a continental component in the ridge crust.

2.3 Chukchi Borderland

The Chukchi Borderland is separated from the Mendeleev Ridge by a bathymetric low (~2500 m depth) which opens up and becomes more planar to the south of the study area (Fig. 1). In the past this region has been called the Chukchi Abyssal Plain (Grantz et al. 1998) and the Mendeleev Basin (Lebedeva-Ivanova et al. 2006). Due to the narrow extent of the basin and to avoid ambiguity, this feature is here referred to as the Chukchi–Mendeleev Borderland. In this paper we examine the relationship between the Chukchi Borderland and the Mendeleev Ridge, which has not been directly studied in the past, and is not well understood.

The Chukchi Borderland has long been considered a thinned continental fragment (Fig. 1; Hunkins 1966; Hall 1990; Laxon & McArdoo 1994; Klemperer et al. 2002; Wolf et al. 2002). Consistent with the rotational model for the Canada Basin, Grantz et al. (1998) suggest the Chukchi Borderland rifted off the Canadian Arctic shelf, together with Arctic Alaska. Piston cores collected on the slope of the Northwind Ridge, sampled Palaeozoic through Late Jurassic strata, which were found contemporaneous with sections from the Sverdrup Basin.

To account for the present location of the Chukchi Borderland, which creates a space problem for simple rotation, Grantz & May (1982) suggests later, clockwise rotation of the Chukchi microplate away from the E. Siberian shelf, possibly contemporaneous with spreading in the Canada Basin and prior to the emplacement of the AMR.

2.4 Makarov Basin

The Makarov Basin lies between the AMR and the Lomonosov ridge (Fig. 1). The basin itself is subdivided by a basement ridge paradoxically named the Arlis Gap (Fig. 1). The basement ridge, as revealed by gravity (+10–15 mGal) and magnetic (800 nT) anomalies, acts as a dam to sediments; thus sediments are ~3.5-km-thick south of the ridge and ~1.5 km thick north of the ridge (Kutschale 1966; Sorokin et al. 1999). The region south of the Arlis Gap is the Podvodnikov Basin (Wrangel Abyssal Plain) and the area north of the ridge has been called the Fletcher Abyssal Plain though is now simply called the Makarov Basin (Jakobsson et al. 2003).

The origin of the Makarov Basin is enigmatic. Taylor et al. (1981) tentatively identify lineated magnetic anomalies 21–34 (49–84 Ma) and propose a seafloor spreading origin. Gravity modelling found the Makarov Basin crust to be 23 km thick (Ostenso 1964). Jackson et al. (1986) hypothesize that similar seismic velocity gradients indicate that the crust underlying the Makarov Basin is structurally contiguous with the AMR. A more recent seismic expedition visited the Makarov Basin where velocity gradients (upper basement = 5.0–5.2 km s⁻¹; lower crust = 6.7 km s⁻¹) revealed both oceanic layer 2 and 3, respectively (Sorokin et al. 1999). Also observing a total crustal thickness of 23 km, the authors prefer the interpretation of thickened oceanic crust but do not rule out a continental origin.

2.5 Circum-arctic cretaceous volcanism

Cretaceous volcanic rocks are observed in multiple locations around the circum-Arctic and may constrain the timing of extensional events in the Amerasian Basin as well as the formation and evolution of the AMR. Tholeiitic and alkali basalts are found in the Canadian Arctic and northern Greenland (Embry & Osadetz 1988; Embry & Dixon 1990; Weaver et al. 2006; Villeneuve & Williamson 2006), as well as in Franz Joseph Land and Svalbard (Maier 2001; Estrada & Kunst 2004). From the Delong Archipelago in the East Siberian Shelf, Fujita & Cook (1990), Drachev et al. (1999) and Drachev & Saunders (2006) observe picritic basalts and continental tholeites. As more Cretaceous volcanic events are described and dated, it appears that these events are related, temporally, spatially and compositionally. Some have suggested that individual occurrences of Cretaceous volcanism may be related to the formation of the AMR (Drachev et al. 1999; Estrada & Kunst 2004), and that these circum-Arctic volcanics constitute a Large Igneous Province (LIP; Tarduno et al. 1998; Maher 2001; Drachev & Saunders 2006). Models describing the development of the AMR, and the tectonic history of the Arctic Basin must explain the potential relationship with these volcanic provinces.

3 EXPERIMENT OVERVIEW, METHODOLOGY AND RESULTS

3.1 Seismic reflection

3.1.1 Experiment overview and methodology

The seismic experiment was designed and primarily implemented by a group from the University of Bergen. In 2005 mid-August approximately 730 km of MCS reflection data were recovered down the western flank of the Chukchi Plateau, across the Chukchi–Mendeleev Basin and over the Mendeleev ridge along with coregistered swath bathymetry, gravity data and seismic refraction data (Fig. 1).

The seismic source was two 250 cu in G-guns. The streamer length was limited by ice conditions to ≤300 m. Wear and tear from towing the analogue streamer through ice degraded the hydrophones and the number of active channels ranged from 24 to as few as 11. The signal was digitized on board using two Geometrics Geode seismographs. Shot spacing was 20 s (~40 m) with a 2 ms sampling rate. Shot depth was approximately 5 m and varied with ship speed and ice conditions. Hydrophone spacing was 12.5 m and stacked data are grouped into 6.25 m common depth point (CDP) bins with an average fold of 4. We assumed straightline streamer geometry for the purposes of normal moveout (NMO) correction and stacking.

The MCS data required significant manual trace editing and automated noise attenuation to eliminate random electrical noise.
Frequency–wavenumber (FK) filtering was used to eliminate low velocity noise caused by the streamer travelling through ice. After the trace editing and filtering, individual traces were bandpass filtered at 6–100 Hz. The NMO correction was made assuming constant velocity layers for the water column, sediments and basement. While imprecise, these velocities are largely consistent with results from the seismic refraction modelling. Initial attempts to derive more accurate layer velocities through velocity analysis were not successful as the data were not adequately dense to provide sufficient signal coherency. The data were then stacked and migrated with Stolt's FK, constant velocity (1490 km s$^{-1}$) algorithm (Yilmaz 2001).

SIOSEIS (Henkart 1981) as well as Promax were used for data processing in order to utilize the most effective functions from each program. Sediment unconformity and basement horizons were identified and picked digitally for use in both the seismic refraction and gravity modelling.

