Late Pliocene basaltic volcanism on the Western Barents Shelf margin: implications from petrology and $^{40}$Ar–$^{39}$Ar dating of volcaniclastic debris from a shallow drill core

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Homogeneous and glassy volcaniclastic debris of excellent preservation was cored during IKU shallow drilling southwest of Bjørnøya in 1989. The present paper focuses on the age and origin of the volcaniclastics based on petrological and isotopic studies of glasses and whole-rocks and their occurrence in relation to the sediments. Chemical analyses of ash, lapilli and blocks all show fairly similar, relatively alkali-rich subalkaline basaltic composition. Sm–Nd and Rb–Sr isotopic signatures are comparable with melting from anomalously enriched mantle or alternatively modification by interaction with anomalously enriched mantle or lower crustal components. $^{40}$Ar–$^{39}$Ar dating by the stepwise heating technique determined a crystallization age of ca. 2.3 Ma for the volcanic eruptions. The activity was terminated by sliding and re-sedimentation of both the volcanics and older Tertiary sedimentary deposits at the Western Barents Shelf margin. The present data are used to constrain the timing of the thermal–tectonic evolution and the erosional history in the vicinity of the Vestbakken Volcanic Province and Stappen High areas.

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Enormous Cenozoic sediment aprons accumulated oceanward of the western margin of the Barents Shelf, due to erosion of Phanerozoic sediments from the shelf. Part of this succession was penetrated by five cores drilled at sites southwest of Bjørnøya by IKU in 1985 and 1989 (Figs. 1 and 2). The youngest of the bedrock cores (7316/03-U-01), located at 297 m water depth, contains an interval almost 150 m thick of sediments with abundant volcaniclastic debris beds. The volcaniclastic material consisted of remarkably well-preserved glassy ash, lapilli and blocks. The present study aims to interpret the origin of the volcanic debris, considering (1) its position in relation to the sedimentary deposition and (2) the petrology of the volcanic clasts, and to compare this occurrence with volcanic development in adjacent areas. K–Ar and $^{40}$Ar–$^{39}$Ar dates are performed on selected clasts, and the results are discussed in relation to the petrographic data.

The study area consists of a series of faulted blocks of Paleocene, Eocene and Late Tertiary sedimentary rocks shown on a section between the Stappen High to the east and the Vestbakken Volcanic Province (VVP) to the west (Figs. 1 and 2). The Stappen High Experienced uplift in several periods in late Cretaceous to Late Tertiary time, partly related to movements along the Knølegga Fault Zone (cf. Gabrielsen et al. 1990).

Based on seismic data and five shallow drill cores, the geological evolution of the study area is interpreted in a complementary paper (Sættem et al. 1994). The Paleogene succession consists of marine deposits of fine-grained clastics originating from erosion of the Stappen High and siliciclastic deposits (Middle Eocene). The Oligocene is marked by an erosional unconformity related to uplift, subaerial exposure and sea-level fall (Sættem et al. 1994). In Neogene time the sedimentation was characterized by mass flow from the east, and these sediments were mainly sourced from the reworking of Cretaceous and Lower Tertiary clastic deposits (Goll & Smelror in prep.). A regional unconformity at the top of the studied succession marks the base of a clastic wedge of Pliocene–Pleistocene age (Sættem et al. 1992; Eidvin et al. 1993).

Early Eocene lava flows extend over large areas from the Stappen High oceanward to the VVP (Fig. 2). These may be part of the widespread Early Tertiary volcanism related to the opening of the Atlantic Ocean and the Norwegian–Greenland Sea. Remnants of this volcanism are known from the classic Tertiary Igneous Province of NW Scotland, northern Ireland, the Faeroe Islands and Greenland and from vast extents of the submarine portions of the rifted continental margins on both sides of the North Atlantic Ocean (cf. White 1988) including the Voring plateau (Eldholm 1991). After this main opening phase, volcanic activity in the border lands was limited.

Neogene volcanism in a continental setting is known from NW Spitsbergen, where tholeiitic continental basalts were erupted in Late Miocene time (Prestvik 1978; Skjelkvåle 1987), and strongly alkaline basaltic
volcanism occurred during a much later event in Quaternary time (Skjelkvåle et al. 1989). Post-opening volcanism is also evidenced in the Vestbakken Volcanic Province in a different tectonic setting. This is shown by the presence of intrusive structures transecting the Early Eocene basalt reflectors (Faleide et al. 1988). The nature of this volcanism is not known, and the boundary between oceanic and continental crust is poorly defined in that area.

Younger volcanic activity is represented mainly by activity related to spreading along the mid-ocean ridge segments (cf. the Kolbeinsey, Mohn's and Knipovich Ridges), and at the Iceland and Jan Mayen hotspots.

The present study area is located on the continental margin to the east of the northward deflection of the Mohn's–Knipovich Ridge axes, and it represents a southern extension of the zone of thermal activity along western Spitsbergen (Fig. 1). This zone is located on the eastern (continental) side near the transition zone between continental–oceanic crust (cf. Skilbrei 1991). This is a margin of extensive Cenozoic epeirogenic uplift (Vågnes & Amundsen 1993; Sættem et al. 1994) possibly related to boundary processes between hot oceanic asthenosphere and cold deep continental lithosphere (Vågnes & Amundsen 1993). The tectonic setting of this area, however, is not well understood.