3.1.2 Results

The MCS survey over the Mendeleev Ridge was broken into seismic lines 17, 18, 20–25 (Fig. 1). Line breaks occur where equipment was recovered due to ice conditions or maintenance. These shorter lines can be aggregated to make two continuous lines, A–A′ (17, 18, 20a) and B–B′ (20b–23). These two lines correspond to the two gravity profiles, and provide a natural segmentation in which to interpret the structures of the ridge (Fig. 1). The MCS images are shown in Figs 4–8.

Reflection images reveal two distinct sedimentary units. The upper sediment ‘Unit I’ (thickness: 0.0–0.7 s TWT, or approximately 0–560 m) is laterally continuous and drapes most of the Mendeleev Ridge with average thickness of $\sim0.3$ secs TWT ($\sim250$ m). Due to the multiple pinchouts of the lower layer of Unit I (Ib) (CDP 21 200–21 800), there are likely at least two distinct sediment horizons within Unit I (Fig. 4). This unconformity may represent a hiatus between two successive passive pelagic units. As there is no significant deformation of Ib, all of Unit I was likely deposited subsequent to the most recent tectonism. A large subsurface channel appears at the base of the Chukchi Plateau (CDP 9600–11 000). Within the channel unit I has been scoured out and subsequently refilled with chaotic deposits approximately 8 km wide and 450 m deep (Fig. 4).

Separating Unit I from lower sediments is an easily identifiable acoustic boundary that is traceable across the Mendeleev Ridge (Figs 4–8). This boundary is interpreted here as a regional unconformity.

The lower, higher reflectivity Unit II (0.0–1.0 s TWT or approximately 0–900 m) is highly deformed in places. The thickness of Unit II is significantly greater in grabens, and is more variable than that of Unit I. The greatest deformation occurs where the basement is significantly normal faulted. In Fig. 6 (CDP 8200–9300), sediments fan towards half-graben walls suggesting deposition and deformation occurred simultaneously. Unit II was likely deposited prior to and contemporaneous with tectonic deformation.

Unit II is mostly conformable over areas not affected by faulting, and has a reflection character similar to Unit I. Though along steep slopes, specifically on the flanks of the Chukchi Plateau, much of the unit appears to be a product of mass wasting (CDP 5000–8300) (Fig. 4). Unit II is also affected by the large channel scours of Unit I.

In Fig. 5 (CDP 8000) there is a low relief fault that could be strike-slip. It incises both the basement and Unit II exhibiting a small flower structure. Small, high slope faults within some grabens also suggest strike slip motion. In Fig. 7 (CDP 12 000–13 200), younger sediments within Unit I onlap the basement as well as older sediments conformably overlying the basement, possibly indicating rift-flank uplift. The sediment sections are described in further detail within Bruvoll et al. (2010).

The basement of the Mendeleev Ridge and the transition to the Chukchi Plateau exhibits very high relief due to extensional faulting. Rounded basement topography in some locations (e.g. Fig. 4, CDP 7600–8000; Fig. 6, CDP 12 500–12 800, 17 000–20 200) may suggest volcanism but is more likely the result of oblique ship crossings. Mostly the basement appears homogenous, though in areas we observe discrete horizontal to subhorizontal coherent reflections. Along the flank of the Chukchi Plateau there is a visible graben-like structure within the basement, that may represent volcanic or sediment infilling of a pre-existing structure (CDP 8800–10 500) (Fig. 4). Subbasement reflectors observed farther down the line in Fig. 4 (CDP 17 200–18 500) and in Fig. 5 also cannot be conclusively interpreted as they may represent lithified stratigraphy, volcanic flows, or ‘ringing’ of the strong basement reflector. In Fig. 6 (CDP 7600–9200, TWT 2.4–3.1 s), there is a strong subbasement reflector tilted and offset within several half-graben structures. And in Fig. 7 multiple reflections within the basement (CDP 12 100–13 600) are tilted away from a large graben. The inclined dip of these reflectors is in contrast to the—horizontal regional trend, further suggesting rift-flank uplift.

The reflection multiple of the basement reveals many deeper reflecting layers not visible above the multiple, but these reflectors cannot be confidently interpreted. They are likely multiples of multiples or peg-leg multiples, but may represent lithified stratigraphy or volcanic flows.

3.2 Seismic refraction

3.2.1 Experiment and methodology overview

Seismic refraction was collected along MCS profiles using the same gun source and one-component marine sonobuoys (Sparton Model AN/SSQ-57SPC) as receivers (Fig. 1). When the seismic reflection experiment was active, and where ice conditions permitted, one active sonobuoy was deployed in the water. In total, 24 sonobuoys were deployed over the Mendeleev Ridge. Five of these contained traceable refractions. Total sonobuoy offset was limited by line of sight and FM carrier loss. Refraction data were bandpass filtered to 6–60 Hz. We applied predictive deconvolution to sharpen the pulse, aiding identification of phase arrivals.

The objective in refraction modelling is to produce a velocity–depth model that accurately predicts the observed travel-times as picked from the refraction data. A 2-D ray tracing algorithm was used to calculate layer velocities with RAYGUI (Zelt & Smith 1992; Song & ten Brink 2004). Results presented here are based only on forward modelling through velocity–depth models. Model shot-offset was calculated from the direct arrivals through water assuming a water velocity of 1450 m s$^{-1}$ and straightline geometry. As an assessment of model fit, whole model root mean square (rms) differences between observed and calculated arrivals were calculated. Acceptable models reproduced the picked arrivals to within 0.1 s.

Including reliable a priori information, for example, MCS horizons as well as amplitude, geological and tectonic information, yields more accurate modelling results (Zelt 1999). Initial
velocity–depth models shown here were forced to honour the subsurface geometry as picked from the MCS data and include bathymetry, unconformity and basement horizons. Originally only one sediment layer was included in the velocity–depth models. However, traveltime misfits in the refraction models and the clear presence of a regional unconformity in the MCS data led to the inclusion of the unconformity in the velocity–depth models. Zero-offset phase arrivals from the sonobuoy records were all compared with equivalent boundaries on MCS records to ensure model accuracy. This time comparison unequivocally revealed an excellent fit between the two data types.

For each model, velocity boundary depths were converted from traveltime with velocities assigned to the top and bottom of each horizon, increasing linearly with depth (Figs 9 and 10). To achieve better fit, layer velocities were adjusted with each forward run in an attempt to satisfy both reflection and refraction arrivals. This trial and error method was repeated until the rms error $\leq 0.1$ s. A successful model requires that assigned velocities must produce traveltime curves that match the observed zero-offset arrivals within the $0.1$ s margin of error as well as reproduce the gradient of the observed phase arrivals.

The goal of the seismic refraction analysis is to provide physical constraints on the sediment and crustal material of the Mendeleev Ridge, as well constrain the gravity models. During this experiment we could not image the crust–mantle boundary (32 km depth), so the ultimate objective of this modelling was to describe the velocity structure of the upper crust, and to compare these results with those from nearby Arctic features as well as analogous structures globally.

3.2.2 Results

Results are based on the modelling of five sonobuoy records over the flank of the Chukchi Plateau and along the Mendeleev Ridge (Fig. 1). The average direct arrival offset achieved was $\sim 25$ km. Most refracted arrivals are traceable to offsets of 10–15 km with only

Figure 9. Seismic refraction modelling; Sonobuoy 34. (a) Record section of sonobuoy data. Arrival picks used for ray tracing displayed in blue. (b) Velocity model. Solid lines are ray paths and dashed lines are layer boundaries taken from MCS data. Synthetic waves are reflected off the seafloor, regional unconformity and basement. Synthetic refracted waves are modelled through the water column for direct arrivals and through the basement to resolve upper basement velocities. (c) Calculated traveltime curves (solid lines) are plotted with observed picks (hachures) to determine model fitness. Sonobuoy location shown in Figs 1 and 4.
one sonobuoy showing refractions to >20 km offset. This limitation was the result of utilizing lower volume guns optimized for the MCS experiment and as a result, observed refracted waves sampled ≤6 km of crust (Figs 9 and 10). The velocity structure of each sonobuoy record is presented in Fig. 11.

Sonobuoy 34 (Figs 4 and 9): $\text{rms} = 0.064 \, \text{s}$. Refracted arrivals are visible to an offset of ~8 km and penetrate into the basement (turning depth) ~0.8 km. The data were collected coming down the flank of the Chukchi Plateau and is a region of high bathymetric variation (Fig. 1). The partial misfit between observed and calculated seafloor arrivals is likely due to the influence of local bathymetry.

Sonobuoy 35 (Fig. 4): $\text{rms} = 0.057 \, \text{s}$. Refracted arrivals are visible to an offset of ~14 km and penetrate into the basement ~2.8 km. The data were collected over the Chukchi–Mendeleev Basin. Basement velocities are lower than for sonobuoy 34 and may represent a change in basement material or may simply indicate more intense fracturing of the rock. There is a significant misfit for the early basement refractions. A large graben like subbasement structure can be seen on MCS line 17 through which these refractions travel (Fig. 1). It is likely that the observed refractions are influenced by this lower structure. The uppermost basement is responsible

Figure 10. Seismic refraction modelling; Sonobuoy 58. (a) Note that offsets >15 km are not shown in this record section. (b) Unlike the previous velocity models, an extra velocity boundary in the basement was required to reconcile later arrivals (c). Sonobuoy location shown in Figs 1 and 8.

Figure 11. Compiled velocity structure stack plots. Velocity structure of the upper crust over the Mendeleev Ridge from the 5 modelled sonobuoys (‘sono’). Water: 1.45–1.49 km s$^{-1}$. Sediment Velocities: 1.5–2.3 km s$^{-1}$. All velocities >3.4 km s$^{-1}$ represent basement material. See Fig. 1 for sonobuoy locations.
for basement reflections as can be seen by the good calculated versus observed fit. However, a model could not be constructed which included the reflecting uppermost basement, and maintained the refracting character of the subbasement structure. Despite the refraction misfit, the whole model misfit, rms = 0.057, is still well within the 0.1 s criteria.

Sonobuoy 37 (Fig. 4): rms = 0.030 s. Refracted arrivals are visible to an offset of ∼10 km and penetrate into the basement ∼1.6 km. The data were collected over the Mendeleev Ridge. Note that this velocity–depth model contains only one sediment horizon.

Sonobuoy 44 (Fig. 6): rms = 0.055 s. Refracted arrivals are visible to an offset of ∼11 km and penetrate into the basement ∼2.2 km. The data were collected coming down off the crest of the Mendeleev Ridge, stepping down half-graben structures.

Sonobuoy 58 (Figs 8 and 10): rms = 0.075 s. Refracted arrivals are visible to an offset of ∼25 km and penetrate into the basement ∼5 km. The data were collected farther north over the Mendeleev Ridge. To reconcile deeper travelling waves, another velocity layer is required with a smaller velocity gradient (6.2–6.4 km s⁻¹ at 10 km depth), and assigning relief to this deeper boundary reduces misfit for the later arrivals.

### 3.3 Gravity

#### 3.3.1 Experiment and methodology overview

Marine gravity data were collected during the transarctic crossing using a Bell BGM-3 gravimeter (Bell & Watts 1986). While underway the gravity data were corrected for Eotvos effects and geographic position. A 2-min Gaussian filter was used to eliminate transients due to ship heave, smoothing the output gravity signal, which was then decimated to 1-min samples. A post cruise gravity tie permitted drift correction. Drift was low, ~0.01 mGal d⁻¹ during the entire Healy deployment.

Forward modelling of the gravity data was accomplished using GM-Sys software (Wen & Bevis 1987). Initial densities of sediments were derived through empirical relationships regarding compressional wave velocities, rock type and depth (Ludwig et al. 1970; Christensen & Mooney 1995; Brocher 2005). Gravity data are segmented into two continuous profiles, each incorporating several seismic lines: Gravity Line A–A’ contains seismic lines 17, 18, 20a; Line B–B’ contains seismic lines 20b–23. For gravity profile locations, see Fig. 1. Gravity line A–A’ begins on the flank of Chukchi Plateau and ends near the axis of the Mendeleev Ridge, whereas gravity line B–B’ begins at the axis of the ridge and ends at the margin of the Canada Basin. Both profiles are ∼225 km in length.

Gravity and coregistered bathymetry data, along with regional unconformity and basement horizons picked off seismic records were projected to model profiles, perpendicular to the local ridge axis. Projecting data in this way maximizes variations of mass over the shortest lateral distance so as not to miscalculate anomalies using the technique developed by Talwani & Ewing (1960), which presumes infinite strike length. Gravity data then were resampled to 0.5 km intervals in model space.