The studied core with volcanics (7316/03-U-01) was sited on the upper part of a wedge of Miocene–Pliocene sediments that is bounded by two regional unconformi-
Late Pliocene volcanism, Barents Shelf

Fig. 2. (a) Location map for the IKU shallow drillcores (filled circles) SW of Bjørnøya, and the approximate location of the schematic W–E profile shown in (c). Inset gives the location of the geological map in (b). Main structural elements are from Gabrielsen et al. (1990). Abbreviations: VVP = Vestbakken Volcanic Province, SH = Stappen High, SVB = Sørvestnaget Basin, BB = Bjørnøya, KFZ = Knelegga Fault Zone. (b) Geological map (base Quaternary) of the study area based on Sættem et al. (1993). (c) Interpretative geological cross-section from the VVP to the SH. Note location of core 7316/03-U-01 in the upper part of the Miocene–Pliocene wedge (compilation by Stein Fanavoll). KFZ = Knelegga Fault Zone.
Fig. 3. Seismic cross-section of the cored interval in the upper part of the Miocene-Pliocene wedge (core 7316/03-U·Ol) showing chaotic reflection patterns (from Bugge et al. 1990). The seismic reflections 1–4 are related to the lithostratigraphic section by arrows in Fig. 4.

ties (Fig. 2). Biostratigraphic dating of the core was complicated by the extensive fossil reworking with species of Cretaceous, Paleogene and Miocene age (Goll & Smelror in prep.). In the present study we provide radiometric dates for the volcanics, which are as young as Late Pliocene. This implies that all fossils within the cored interval are reworked. Radiometric dating of the volcanics also provides the timing of sediment deposition and constrains the age of the two regional seismic reflectors that are recognized in the cored interval. This also has implications for the interpretation of the erosional history which is discussed elsewhere (Sættem et al. 1994).

The core was drilled through three different units defined on the seismic sections. The seismic reflection patterns are chaotic (Fig. 3) resembling channel erosion and infill, possibly slides and slide disturbed deposits in the middle and upper part (cf. Bugge et al. 1990; Settem et al. 1994). The upper section with volcanic debris is truncated by a regional erosion surface that can be traced westward into the Vestbakken volcanic province (Faleide et al. 1988; Gabrielsen et al. 1990; Settem et al. 1994). This level is above the distinct reflector with Eocene basalts, and apparently corresponds to the level with younger intrusions in the VVP described by Faleide et al. (1988).

Sedimentological setting and appearance of volcanic debris

A simplified lithostratigraphic section is shown in Fig. 4. The lower part of the core from the base up to ca. 95 m consists of very fine-grained sandstone with scattered thin beds with volcanic debris (lapilli-tuff sandstone). This sandstone succession shows a gradational transition (fining up) to silty shale between 95 and 94 m. Overlying this is an interval of silty shale (94–71.6 m), interrupted by a 4 m thick package of volcanic debris beds with interbedded sandstone. The upper part of the core, from 71.6 m to 31.9 m (below the Quaternary), consists of fine-grained to fine medium-grained sandstone with abundant volcanic debris beds of variable thickness. The volcanic debris beds are particularly abundant in the lower part of this interval. A variety of angular to rounded pebbles of quartz and claystone fragments are dispersed in the sandstones and most commonly in the upper subunits (cf. Settem et al. 1994).

Sandstones and silty claystone

The sandstones are poorly consolidated sublitharenites to litharenites consisting of quartz, plagioclase, microcline and rock fragments of reworked glaucony, chert and sandstone. Biotite, muscovite, chlorite, epidote, garnet, amphibole, zircon, rutile, sphene and opaque ore minerals are present in accessory amounts. Very fine-grained sandstone in the lower interval is well sorted with high porosity, while fine and fine to medium-grained sandstones in the upper subunit display moderate sorting, but with higher compositional maturity.

The abundant recycled sedimentary rock fragments and the reworked fossils in the core (Goll & Smelror in prep.) indicate that much of the sediments formed by erosion of Mesozoic and Cenozoic sediment deposits. The source area of the first cycle sediments included green schist and amphibolite facies metamorphic rocks.

The silty claystone intervals are mineralogically similar to the sandstones, but with a higher content of mixed-layer clay (and smectite), mica-illite (including glaucony) and kaolinite, mainly reflecting the difference in grain size. Glaucony of possible authigenic origin may be present in a greenish interval between ca. 90 and 100 m. Trace amounts of zeolite and amphibole are present in sandstone and silty claystone intervals that include volcanic debris beds. Opal (A) was identified in the upper sandstone, and more abundantly in some volcanic debris beds.

Volcanic debris beds

The volcanic debris beds consist of poorly sorted mixtures of black ash, lapilli and occasional coarser blocks, all of basaltic composition, intermixed with glauconite-bearing sand similar to the host sediments (Fig. 5, 6 and 8). The composition ranges from sandy lapilli tuff to sandstone with dispersed volcaniclastic debris. The majority of the volcanic clasts have retained the irregular and angular shapes of the scoria. Some of the lapilli-tuff-
Late Pliocene volcanism, Barents Shelf

Fig. 4. Lithostratigraphic section (left, from Sættem et al. 1993) and sketches of selected volcanic debris beds (A–T) within sandstone. The proportions of clay, silt and sand are shown by symbols in the left (lithostratigraphic) column, where volcaniclastic beds are indicated by open clast symbols whereas filled triangles record other pebble-sized clasts. G = glauconite. The volcaniclastic debris beds (A–T) are sketched from studies of cores and core photos. The diameter of the core is ca. 5.5 cm. Basaltic blocks, lapilli and ash are shown by the filled (black) irregular clasts and larger dots within sandstone (fine dotted). Arrows mark the locations of the seismic reflections in Fig. 3.
Fig. 5. Photos of core slabs demonstrating the occurrence of volcanic material in core 7316/03-U-01. (a) 50.20–50.50 m. A thin bed with volcanic debris in sandstone. Oblique bedding is due to loading (b) 57.00–57.40 m. Note inversely graded 35 cm thick volcaniclastic bed above oblique contact with underlying sandstone. Light colour in upper part reflects opal cementation. (c) 78.00–78.30 m. Basaltic block/pillow within lapilli sandstone appears in the lower part. Note light coloured opal-cemented tuffaceous sandstone in upper part. (d) 100.30–100.60 m. Normal graded volcaniclastic bed. (e) 153.40–153.65 m. Lowermost continuous bed with volcaniclastic debris overlain by poorly cemented sandstone and a cemented volcaniclastic debris bed. Note loaded contract at the base of the upper bed.