While the short wavelengths are dominated by bathymetry, the longer wavelength gravity field is controlled by the deeper crustal structure and the geometry of the Moho boundary. A gravity admittance study by Williams & Coakley (2005) showed that the state of isostatic compensation on the AMR can be approximated by Airy isostasy. As a starting model, the Moho boundary was located assuming Airy compensation of the ridge. The maximum Moho depth was pinned at 32 km, directly beneath the highest bathymetric position on each ridge profile (Hall 1970; Lebedeva-Ivanova et al. 2006). The shape of the calculated Moho depends on crustal density, and thus each gravity model includes a Moho boundary specific to that assigned density. After initial validity tests, final models include bulk crustal densities between 2.76 and 2.9 g cm⁻³.

#### 3.3.2 Results

Line A–A’ (Fig. 12): Observed free-air gravity anomalies range from ∼32 to 44 mGal. Sediment densities converted from the refraction experiment were adequate to reproduce the short wavelength gravity field. In both profiles A–A’ and B–B’, sediment densities for sediment horizons I and II were 1.97 and 2.05 g cm⁻³, respectively. A single crustal density, along with the calculated Moho boundary was sufficient to reproduce the long wavelength field. The best-fitting crustal density was also consistent for both profiles at 2.86 g cm⁻³. Mantle density was kept uniform for both profiles at 3.3 g cm⁻³.

Invoking the calculated Moho, a maximum thickness of 32 km is observed beneath the crest of the Mendeleev Ridge and thins both to the east towards the Makarov Basin, and to the west towards the transition with the Chukchi Plateau. At this transition, the crust thins to 20 km and thickens to 32 km beneath the Chukchi Plateau. Ultimately, no lateral heterogeneity of the crust was required to reproduce observed anomalies with the exception being the transition between the Chukchi Plateau and Mendeleev Ridge.

Applying the predicted crustal thickness of 32 km for the Chukchi Plateau yields a distinct gravity deficit over the Chukchi Plateau, with respect to the Mendeleev Ridge.

Previous studies by Hunkins (1966), Hall (1970), Weber & Sweeney (1990) and Grantz et al. (1998) suggest the Chukchi Plateau is a region of thinned continental crust. Hunkins (1966) and Wolf et al. (2002), estimate crustal thickness of the outer Chukchi Shelf at ∼32 km depth. The Chukchi Plateau however is likely more extended than the shelf. Weber (1986) recalculated crustal structures from the Chukchi Plateau and Canada Basin from gravity profiles first presented by Hall (1970) in order to reconcile newer results from CESAR. Chukchi crust was estimated to be ∼22 km thick, with upper basement density of 2.75 g cm⁻³ and crust–mantle density contrast of 0.25 g cm⁻³.

While acknowledging the inherent non-uniqueness of gravity modelling, honouring the previous studies over the Chukchi Plateau shows that the most appropriate correction to the crustal model involves both thinning the Chukchi crust and reducing crustal density. To account for this gravity deficit for the Chukchi Plateau the crust is thinned to ∼29 km and the crustal density set to 2.85 g cm⁻³, while the Mendeleev Ridge crust is kept at 2.86 g cm⁻³. (Fig. 12c). This reduces the misfit to a large degree, and the adjustment is not so extreme as to disrupt the long wavelength fit for the whole of the profile.

Another local misfit occurs near to the crest of the Mendeleev Ridge. The high slope of the observed anomaly suggests the presence of a small, higher density body at shallow depths beneath the crest of the ridge. The misfit is small however, and is not considered to represent a significant discrepancy in crustal material.

Line B–B’ (Fig. 13): Observed free-air gravity anomalies range from ∼33 to 27 mGal along this profile. Gravity modelling along the more northerly flank of the Mendeleev Ridge reveals much the same result with the ridge axis pinned to 32 km. The ridge thins significantly towards the margin of the Canada Basin where the crust...
Figure 12. Gravity profile A–A’. Gravity Modelling Results. See Fig. 1 for profile location. Calculated and observed anomalies, along with calculated model error are plotted in the upper panels of (a), (b) and (c), while crustal–density structures are displayed in the lower panels. (a) Upper crustal section shows two sediment layers with no basement structure or Moho assigned. Note the calculated gravity surplus over the MR that indicates crustal thickening of the MR is required. (b) Full crustal structure with the best-fitting density for the MR crust as 2.86 g cm\(^{-3}\). This density assignment leaves a gravity deficit over the flank of the Chukchi Plateau and in (c), where thinning and reducing the density of the Chukchi Plateau crust reduces model error. A geological boundary between the Chukchi Plateau and MR is inferred.
Figure 13. Gravity profile B–B’; gravity modelling results. See Fig. 1 for location. (a) Upper crustal section shows two sediment layers with no basement structure or Moho assigned. Note the calculated gravity surplus over the MR that indicates crust thickens towards the crust of the MR. (b) Full crustal structure with the best-fitting density for the MR crust as 2.86 g cm$^{-3}$. (c) Underplated layer of 3.04 g cm$^{-3}$ at depths >26 km partially reduces model misfit.
is $\sim 20$ km thick. Again sediment densities are $1.97$ and $2.05$ g cm$^{-3}$ for sediment horizons I and II, and the best-fitting crustal density is $2.86$ g cm$^{-3}$ when only one crustal layer is invoked (Fig. 13b).

Model misfits occur at the edges of the model where more mass is not accounted for in the calculated anomaly. On the Canada Basin side of the profile, this misfit may be due in part to insufficient picks of the MCS identified basement, which results in $30$ km gap in the Moho boundary. In the model, this gap was linearly interpolated. There is some indication that at basement lows, the calculated anomaly is excessively low, suggesting that perhaps greater extension and thinning occurred in these localities than that given by the shape of the calculated Moho boundary. Another likelihood is that the sediments deep within the grabens have greater densities than $2.05$ g cm$^{-3}$.

In an attempt to correct for these edge misfits, an underplated, high-density ($3.04$ g cm$^{-3}$) layer was invoked where the Mendeleev crust was $\geq 26$ km (Fig. 13c). This addition reduces the model error. A lower crust layer was also invoked in gravity models over the Alpha Ridge, where the upper and lower crustal densities are $2.88$ and $3.04$ g cm$^{-3}$, respectively (Weber 1986). And Lebedeva-Ivanova et al. (2006) describe a high velocity, $7.4$–$7.8$ km s$^{-1}$, layer at the base of the Mendeleev Ridge that might signify underplating. Underplated layer was not invoked for gravity line A–A, as no such layer was necessary to reproduce the observed gravity anomalies.

4 DISCUSSION

4.1 Physical constraints on the geology of the Mendeleev Ridge crust

4.1.1 Upper crust

As observed on the MCS data, basement reflection character is predominantly homogenous and interpretations of subsedimentary coherent reflections are ambiguous, possibly representative of volcanic flows, lithified Mesozoic or older sediments, or sediments intercalated with basalt flows as commonly observed in oceanic plateau environments (Berger et al. 1992; Konnecke et al. 1998). Interpretation is inhibited by depth of penetration and the seafloor multiple. The multiples themselves reveal coherent reflectors in the crust not observed above the multiple, but most of this energy is likely from peg-leg multiples and thus are not interpreted (Fig. 5).

Compressional wave velocities for the upper crust are relatively uniform across the study area with the possible exception of sonobuoy 34 over the flank of the Chukchi Plateau, which appears to sample higher velocity material (Fig. 11). These upper crustal velocities are not lithologically diagnostic due to variations in porosity at shallow crustal depths. Basement velocities reported here may represent high velocity sediments (carbonates), or oceanic layer 2. Oceanic layer 2 is composed of the uppermost extrusive volcanic zone and may include sills, dykes, pillow, basalts and intercalated sediments. Lower seismic velocities are recorded in this youngest volcanic layer due to the abundance of fractures, faults and pore space, but increase rapidly with depth (Ewing & Houtz 1979; White et al. 1992).

The upper crustal velocities observed here ($3.5$–$6.4$ km s$^{-1}$) are slightly lower than those of the northern Mendeleev Ridge ($\sim 5.0$–$6.3$ km s$^{-1}$) and Alpha Ridge ($\sim 4.2$–$6.0$ km s$^{-1}$), though they are similar (Fig. 14). This may indicate that a different crustal material was sampled, or it may simply reflect the higher resolution of this survey, which does not penetrate to great depths, but is sensitive to the uppermost basement.

In a seismic experiment of a scale similar to our study, upper-basement velocities of $4.2$–$4.6$ km s$^{-1}$ were sampled over the Alpha Ridge by Jokat (2003). Hunkins (1961) presented velocities over the Alpha Ridge of $4.7$ km s$^{-1}$ for the uppermost $2.8$ km of basement.

While absolute values of upper basement velocities are not diagnostic of upper crustal material, velocity–depth gradients are more predictable for particular lithologies. Figure 15 shows the velocity–depth functions from this study superimposed on the results of White et al. (1992), who compiled velocity–depth functions of normal oceanic crust from the Pacific and Atlantic Oceans. Conclusions from this test are tentative though as the Mendeleev Ridge is unlikely to be composed of normal oceanic crust. It is approximately four times as thick, and as our interpretation of the MCS data shows, it has undergone significant post-formational extension. Upper basement velocities from the Mendeleev Ridge are mostly inconsistent with normal oceanic crust in that they are typically lower than those of oceanic crust of an equivalent depth. The best fit

\begin{figure}[h]
\centering
\includegraphics[width=\textwidth]{figure14.png}
\end{figure}
Figure 15. Velocity–depth functions from this study (red) plotted with stacked velocity–depth functions for normal oceanic crust (black) from the (a) Pacific and (b) Atlantic Oceans. Augmented from White et al. (2002).

is achieved with Atlantic Ocean results, 59–127 Ma, the most appropriate age for comparison. Sonobuoys 34, 37 and 58 plot within the envelope for normal oceanic crust (35 is considered an outlier due to the structural irregularities of the basement). These have the potential to represent normal oceanic crust, but the general comparison suggests that the velocities from the Mendeleev Ridge are incompatible with normal oceanic crust.

The basement velocity functions are also compared to those observed in Iceland, (Flovenz & Gunnarsson 1991), as well as Hatton Bank and the Voring Margin, both rifted volcanic margins within the North Atlantic Volcanic Province (NAVP; Fig. 16a). Eldholm & Grue (1994) compile results from these provinces to contrast the velocity–depth functions for the extrusive upper basement. Fig. 21(b) reveals a striking similarity between the velocity structure of oceanic plateaus and volcanic margins, possibly indicating similar compositions and processes governing magmatic emplacement. The velocity gradients presented in this study for the AMR are found compatible with the shallow results from both Iceland, and the volcanic margins. However, considering that velocities presented here only sample the upper crust, and that Iceland and the NAVP are younger features, implications from this comparison are also tentative.

4.1.2 Deeper crust—evidence of homogenous crust and contiguity with the alpha ridge

Gravity data suggest that the Mendeleev Ridge crust is laterally homogeneous as only a single crustal density layer of 2.86 g cm$^{-3}$ is required to sufficiently reproduce the gravity field for both $\sim 225$ km profiles. While in reality, crustal densities would increase gradually with depth, such gradation is not introduced here due to the success of invoking a single density layer. These results are compatible with a mafic, largely magmatic crust and incompatible with the typical heterogeneity of continental crust.

Our results are fully consistent with gravity studies over the Alpha Ridge (Weber 1986; Sobzak et al. 1990; Weber 1990; Weber & Sweeney 1990). Weber (1986) shows that over the Alpha Ridge, a two-tiered crustal model is sufficient to reproduce regional anomalies with bulk crustal densities of 2.88 g cm$^{-3}$ to 26 km depth and 3.04 from 26 km to the Moho boundary. And Alvey et al. (2008) utilise a similar bulk density of 2.85 g cm$^{-3}$ for the ridges and basins of the Amerasian Basin in order to determine crustal thickness from regional gravity data.

The results are also consistent with the crustal structure for the northern Mendeleev Ridge presented by Lebedeva-Ivanova et al. (2006). While specific densities were not published, calculating densities from the presented seismic velocities reveals a crustal density structure of 2.53–3.08 g cm$^{-3}$ at 29 km and an underplated layer of 3.08–3.29 g cm$^{-3}$ as converted from the seismic velocity profiles (Ludwig et al. 1970; Brocher 2005).

The seismic refraction experiment presented by Lebedeva-Ivanova et al. (2006) is the only such work to observe the crust–mantle boundary at the Mendeleev Ridge, which at the ridge crest is observed at 32 km. Through interpretation of individual velocity layers and the velocity structure as a whole, the authors suggest the Mendeleev Ridge is composed of ‘thinned underplated continental crust or thickened oceanic crust’, but they prefer a continental origin.