Fig. 6. Optical micrograph of tuffaceous lapilli sandstone at 49.57 m (bed D) showing vesicular and less vesicular lapilli clasts (pale brown) in a matrix of ash and finer lapilli (light brown) and very fine-grained sand (white subrounded/subangular grains = quartz, dark = glaucony, mica and clay matrix). Vesicles are shown as green when filled with coloured epoxy, while empty vesicles appear white.)
Late Pliocene volcanism, Barents Shelf

Fig. 7. (a) Photomicrograph of basaltic lava selected for Ar–Ar dating (32.05 m, parallel polars), showing microphenocrysts of olivine and very thin plagioclase microlites (white) in a dark glassy matrix. Note the appearance of a homogeneous glassy rind around the vesicles and of smaller pores within the larger (pores are coloured by green epoxy). Excess argon may have been trapped within the micropores and glassy rinds. (b) Photograph of basaltic lava selected for Ar–Ar dating (77.70 m, parallel polars). Note higher content of well-developed microphenocrysts dominated by plagioclase microlites (white). (c) Close-up of (b) showing plagioclase laths and clinopyroxene within a glassy matrix. The absence of excess argon within this sample as compared to the sample in (a) may be caused by Ar-escape related to the more extensive microphenocryst growth during the eruption and the formation of microcracks during the rapid cooling of the glass.
rich beds have a whitish color due to cementation by silica (opal, Fig. 5).

Volcanic debris beds in the lower sandstone interval are thin (generally 2–10 cm thick). Six volcaniclastic beds from this part are illustrated in Fig. 4 at P, Q, R, S, T. These are all matrix supported and generally poorly sorted. The coarsest lapilli (2–3 cm clast) tend to be dispersed in the lower part (beds P, R) or in the upper part (beds S, Q) of distinct beds, indicating the presence of both normal and inverse grading. A bed at 109.5 m displays inverse grading in the lower centimeters by the increasing concentration of lapilli.

The package of volcanics in the central part of the core between 80 and 75 m includes three volcaniclastic beds (M, N, O – up to more than 1 m thick). Bed O shows a high concentration of lapilli clasts and displays normal grading from a nearly clast-supported lower part to a matrix-rich upper part. Occasional coarser clasts (2–3 cm) are dispersed in the upper part. Another bed (N) contains block-sized basalt clasts in the lower half. This bed is matrix rich and also displays increasing matrix content in the upper half. Slumping is evidenced by sandy lapilli-free lobes within the central part (beds N, M). The apparently unsorted structure of bed M is influenced by slumping.

Volcaniclastic beds in the upper sandstone interval have variable thickness between 2 and 65 cm. The thickest beds (I, D) are matrix-supported and non-sorted with a moderate concentration of lapilli clasts throughout the intervals and with dispersed coarser clasts. A thin bed (G) displays normal concentration grading, while another (F) shows inverse concentration grading. The latter includes steeply oriented fringes of semi-consolidated lapilli-free sand and a sand inclusion with lapilli. Another bed (67.2 m) includes a normally graded lower part and an inversely graded upper part by concentration of lapilli. Lapilli-rich beds with dispersed coarser clasts near the top are exemplified by ‘K, L’ and completely non-sorted beds are also present (C). Slumping is illustrated by the appearance of bed B.

In addition to the well-defined volcaniclastic beds, isolated coarse lapilli clasts are occasionally found dispersed in the sand (37.5 m). Some of the latter may, however, have been displaced during drilling.

In summary, the volcaniclastic debris beds are commonly oriented obliquely to the core and the faint lamination locally observed in the sediments, and they also show evidence of slumping and occasionally of shear/faulting against the underlying sediments (cf. H, M). Altogether, the structures are chaotic and typical for debris flow deposits (see discussion below).

Petrography of glasses and sediment inclusions

Ash-sized clasts in the core are mainly in the range between 0.01 mm and 2 mm, and the lapilli-sized clasts present range from a few mm up to 1–2 cm. Lapilli up to 6 cm in size and block-sized clasts are occasionally present.

The ash-sized clasts and most of the lapilli-sized clasts consist of remarkably well-preserved homogeneous glass (sideromelane) appearing pale brown in the optical microscope, and less abundant darker brown or black coloured fragments (tachylite). The ash-sized clasts are angular, equant or irregular (cuspate) with curved surfaces defined by pre-existing larger vesicules (Figs. 6 and 8). The ash fragments may have formed by two mechanisms: (1) disruption and brittle deformation of larger vesicular lapilli clasts and (2) shattering due to quenching in contact with water, ice or sediments (hyaloclastite).

The lapilli-sized clasts are dominantly slightly vesicular sideromelane, but show different concentration and size of vesicules. Lapilli with deformed (stretched) vesicules are occasionally present. Clasts of centimeter size and block-sized clasts are commonly tachylitic, but may include relics of homogeneous sideromelane in central parts. Palagonitized volcanic debris was detected in one of the cemented sandy lapilli-tuff beds (Fig. 8).

The host sediments are closely intermixed with the volcanic debris, and commonly occur in the larger open vesicules in the glass fragments. These may include smaller ash fragments intermixed with the sediments (Fig. 8). Most of the relations indicate that sediments and glass were intimately mixed after the solidification of the melt, but some clasts also show evidence that sediments were incorporated into the basaltic melt when it was still hot. This is illustrated in Fig. 8, where late-formed quench crystals are radiating out from an inclusion of a rounded detrital quartz grain (77.10 m).

Isolated sedimentary xenocrysts in the glass were also detected in dark lapilli-sized clasts at 111.45 m, 106.73 m and 77.97 m, indicating that sediments may have been assimilated in the melt during the eruption.