In a benchmark paper, Christensen & Mooney (1995) compile global results for the velocity structure of continental crust. Rifted continental crust reveals the highest velocity gradients of the various tectonic environments presented. Velocity gradients for rifted
continental crust are considerably less than those observed by Lebedeva-Ivanova et al. (2006), potentially arguing against a continental origin for the Mendeleev Ridge. The velocity structure presented by Lebedeva-Ivanova et al. (2006) is more consistent with thickened oceanic crust observed at oceanic plateaus (Fig. 16; Hussong et al. 1979; Flovenz & Gunnarsson 1991; White et al. 1992; Coffin & Eldholm 1994), or volcanic continental margin crust, as described by Eldholm & Grue (1994) over the NAVP (Hatton Bank, Voring Margin, Jan Mayen Ridge, SE-Greenland, NE-Greenland, More, Lofoten and Vestbakke Volcanic Province).
4.2 Structural history

MCS images of the Mendeleev Ridge reveal many extensional structures, with large graben and half-graben structures visible throughout. Bathymetry is in large part controlled by normal faults, which also influence sediment distribution. Similar extensional features are observed on the Alpha Ridge, though perhaps to a lesser degree (Jokat 2003).

Fig. 18 shows structural map of the ridge. Regional bathymetric contours from the International Bathymetric Chart of the Arctic Ocean (IBCAO; Jakobsson et al. 2000) are superimposed onto the regional gravity grid from the Arctic Gravity Project (AGP; Kenyon & Forsberg 2001), along with the ship's bathymetry and interpreted structures. To produce the map, structural elements from MCS images are projected to the ship's bathymetry. The orientation of faults can typically be observed on the ship's bathymetry, for example, the lineament at the base of a scarp. These locally observed structures are finally analysed to test whether they can be extrapolated according to regional bathymetry, gravity or magnetic data.

The primary extensional axis is E–W to NE–SW. Regional gravity data are found to be a good predictor of the large scale structures. Anomaly highs are correlated with horsts, anomaly lows are associated with grabens, and anomaly gradients are normal to the inferred axis of extension.

Similar to the Mendeleev Ridge, MCS data from the Chukchi Plateau suggest large scale ~E–W extension. Given the common extensional texture, it is inferred that the surveyed western edge of Chukchi Plateau experienced the same extensional event that deformed the basement and lower sediment Unit II of the Mendeleev Ridge. Gravity modelling results suggest a geological boundary between the Mendeleev Ridge and the Chukchi Plateau. The simple density–depth model that accurately predicts gravity anomalies over the Mendeleev Ridge is not effective over the Chukchi Plateau suggesting the two features have distinct compositions and separate origins (Fig. 13c).

4.2.1 Regional unconformity and timing of most recent tectonism

The regional unconformity separating sediment Units I and II appears to mark the end of extensional deformation of the Mendeleev Ridge and perhaps the Alpha Ridge. The unconformity can be traced onto the Chukchi Plateau until it pinches out at the crest of the plateau where ice erosion may have removed upper sediments (Polyak et al. 2003). Hall (1970) recognized that sedimentation post-dated deformation of the Mendeleev Ridge, though he attributed this deformation to fracturing rather than extension. Over the Alpha Ridge, Jokat (2003) also identifies a strong reflector separating two sediment units at similar subsurface depth to those reported here (~250 m), and also observed extensional block faulting of the upper crust.

Observed sedimentation rates from the Lomonosov Ridge can be used to make a crude estimate of the age of this unconformity. Moran et al. (2006) chronicles the recent ACEX drilling of the Lomonosov Ridge. Backman et al. (2008) present a refined age model for the ACEX cores as well as describes a hiatus over the Lomonosov Ridge spanning 18.2–44.4 Ma. Calculated sedimentation rates for the Neogene and Palaeogene were 12.4 and 17.6 m Myr$^{-1}$, respectively. Assuming ~0.3 s TWT to the unconformity observed in this study (Figs 4–8), 1.6 km$^{-1}$ sediment velocity and sedimentation rates calculated from the ACEX drilling, the base of sediment Unit I over the Mendeleev Ridge is dated at ~19 Ma. This age is anomalous as it significantly postdates the opening of the
Amerasian Basin (\(\sim 153–127 \text{ Ma}\)), the expected formation of the Alpha and Mendeleev ridges (120–78 Ma) and the initial opening of the Eurasian Basin (\(\sim 56 \text{ Ma}\)).

If in fact the Lomonosov sedimentation rates are appropriate for the Mendeleev Ridge then tectonism may have persisted in the Amerasian Basin well into the Cenozoic. This is an anomalous result as such recent tectonism has not been documented in the Amerasian Basin.

We must consider though that the Lomonosov sedimentation rates may be inappropriate for the AMR, particularly during the early history of the Lomonosov Ridge when it was closer to the Barents Shelf and less isolated from sediment sources than the AMR. From this we may assume that lower sedimentation rates are more appropriate. As originally put forth by Jokat (2003), under this regime it is possible that the unconformity signals the onset of rifting that led to the opening of the Eurasian Basin at \(\sim 56 \text{ Ma}\).

We must also consider that the unconformity is not tectonically controlled. Working from the same data presented in this paper, the palaeo-oceanographic and depositional environment of the Mendeleev Ridge are more thoroughly considered in the work by Bruvoll et al. (in preparation) where sediment sections over the Mendeleev, Alpha and Lomonosov ridges are analyzed and compared. These authors present a different hypothesis for the regional unconformity, preferring instead a primarily depositional origin, with the opening of the Fram Straight as a possible driving mechanism.

4.3 Models for the origin of the Alpha Mendeleev Ridge (AMR), implications for the Amerasian Basin and potential global analogues

In past studies, emplacement of the AMR is constrained to the mid-late Cretaceous (120–78 Ma) due to: the age (\(\sim 82 \text{ Ma}\)) of the basalts recovered from the ridge (Mühe & Jokat 1999; Jokat 2003), the oldest sediments (Campanian) sampled from the Alpha Ridge (Thiede et al. 1990), heat flow observations (Langseth et al. 1990) and the magnetic anomaly patterns which suggest formation during the Cretaceous normal period (Weber & Sweeney 1990). This range of dates post-dates seafloor spreading in the Canada Basin (\(\sim 148–127.5 \text{ Ma}\)) and pre-dates spreading in the Eurasian Basin (\(\sim 56 \text{ Ma}\)–present).