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Fig. 8. Backscattered electron images of lapilli clasts. Abbreviations: Q = detrital quartz, g = volcanic glass, 0 = olivine. Scale bar is in micrometres. (a) Vesicular lapilli clast (white) in sandstone matrix (dark grey) with quartz grains (light grey) and ash fragments (white). Note sediment-filled open vesicules which include tiny ash fragments of similar composition as the coarser clasts (49.57 m). (b) Skeletal olivine microphenocrysts (grey) within a glassy lapillius clast (light grey) (153.35 m). (c) Rounded quartz xenocryst (dark grey) in lava matrix containing microlites of plagioclase (light grey) and clinopyroxene (light). Note radiating fabric of microlites around the quartz (77.10 m). (d) Detail of (c) showing the quartz–lava interface. Note high density of crystals that have nucleated on the detrital quartz inclusion. This suggests rapid cooling of the magma in contact with the sediments. (e) Sili–opal cemented sandy tuffaceous lapillistone at 77.97 m. Oval structures are opal-cemented sediment-filled vesicules within the lapilli (white). Black = vesicules and lack of material. Dark grey grains within the vesicules are detrital quartz and feldspar.
Interpretation of the eruption and deposition

The generally crystal-free and homogeneous glass composition shows that the magma was subjected to very rapid cooling. This may have taken place in contact with water, ice and/or wet sediments. The above-mentioned growth of quench crystals on sediment inclusions shows that sediments were included when the melt was still hot, and thus that parts of the melt were erupted through or into the poorly consolidated sediments. The vesicularity of the glasses is variable. Whereas some of the smaller and massive ash fragments resemble subglacial eruption textures (cf. shards in figure 10.16 in Fisher & Schmincke 1984), the more vesicular ash and lapilli-sized glass possibly includes both subaqueous and subaerial products (cf. examples from Surtsey and Reykjanes, figures 10.14 and 10.16 in Fisher & Schmincke 1984). The differences in vesicularity could be related to phases of different explosivity during eruption, as reported for recent volcanos (cf. Surtsey, Kokelaar 1983). However, such eruption products may show great variation in vesicularity even on the centimeter scale.

The sedimentary deposition was by various mass-flow mechanisms in marine environments, and the seismic and sedimentological data indicate that sediment transport occurred by both high density turbidity currents and debris flows (cf. Sættem et al. 1994). The variable and commonly poor sorting in the volcanic beds represents different energetic regimes of debris flows. The structures were modified by synsedimentary slumping, probably related to rapid sedimentation or to pushing by loading in adjacent parts of the deposit. The sandstones and volcanic debris beds in the upper core interval represent debris flow and other mass-flow deposits in an upper slope environment. Different beds display low and high turbulence parts of flows, and chaotic structures may also result from loading and probably sliding of newly deposited sediment masses. Preservation of the oblique sand inclusions and sediment incrustations described for bed F (Figs. 4 and 5) may reflect conditions of low turbulence during deposition, possibly a very local sliding of semiconsolidated sediments.

The high concentration of the volcaniclastics and the varied size and shape and the excellent preservation of the glass exclude a long and multistage sedimentary history after the eruption. The volcaniclastics were probably eroded from newly extruded volcanic deposits. An eastern sediment transport direction indicated from the seismic data suggests that the eruption site was located somewhere between the Stappen High and the site of the core, and thus well to the east of the post-rifting intrusions described by Faleide et al. (1988).

Volcanic eruptions are known to be accompanied by tectonic activity (cf. earthquake activity) and the building up of unstable structures (piles of volcaniclastics). The volcanism described in the present study occurred in a setting of poorly consolidated Tertiary and older sediments. It is therefore reasonable that the volcanotectonic activity in this case also may have acted as a triggering mechanism for the sliding and mass flows. The excellent preservation of the glasses indicates that they were rapidly buried and protected from circulating fluids, possibly during or shortly after the eruption. The whole succession is interpreted as formed by rapid re-sedimentation of Eocene–Pliocene sediments shortly after the younger Cenozoic volcanic products were erupted through them. In the paper by Sættem et al. (1994) possible glacial influences on the sediment supply are also discussed.

Petrology of the volcanics

Microphenocrysts

Most of the ash-sized glass fragments are devoid of crystals. However, dispersed grains of olivine may be present within the coarser of the ash-sized fragments and also in lapilli and block-sized clasts. The olivine grains occur as euhedral, embayed or skeletal phenocrysts in the glass (Figs. 7 and 8). Euhedral Cr-spinel has been detected as inclusions in olivine. The composition of the olivine phenocrysts is fairly homogeneous (Fo81–Fo86) throughout the core. Block-sized fragments may also include matrix olivine with slightly more evolved composition (Fo75–Fo82).

Plagioclase microlites (labradorite, An40–An65) appear in very thin needles (0.06 up to 1 mm long) in some of the dark and oxidized larger clasts (lapilli to block size). Clinopyroxene was detected occasionally in quench textures with branching plagioclase–clinopyroxene intergrowths (down to 1 μm thin crystals, cf. sample 32.05 m). Microphenocrysts of salitic–augitic clinopyroxene were recorded in one lapillus clast (at 77.70 m, Fig. 7b). This forms an intergrowth with up to 0.6 mm long plagioclase laths (An86). Minor olivine (Fo74–80) and abundant euhedral to skeletal ilmenite grains are also present. The ilmenite may have formed in a late stage associated with resorption of olivine. The microphenocrysts are surrounded by homogeneous glass that formed by quenching of the remaining melt after the phenocryst growth.

Glass and whole-rock compositions

The chemical composition of the volcaniclastic debris was analysed in order to characterize the volcanism and to determine whether different types of volcanic rocks are present. This was done by quantitative electron microprobe analyses of ash and lapilli clasts from 19 levels in the core between 153.35 m and 38.13 m. Analysis was carried out by using a slightly defocused electron beam (10 μm) to prevent loss of the alkali elements. In addition, XRF whole-rock analysis was performed on two of the larger clasts for comparison. The chemical composi-
tion of the unaltered glass is representative of the composition of the melt at the stage of eruption.

The glasses are fairly silica-rich, olivine and hypersilicate normative basalts with ca. 51% (±0.8) SiO₂, 6% (±0.5) MgO and 3.5% (±0.4) Na₂O (Table 1). A subalkaline to alkaline transitional affinity is indicated in the Na₂O + K₂O versus SiO₂ diagram, using the dividing lines of Kuno 1966, Irvine & Barager 1971 and Miyashiro 1978 (Fig. 9). The K₂O and Na₂O contents are significantly higher than in the tholeiitic ocean ridge basalts which are associated with spreading, as exemplified by the Quaternary volcanics dredged from the Mohn's–Knipovich Ridge to the west of the study area (cf. Neumann & Schilling 1984). A calc-alkaline character is indicated by the SiO₂–K₂O and the FeO–MgO–K₂O+Na₂O diagrams (cf. Irvine & Baragar 1971), whereas the FeO/MgO–SiO₂ diagram (Miyashiro 1974; Rickwood 1989) indicates a tholeitic–calc-alkaline transitional composition. The K₂O–SiO₂–P₂O₅ ratio plots in the continental basalt field of Pearce et al. (1975).

The chemical analyses from different levels in the core show no regular variation with core depth and mostly show overlapping compositions. Some scattering is present, but there is no indication that different compositional groups are present. More than 80% of the total analysed grains represent homogeneous pale brown glass. A mean value with standard deviation for these pale brown glasses is shown in Table 1. This composition is indistinguishable from the composition of the dark brown glass fragments (Table 1). The black oxidized clasts also show similar compositions, but include a larger scatter, some of which may be due to minor secondary alteration associated with the oxidation.

XRF whole-rock major and trace element analyses of two of the larger basaltic blocks at 32.05 m and 78.05 m are closely similar (Table 1). The major element compositions also show excellent agreement with the analyses of the homogeneous ash and lapilli clasts, except that the MgO content is slightly lower in the latter. This difference reflects the presence (bulk rock) or absence (glasses) of phenocrysts of olivine.

Trace element compositions of the two whole-rock samples are also nearly identical (Table 2). The trace element distribution shows an enrichment of LIL elements relative to the average MORB composition (Fig. 10) comparable with the mildly alkaline nature of the volcanics. The trace element distribution may be similar to ocean island basalts and to some continental plateau

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**Fig. 9**. Alkali-silica plot for basaltic glass from the present study (BJW). Microprobe analyses are shown by encircled area (cross = standard deviation from Table 1). Two whole-rock basalt analyses from BJW (Table 1) are shown by dots. The compositions are compared with basalts from Spitsbergen (SPI-Q = Quaternary and SPI-T = Tertiary basalts, Skjelkvåle et al. 1989, Prestvik 1978, Hagen 1989), Mohn's Ridge E, W (MR-E, MR-W), Knipovich Ridge (KNR) and the Jan Mayen Platform (JMP, x) Neumann & Schilling 1984). Dividing lines after Irvine & Barager (1971), where the FeO/MgO–SiO₂ diagram (Miyashiro 1974; Rickwood 1989) indicates a tholeitic–calc-alkaline transitional composition. The K₂O–SiO₂–P₂O₅ ratio plots in the continental basalt field of Pearce et al. (1975).

**Table 1.** Chemical composition of volcanic debris represented by mean values of electron-microprobe analyses of glasses (a, b, c) with standard deviation shown in parentheses. XRF whole-rock analyses of two of the largest blocks are also shown (d, e). Ferric/ferrous iron ratio was determined by titration for samples d and e. A Fe₂O₃/FeO ratio of 0.15 was used for recalculation of the glass composition in a, b and c.

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<th>Element</th>
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<th>b</th>
<th>c</th>
<th>d</th>
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<tr>
<td>P₂O₅</td>
<td>0.25 (0.07)</td>
<td>0.26 (0.05)</td>
<td>0.30 (0.09)</td>
<td>0.22</td>
<td>0.23</td>
</tr>
<tr>
<td>L.O.I.</td>
<td>n.a.</td>
<td>n.a.</td>
<td>n.a.</td>
<td>0.72</td>
<td>1.11</td>
</tr>
<tr>
<td>Sum</td>
<td>98.20</td>
<td>98.27</td>
<td>98.19</td>
<td>99.87</td>
<td>100.41</td>
</tr>
<tr>
<td>Number of analyses</td>
<td>111</td>
<td>17</td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>
Table 2. XRF analyses of trace elements (ppm) of the whole-rock samples (d, e) above.

<table>
<thead>
<tr>
<th>Sample</th>
<th>Ni</th>
<th>Cr</th>
<th>Sc</th>
<th>V</th>
<th>Ba</th>
<th>Rb</th>
<th>Sr</th>
<th>Zr</th>
<th>Y</th>
<th>Nb</th>
<th>Ga</th>
<th>Cu</th>
<th>Zn</th>
<th>Pb</th>
<th>La</th>
<th>Ce</th>
<th>Th</th>
</tr>
</thead>
<tbody>
<tr>
<td>32.05</td>
<td>252</td>
<td>314</td>
<td>25</td>
<td>189</td>
<td>352</td>
<td>23</td>
<td>340</td>
<td>138</td>
<td>24</td>
<td>23</td>
<td>20</td>
<td>81</td>
<td>98</td>
<td>3</td>
<td>21</td>
<td>53</td>
<td>3</td>
</tr>
<tr>
<td>78.05</td>
<td>275</td>
<td>313</td>
<td>26</td>
<td>183</td>
<td>375</td>
<td>23</td>
<td>369</td>
<td>143</td>
<td>24</td>
<td>23</td>
<td>20</td>
<td>89</td>
<td>104</td>
<td>3</td>
<td>33</td>
<td>50</td>
<td>2</td>
</tr>
</tbody>
</table>

Isotopic data

The hypothesis of a common origin was further tested by comparing the Sr- and Nd-isotopic compositions of the basaltic block near the top of the core (32.05 m) and a fraction of homogeneous ash fragments (1-2 mm in diameter) from a lower level (49.57 m). The latter were separated by careful hand-picking under the microscope to get the freshest glasses. The $^{143}\text{Nd}/^{144}\text{Nd}$ ratios are identical for the two samples, and the $^{87}\text{Sr}/^{86}\text{Sr}$ values are also very close (Table 3). The similar isotopic composition of these two texturally different samples supports the above conclusion of a common origin for the different kinds of clasts.

Table 3. Sm–Nd and Rb–Sr isotopic analyses of basaltic block at 32.05 m (sample d above) and of a hand-picked separate of homogeneous glassy lapilli fragments at 49.57 m.