Considering the constraints set by this study and previous studies of both the Mendeleev and Alpha ridges, three potential tectonic environments are envisioned that lead to the emplacement of the AMR: (1) A rifted volcanic continental margin, (2) an oceanic plateau formed at a spreading centre—perpendicular to the AMR and (3) an oceanic plateau formed at a spreading centre—parallel to the AMR.

The degree to which each model is consistent with the rotational model for the opening of the Amerasian Basin varies. It is considered though that none of the three models negates a rotational origin for the Canada Basin. All three hypotheses are consistent with the presence of a LIP during the late Cretaceous. Results from the
Canadian Arctic margin, N. Greenland, Svalbard, Franz Joseph Land and the E. Siberian Sea all reveal Cretaceous basaltic volcanism with two main pulses at ∼130 and ∼95 Ma, ∼95 Ma being more commonly observed (Maher 2001; Estrada & Kunst 2004; Drachev & Saunders 2006; Weaver et al. 2006; Villeneuve & Williamson 2006).

Recovery of highly altered alkali basalts from the Alpha Ridge on separate expeditions suggests an intraplate origin for the rocks, or at least small degrees of partial melting (Van Wagoner et al. 1986; Muhe & Jokat 1999). It should be stressed though that very little sample material was recovered and the samples exhibit large degrees of low temperature alteration. Still, the geochemistry is more consistent with the rifted volcanic continental margin hypothesis than the oceanic plateau hypotheses in that greater partial melting at or near a spreading centre would yield greater volumes of tholeiites, Ocean Island Basalts (OIB) and mid-ocean ridge basalts (MORB’s) (Rollinson 1993). However, despite being only a minor component of volcanism (90% of basalts tholeiitic), transitional rocks and alkali basalts are observed off axis in Iceland where lower degrees of partial melt and deeper sources are inferred (Saunders et al. 1997). For a more thorough explanation of these models and global analogues for the AMR, see Dove (2007).

### 4.3.1 Rifted volcanic continental margin

Prior to the opening of the Eurasian Basin, the AMR may have rifted off the Barents Shelf with a geometry similar to that of the Lomonosov Ridge (Fig. 19a). This geometry suggests that the Makarov Basin opened with a spreading axis roughly parallel to the orientation of the AMR, and formed either by passive extension or active spreading, following the rifting and magmatic emplacement of the AMR. The orientation of this rifting is consistent with the observed E–W to SW–NE extensional fabric of the Mendeleev Ridge and Chukchi Plateau (Fig. 18). This geometry is also consistent with the hypothesis of Miller et al. (2006), that the Chukotka (incl. Chukotka Plateau) element of the Arctic Alaska–Chukotka microplate originated from the west (Russia) rather than the Canadian Arctic shelf.

The AMR crust would have been created by the emplacement of Mg-rich melt into actively thinning continental crust over a broad region of asthenospheric upwelling (White & McKenzie 1989; Coffin & Eldholm 1994). Volcanism that precedes initial rifting, dynamic uplifft and the compounding of tensional stresses are all common attributes observed at other volcanic margins.

Global analogues include the NAVP as described earlier (Fig. 16; White & McKenzie 1989; Eldholm & Grue 1994; Saunders et al. 1997; Hopper et al. 2003) and the South China Sea where asymmetrical spreading and complex tectonism are attributed to changing plate directions related to subduction or the rotation of plates while spreading (Hall 2002; Lin et al. 2003). Further analogues include the continental microplates: Seychelles (Todal & Eldholm 1998), Jan Mayen, Tasman Plateau and Gilbert Seamount Complex (Muller et al. 2001).

### 4.3.2 Oceanic plateau formed at a spreading centre—perpendicular to the AMR

The AMR is estimated to contain >10 million km$^3$ of mafic material (Vogt et al. 2006). If the AMR is an oceanic plateau it is the second largest on earth, next to the Ontong Java Plateau. All plateaus anywhere near this volume formed from mantle plumes, with some interaction with an active spreading centre (Coffin & Eldholm 1994).

In this framework the AMR was created by excess melt at a spreading centre similarly oriented to that, which formed the Canada Basin (Fig. 19b). The AMR would be perpendicular to, and symmetric about this spreading axis. The oldest crust would be near the Canadian and East Siberian Sea margins. Due to the presumed younger age of the AMR, incipient spreading would have post-dated the opening of the Canada Basin. But because uncertainties concerning the age of the AMR, overlap in these two spreading events cannot be precluded.

This model requires that the Chukchi Borderland occupied its current position before emplacement of the AMR, presumably rifted off of the E. Siberian Shelf after initial opening of the southern Canada Basin (Grantz & May 1982). This model also suggests that the western margin of the Chukchi Plateau as well as the eastern margin of the present-day Lomonosov Ridge served as transform margins accommodating seafloor spreading and the emplacement of the Mendeleev Ridge. This geometry may be difficult to justify considering how drastically the Chukchi Borderland protrudes into the Amerasian Basin (Fig. 1), and the current E–W to SW–NE extensional fabric of the Chukchi Borderland and the Mendeleev Ridge (Fig. 18). Moreover, while the Marvin Spur on the Amerasian side of the Lomonosov Ridge may provide bathymetric evidence of such motion (Cochran et al. 2006), no large-scale fault has been observed through other geological or geophysical studies.

### 4.3.3 Oceanic plateau formed at a spreading centre—parallel to the AMR

Geometrically this model is similar to that of the rifted volcanic margin hypothesis (Fig. 19c). Seafloor spreading would have predominately ceased in the Canada Basin before being activated, perhaps under the influence of a mantle plume, along two other arms (Alpha and Mendeleev ridges). The Makarov Basin would have formed as a part of this spreading event, opening roughly perpendicular to the former Barents Shelf.

This model is attractive in that it provides a more direct means to move the Chukchi Plateau into its current position by rifting off the E. Siberian Sea before emplacement of the Mendeleev Ridge. It would also create a structural fabric that is coincident with the later, observed E–W to SW–NE extension of the Mendeleev Ridge and Chukchi Plateau.

The AMR has long been hypothesized as an oceanic plateau (Forsyth et al. 1986; Weber 1986; Weber 1990). Most of this work however regards the Alpha Ridge, and only Jackson et al. (1986) extrapolated these results to the Mendeleev Ridge.