<table>
<thead>
<tr>
<th>Sample</th>
<th>Sm</th>
<th>Nd</th>
<th>$^{143}\text{Sm}/^{144}\text{Nd}$</th>
<th>$^{147}\text{Sm}/^{144}\text{Nd}$</th>
<th>$^{143}\text{Nd}/^{144}\text{Nd}$</th>
<th>$^{147}\text{Nd}/^{144}\text{Nd}$</th>
</tr>
</thead>
<tbody>
<tr>
<td>32.05</td>
<td>4.54</td>
<td>20.3</td>
<td>0.1359</td>
<td>0.512325 ± 6</td>
<td>0.5123235 ± 12</td>
<td></td>
</tr>
<tr>
<td>49.57</td>
<td>4.71</td>
<td>21.0</td>
<td>0.1366</td>
<td>0.512324 ± 12</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>Sample</th>
<th>Rb</th>
<th>Sr</th>
<th>$^{87}\text{Rb}/^{86}\text{Sr}$</th>
<th>$^{87}\text{Sr}/^{86}\text{Sr}$</th>
</tr>
</thead>
<tbody>
<tr>
<td>32.05</td>
<td>22.8</td>
<td>337</td>
<td>0.1957</td>
<td>0.705306 ± 8</td>
</tr>
<tr>
<td>49.57</td>
<td>25.0</td>
<td>352</td>
<td>0.2052</td>
<td>0.705288 ± 6</td>
</tr>
</tbody>
</table>

Discussion

The studied volcanics from the location southwest of Bjørnøya differ in composition from the basalts along the Mohn's and Knipovich Ridge segments to the west and also from the alkaline off-ridge volcanicism in the vicinity of Jan Mayen (Neumann & Schilling 1984; Imsland 1978, 1986). The younger MORB and alkaline volcanics to the west have Nd and Sr isotopic compositions that
indicate melting of a more depleted mantle (Waggoner 1989, Fig. 11).

The major element chemistry of the volcanics southwest of Bjørnøya resembles the Late Cenozoic basalts described from Spitsbergen (cf. Prestvik 1978; Hagen 1988; Skjelkvåle et al. 1989), but with slightly more silicic compositions and with greatly different Nd and Sr isotopic signatures, as the Spitsbergen basalts show depleted MORB type mantle compositions (Amundsen 1991) and only minor evidence of interaction with continental crust. The Spitsbergen basalts originated from a depleted mantle and ascended through partly thinned continental lithosphere (Vågnes & Amundsen 1993).

The glass and whole-rock analyses in the present study are characterized by a high content of SiO2 and by mg values (Mg2+/Mg2+ + Fe2+) of 0.58–0.68. The Fo-rich olivines (up to Fo96) are, however, comparable with crystallization from a fairly primitive basaltic melt. Calculated K0 (Fe/Mg)olivine/glass are lower than the olivine/melt equilibrium values of ≥0.30 from experimental studies (cf. Roeder & Emslie 1970; Takashi & Kushiro 1983). This means that some differentiation or modification of the melt has taken place owing to or after the olivine crystallization. The fairly homogeneous chemistry of the olivines indicates crystallization at conditions of extensive diffusion at high temperature conditions. The branching textures of some of the olivine grains may reflect rapid growth and/or resorption.

Silica-rich basaltic compositions have been produced by experimental melting of mantle peridotite at fairly low confining pressures (cf. 10 Kb, Takashi & Kushiro 1983; Hirose & Kushiro 1993). Melt compositions in these experiments resemble the volcanics from the area southwest of Bjørnøya, except that the latter are richer in K2O. These experiments showed that the silica content was controlled mainly by the pressure of melting, whereas Ca, Al and incompatible elements (including K) reflected the degree of melting and compositional variations of the source peridotite. This mechanism may account for the compositions described from the mid-ocean ridges. In order to produce the melt compositions of this study, the mantle protolith must have been enriched in phlogopite, amphibole or another K-rich phase (previously metasomatized mantle?). Alternatively, the melt was contaminated by LIL-enriched melts or fluids derived either from enriched mantle or the continental crust. A possible mechanism is shallow melting of the melt due to lithospheric thinning and mantle upwelling to ca. 30 km. Possible mixing may have taken place at depths where the melt was still hot owing to the complete chemical homogenization of the volcaniclastic products.

K–Ar and 40Ar–39Ar dating of volcanic clasts

**Samples**

Five volcanic clasts from different levels in the core were selected for K–Ar radiometric dating, and two of these were also dated by the 40Ar–39Ar incremental heating method. The dated clasts are the coarsest of the volcanic clasts, comprising 3–4 cm lapilli (77.70 m, 111.45 m) and coarser blocks (and pillow?) (32.05 m, 32.10 m, 78.05 m). All the dated clasts are vesicular and include olivine phenocrysts. Four of the clasts are tachylic with very fine-grained needles of plagioclase which have formed by late-stage quenching during eruption. The lapilli clast at 77.70 m consists of homogeneous sideromelane with abundant microphenocrysts (described earlier) (Fig. 7b, c).

The selected clasts show no evidence of secondary alteration; they all look remarkably fresh, and sediment inclusions were not observed. In the tachylic samples, a peculiar crystal-free glass rind commonly appears around the vesicules (cf. 32.05 m, Fig. 7a).

**K–Ar results**

The K–Ar dates range from 2.34 ± 0.08 to 10.20 ± 0.15 Ma (Table 4). The youngest date was obtained for the sample with microphenocryt-rich sideromelane (77.70 m). There is no systematic variation between the dates and the sample depths. Indeed the upper two samples, which may be part of the same blocky lava, cover much of the spread with dates of 7.35 ± 0.15 and 3.19 ± 0.05 Ma. Above, it was argued that the volcanic clasts probably originated from the same volcanic eruption (or eruptive event) and display similar chemical signatures. The recorded variation in K–Ar dates is therefore not considered as recording real differences in eruption age, as will be discussed below.

In principle, secondary alteration and cements (devitrified glass, calcite, zeolite, clay minerals) may result in a lowering of K–Ar dates, while artificially old dates may result from the presence of xenoliths and xenocrysts, which are likely to contain excess radiogenic argon (cf. Faure 1986). Commonly, secondary alteration may be associated with radiogenic argon loss and changes in the potassium content of the basalt via chemical exchange. Interaction with sea water may add potassium to the basalt, while other hydrothermal interactions may lead to leaching of potassium. In the present samples no such effects are seen, as the K-content of the dated samples is very similar, and also similar to the homogeneous glass described earlier (cf. Tables 1 and 4). This indicates that secondary K+ supply is not a likely mechanism and, as

<table>
<thead>
<tr>
<th>Sample depth</th>
<th>Material</th>
<th>Weight</th>
<th>%K</th>
<th>% radiogenic 40Ar</th>
<th>Age (Ma)</th>
</tr>
</thead>
<tbody>
<tr>
<td>32.05 m</td>
<td>Block</td>
<td>104 g</td>
<td>0.910</td>
<td>33.2</td>
<td>7.35 ± 0.15</td>
</tr>
<tr>
<td>32.10 m</td>
<td>Block</td>
<td>40 g</td>
<td>0.906</td>
<td>34.0</td>
<td>3.19 ± 0.05</td>
</tr>
<tr>
<td>77.70 m</td>
<td>Lapillus, 4 cm</td>
<td>35 g</td>
<td>0.909</td>
<td>39.0</td>
<td>2.34 ± 0.08</td>
</tr>
<tr>
<td>78.05 m</td>
<td>Lapillus, 3 cm</td>
<td>75 g</td>
<td>0.943</td>
<td>45.9</td>
<td>10.20 ± 0.15</td>
</tr>
<tr>
<td>111.45 m</td>
<td>Block/bomb</td>
<td>23 g</td>
<td>0.899</td>
<td>24.4</td>
<td>4.39 ± 0.10</td>
</tr>
</tbody>
</table>

Table 4. K–Ar dates and sample information.
already mentioned diagenetic effects were not observed in the dated samples. Another possibility is that the variation in measured ages reflects the presence of variable amounts of retained mantle argon. In order to test this hypothesis, two of the samples with different K–Ar dates were selected for dating by the 40Ar–39Ar method.

40Ar–39Ar results

Tachylitic basalt (32.05 m) and phenocryst-rich sideromelane lapilli (77.70 m) with widely different K–Ar dates were selected for dating with the Ar–Ar stepwise heating technique. Results are given in Table 5 and Fig 12. Five temperature steps were applied for each sample in the range between 500 and 1500°C, and 40Ar–39Ar ages were calculated from the gas compositions for each step. Most of the gas was released between 900 and 1000°C. Gas released in the lowest temperature heating step of sample 32.05 m showed an excess Ar isotopic composition (giving a calculated age of 15.5 Ma), while the four higher temperature steps defined a plateau that gave a weighted mean age of 2.35 ± 0.12 Ma. By contrast, sample 77.70 m displayed a different pattern, with a slightly younger age, the lowest temperature step reflecting a small amount of Ar loss, while the three intermediate temperature steps that released 94% of the total gas defined a plateau with a weighted mean age of 2.20 ± 0.12 Ma. This is close to the K–Ar date of the same sample.

Table 5. 40Ar–39Ar dates for different incremental heating steps.

<table>
<thead>
<tr>
<th>Sample</th>
<th>Step 1</th>
<th>Step 2</th>
<th>Step 3</th>
<th>Step 4</th>
<th>Step 5</th>
</tr>
</thead>
<tbody>
<tr>
<td>32.05 m</td>
<td>15.45 ± 1.39</td>
<td>2.27 ± 0.19</td>
<td>2.37 ± 0.18</td>
<td>2.82 ± 0.50</td>
<td>2.30 ± 0.32</td>
</tr>
<tr>
<td>Plateau age:</td>
<td>2.35 ± 0.12 Ma</td>
<td>Isochron age:</td>
<td>2.31 ± 0.10 Ma</td>
<td></td>
<td></td>
</tr>
<tr>
<td>77.70 m</td>
<td>0.97 ± 0.88</td>
<td>2.40 ± 0.22</td>
<td>2.11 ± 0.15</td>
<td>2.22 ± 1.11</td>
<td>1.44 ± 1.37</td>
</tr>
<tr>
<td>Plateau age:</td>
<td>2.20 ± 0.12 Ma</td>
<td>Isochron age:</td>
<td>2.21 ± 0.52 Ma</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

39Ar/40Ar–36Ar/40Ar isochron plots from the same temperature steps that define the plateaus yield concordant ages (Fig. 12, Table 5). The 40Ar/36Ar intercepts are...
atmospheric composition, indicating that excess Ar is not present in the plateau steps and the calculated age (2.3 Ma) is a reliable measure of the time of basalt crystallization.

Discussion

While the K–Ar method resulted in clearly different dates for samples 32.05 m and 77.70 m (7.35 ± 0.15 Ma and 2.34 ± 0.08 Ma respectively) the 40Ar–39Ar incremental heating results are similar within the standard deviation (2.35 ± 0.12 Ma and 2.20 ± 0.12 Ma respectively). The release pattern of sample 32.05 indicated the presence of radiogenic argon released at the lowest temperature heating step, explaining the older K–Ar date. In contrast, the K–Ar date of sample 77.70 m is similar to the 40Ar–39Ar date, and this is compatible with the apparent lack of excess argon in the age spectrum for this sample. Altogether, these relations support the hypothesis that the observed variation in the K–Ar dates is an artifact and that this may reflect partial retention of mantle-derived (excess) argon. The effects of diffusive loss of 39Ar are discussed below.

The behavior of the argon isotopic system is influenced by the detailed cooling, vesiculation, and crystallization history and varies with parameters like melt and fluid composition, temperature and pressure (cf. Hayatsu & Waboso 1985). In addition, secondary effects include diffusive loss of 39Ar in the solidified rock, and such artifacts may be introduced during the irradiation in the reactor (D. Rex, pers. comm. 1992). The latter effect may be strongest in the most altered parts of the rocks (cf. Seidemann 1978).