Lawver & Muller (1994) and Lawver et al. (2002) go further and argue that the AMR is an oceanic plateau and an expression of the Amerasian Basin’s rotation over the Iceland hotspot. The author’s reconstruction places the Iceland plume at Ellesmere Island at 130 Ma. This date is mostly incompatible with onshore, presumed equivalent volcanic rocks in that area, where volcanics are dated at ∼95 Ma (Tarduno et al. 1998; Weaver et al. 2006), though ~128 Ma rocks are also observed (Villeneuve & Williamson 2006). Continuing to follow it back, the reconstruction is even more incompatible for the AMR (120–78 Ma) and thus is considered unlikely that the Iceland Plume supplied the magma that formed the AMR.

Potential global analogues for the AMR include Iceland (Flovenz & Gunnarsson 1991; Saunders et al. 1997; Fitton et al. 1997), the Ontong Java Plateau (Hussong et al. 1979; Berger et al. 1992;
Coffin & Eldholm 1994) and the Kerguelen Plateau (Charvis & Operto 1999; Frey et al. 2002; Konnecke et al. 1998). Each of these features was formed in a unique tectonic environment, but common to their origin is the interaction between excess mantle melting and a spreading centre (Lassiter & DePaulo 1997).

These analogues share characteristics of the AMR, which include but are not limited to: a thickened oceanic crust (∼15–40 km), comparable velocity structures (both upper crust and full crustal structure; Fig. 16), and a predominantly homogenous basement character. Where subbasement reflectors are observed, they are typically representative of basalt flows intercalated with sediments. One interesting aspect of the Kerguelen Plateau is the possibility of admixed continental crust amongst the ocean island basalts. While contentious, interpretations based on reflection character, velocity gradients and the recovery of meta-igneous granulite xenoliths, have led to hypotheses describing the presence of captured slivers of volcanic continental crust (Operto & Charvis 1995; Frey et al. 2002), or the possibility of continent nucleation in an oceanic setting (Gregorie et al. 1998).

5 CONCLUSIONS

The AMR is the single largest edifice in the central Arctic Ocean. Until recently, the Alpha Ridge has been more thoroughly studied than the Mendeleev Ridge. Most studies conclude that the Alpha Ridge is an oceanic plateau, formed by plume interaction with a spreading centre, similar to Iceland. However, many authors cannot rule out a continental origin.

Due to the paucity of both geological and geophysical data collected at the Mendeleev Ridge, the origin and history of the ridge has long been enigmatic. Hypotheses regarding ridge formation are often inferred from the Alpha Ridge and range from: plume affected spreading centre to Hawaii-type hotspot track to attenuated continental crust.

During the Mendeleev Ridge leg of the HLY-05-03 expedition, bathymetry, MCS, seismic refraction and gravity data were collected that further enhance our understanding of the ridge and its relationship to neighbouring features, the Alpha Ridge and Chukchi Plateau.

MCS images reveal two primary sediment units and a mostly homogenous upper crust. Interpretations of isolated subbasement coherent reflectors are ambiguous and may represent Mesozoic or older lithified sediments, volcanic flows, or intercalated sediments and volcanics. Extension of the ridges is inferred along an E–W to NE–SW axis, which led to pervasive normal faulting of the basement and lower sediments, and the development of large horst and graben structures.

The two sediment units are separated by an unconformity that appears to mark the end of extensional deformation of the ridge, and
likely persists across the whole of the Mendeleev and Alpha ridges. Total sediment thicknesses range from 0 km at basement exposed scarps to ~1.2 km in the deep grabens. Tentative dating of the unconformity, applying sedimentation rates from the Lomonosov Ridge, suggests tectonism in this region may have persisted well into the Cenozoic (~19 Ma). But as these rates may be inappropriate for the AMR, incipient spreading of the Eurasia Basin may explain the unconformity.

Modelling of the seismic refraction data reveals an upper crustal velocity structure (3.5–6.4 km s⁻¹) that is mostly inconsistent with normal oceanic crust. The results are more compatible with both the crust of volcanic rifted continental margins similar to the NAVP, and the crust of oceanic plateaus similar to Iceland, and the Kerguelen and Ontong Java Plateaus. However, these comparisons remain tentative as we sampled only the upper crust in this experiment. Basement velocities reported here may represent high velocity sediments (carbonates), or oceanic layer 2.

Gravity anomalies over 2–225 km profiles crossing the Mendeleev Ridge can be reproduced with models containing bathymetry, sediment and basement horizons, and a single density crustal layer of 2.86 g cm⁻³. This result is indicative of laterally homogenous, predominantly mafic crust for the Mendeleev Ridge. Model misfits over the Chukchi Plateau suggest that it likely has a different composition than the Mendeleev Ridge, and a separate emplacement history. The similarity of both the velocity and density structures between the Mendeleev and Alpha ridges, corroborated by potential field and bathymetry observations suggests the ridges are a contiguous feature, sharing a common geological origin.

A comprehensive model for the tectonic evolution of the Amerasian Basin requires the thorough understanding of the AMR. A unique solution to the geological origin and history of the ridge is not yet apparent and deep-sea drilling of the ridge may be required to answer this question. Three emplacement models for the AMR that satisfy constraints set by this and previous studies are: (1) rifted volcanic continental margin, (2) oceanic plateau formed at a spreading centre—perpendicular to the AMR and (3) oceanic plateau formed at a spreading centre—parallel to the AMR. Models 1 and 3 are more compatible than model 2 with the observed E–W to NE–SW extensional fabric of the Mendeleev Ridge and Chukchi Plateau.

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REFERENCES


Figure 4. Interpreted multichannel seismic (MCS) line 17 (below) with uninterpreted close-up views of key features shown in insets above. Two-way travel time (TWT) is given on the Y-axis, with common depth point (CDP) given on the X-axis. Locations of MCS lines are shown in Fig. 1. Rectangular boxes with seismic velocities show results from seismic refraction modelling of sonobuoy data.
Figure 5. Interpreted multichannel seismic (MCS) line 18 (below) with uninterpreted close-up view of key feature shown in inset above. Locations of MCS lines are shown in Fig. 1.
Figure 6. Interpreted multichannel seismic (MCS) line 20 (below) with uninterpreted close-up views of key features shown in insets above. Locations of MCS lines are shown in Fig. 1. Rectangular box with seismic velocities shows results from seismic refraction modelling of sonobuoy data.
Figure 7. Interpreted multichannel seismic (MCS) lines 21, 22 and 23 with uninterpreted close-up view of key features shown in top left inset. Locations of MCS lines are shown in Fig.1.
Figure 8. Interpreted multichannel seismic (MCS) lines 24 (above) and 25 (below). Locations of MCS lines are shown in Fig.1.