The influence of cooling history on the excess argon retentivity is illustrated by studies of pillow basalts dredged from the ocean floor. Seidemann (1978) found a decrease in amount of excess argon from the rapidly cooled glassy rind of the pillow to the holocrystalline pillow center, reflecting that extensive degassing was related to the diffusive reorganization during the crystallization process. Thus retention of excess argon is favored by rapid chilling, and Dalrymple & Moore (1968) found that chilled glassy rims of pillow basalts, erupted at greater depths than 1500 m below sea level, commonly included excess argon. More recent studies have shown that some argon is soluble in basalt melt also at surface conditions and this may be preserved if the chilling is very rapid (cf. Marty et al. 1983; Hayatsu & Waboso 1985). In principle, the retentivity should depend on the ability to capture argon in sealed microvoids or in relatively open microstructures intergranular to the minerals or in rapidly grown minerals. An indication of the structural localization of the excess argon is given by the age spectra, but in general there is no simple interpretation. Ozima et al. (1977) concluded that excess radiogenic Ar was located in high temperature sites in deep-sea basalts. This would indicate that the excess argon was located in crystals or within the glass structure. In contrast, Seidemann (1978) found that excess 40Ar was released throughout the extraction range, implying that the excess argon was partitioned into both phenocrysts and glass, including the low melting parts. Seidemann (1978) concluded that the heating technique could not be used to eliminate the effects of excess radiogenic argon. A third possibility is demonstrated in an experimental study by Marty et al. (1983), who concluded that the excess argon was located in vesicles in the basalt. This was found to be released during the low-temperature incremental heating steps (700–800°C), while argon located in the homogeneous glass was released at higher temperatures (1000–1100°C).

The Ar-release behavior described by Marty et al. (1983) is similar to that described for sample 32.05 m in the present study, and a localization in microvoids or in weakly bound positions interstitial to quench crystals in the tachylite is possible. The different Ar-release patterns of samples 32.05 m and 77.70 m may be explained by the textural differences of these samples, which indicate slightly different cooling histories. Sample 32.05 m, which revealed excess argon in the first incremental heating step, was characterized by rapid cooling and a high nucleation rate; but with limited time for growth of phenocrysts during magma ascent. This is evidenced by the very minute needles of plagioclase quench crystals and typically, the vesicles are sealed by crystal-free glassy rims. Locations for the excess argon would be in the sealed matrix parts, or possibly within sealed microvoids. By contrast, the somewhat coarser and differentiated microphenocrysts of plagioclase, clinopyroxene, ilmenite, and possibly resorbed/re-equilibrated olivine of sample 77.70 m show that the crystallization took place more slowly at an earlier stage, with time for diffusive loss of excess argon. This was followed by a later stage of very rapid chilling with production of homogeneous glass matrix as the melt came into contact with water or ice. Thus excess argon may have had time to escape before the final event with rapid chilling. If argon was captured in vesicles, it could have escaped through perlitic microcracks during or shortly after the final cooling.

Conclusions

The present study provides evidence of basaltic volcanism on the Western Barents Margin in Late Pliocene time. The volcanic eruption occurred at the continental side of the continent–ocean transition zone, possibly in the vicinity of the Stappen High. The eruption was mainly subaquatic and possibly also included subaerial phases. The magma was extruded through (and into) an area of unstable deposits of poorly consolidated Tertiary and older sediments. A volcanic pile may have accumulated at the eruption center. The volcanism was accompanied by tectonic activity which may have induced sliding of older sediments and the newly deposited...
eruption products leading to sediment reworking by various mass-flow and debris-flow mechanisms. Time equivalence to intrusive structures in the Vestbakken Volcanic Province, and possibly to part of the thermal activity on Spitsbergen, indicates that this activity may be part of a regional thermal and volcanic event on the Barents Shelf.

Before the eruption, olivine (+Cr-spinel) had crystallized from a fairly primitive basaltic magma composition. The magma was produced either by melting of an anomalously enriched lithospheric mantle, or the composition of the magma was modified and homogenized by interaction with enriched mantle or more probably by continental lower crustal material. The melt generation was at shallow depths (ca. 30 km) and magma ascended through continental lithosphere.

The western margin of the Barents Shelf was subjected to rifting and formation of sheared margin segments in Early Cenozoic time (cf. Eldholm et al. 1987). Shallow seismic data indicate shear faulting in the Stappen area also in Late Cenozoic time (Sættem et al. 1994). The latter could be related to the Late Cenozoic igneous activity. The tectonics at the boundary between the oceanic lithosphere and continental lithosphere is, however, not properly known.

The extensive uplift at the Western Barents Shelf margin has been discussed in several papers (cf. Vågnes & Amundsen 1993; Sættem et al. 1994), and implications for the timing of erosion from the present study are discussed elsewhere (Sættem et al. 1994). Vågnes & Amundsen (1993) related the extensive uplift to a thinning of the continental lithospheric mantle at the continent-ocean transition zone. If this model is correct, the volcanism could have resulted from melting of the lower lithosphere related to the thinning. An alternative hypothesis is that part of the uplift was related to underplating of spreading-related basaltic magma below the Western Barents Shelf margin accompanied by local melting of the enriched continental lithosphere.

Acknowledgements. – The study of the volcanics from southwest of Bjrnaya resulted as part of IKU Shallow Drilling 1989, and most of the analytical work was funded as an integrated part of the basic core descriptions. Discussions with colleagues at IKU during the project work (see authors included in Bugge et al. 1990) and during later follow-up studies, mainly with Joar Sættem, and input of seismic data by Stein Fanavoll are acknowledged. Project support and permission to publish by the participating oil companies (Amoco Norway Oil Company, A/S Norske Shell, Congocho Norway Inc., Elf Aquitaine Norge A/S, Mobil Exploration Norway Inc., Norsk Hydro a.s., Oljedirektoratet, Saga Petroleum a.s., Statoil, Total Marine Norsk A/S), are also acknowledged.

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