ICAM IV

Dartmouth, Nova Scotia, Canada

September 30-October 3, 2003

Proceedings of the Fourth International Conference on Arctic Margins

U.S. Department of the Interior
Minerals Management Service
Alaska Outer Continental Shelf Region
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Edited by:

Robert A. Scott
Dennis K. Thurston
INTRODUCTION

The International Conference on Arctic Margins (ICAM) was founded by the U.S. Department of the Interior Minerals Management Service (MMS) in 1991 with the underlying two-point theme of 1) scientific understanding of the Arctic and 2) international cooperation in Arctic research. To these ends, ICAM has provided a forum for the exchange of information, collaboration in research, and presentation of results.

The inaugural 1992 ICAM conference was held in Anchorage, Alaska. It was hosted by the MMS, Alaska Geological Society, and the University of Alaska, Fairbanks. Nearly four hundred scientists from 12 nations attended the Anchorage meeting and made nearly two hundred technical presentations on Arctic earth sciences. The Proceedings of the 1992 ICAM contain 70 scientific papers and are available from the Department of Interior Minerals Management Service, Alaska Region.

The 1994 ICAM in Magadan, Russia was hosted by the North East Science Center of the Russian Academy of Sciences Far East Branch and was attended by over 130 scientists. The conference featured sessions on Stratigraphy, Paleoclimate and Paleogeography, Regional Terrane Correlation and Geophysics, Resource Potential, Permafrost, and Mining Ecology. A round table discussion was held on Present and Future Cooperative Alliances Between Science, Industry, and Government. There were numerous symposia and selected workshops on specific topics. The Proceedings of the 1994 ICAM were published in 1995 by the Russian Academy of Sciences Far Eastern Branch, Magadan and contain 45 scientific papers.

Germany, through the Federal Institute for Geosciences and Natural Resources (BGR), the German Society for Polar Research and the Alfred Wegener Institute for Polar and Marine Research hosted the 1998 ICAM III conference in Celle Germany. The ICAM III Proceedings contain sixty seven papers and were published as two special volumes of the Journal Polarforschung (numbers 68 and 69) in 2000 and 2001 by the German Society for Polar Research and Alfred Wegener Institute for Polar and Marine Research.

ICAM IV convened September 30-October 3, 2003, and was sponsored by Natural Resources Canada through the Geological Survey of Canada (Atlantic) and was supposed to be held at the Bedford Institute of Oceanography, Dartmouth Nova Scotia. However, due to the impact of the direct hit of Hurricane Juan at Halifax the day before the meeting, the resourceful and undaunted organizers led by Ruth Jackson of the Geological Survey of Canada, found space at a local hotel to hold the meeting. And not even a Hurricane could stop the meeting. This fourth conference was devoted to discussions of the geology, geodynamics and resources of the margins within our joint Arctic community.

Just in time for the International Polar Year, ICAM V – the Fifth International Conference on Arctic Margins will be held September 3-5, 2007 in Tromsø, Norway. It is sponsored by the Geological Society of Norway (NGF) in cooperation with the European Association of Geoscientists and Engineers (EAGE).

Conference themes include:

**Geodynamics and Tectonic Evolution of Arctic Region**
Search for past plate boundaries; Megasequences in the Circum Arctic; Magmatism; Palaeogeography

**Circum Arctic Orogens**
Uralides; Caledonides/Ellesmerian; Brooks Range

**Sedimentary Basin Development**

**Uplift and Erosion of Arctic Margins – Timing and Causes**

**Continental Slopes and Deep Basins**
Sedimentary processes; Origin and evolution of the Alpha-Mendeleev Ridges

**Arctic Gateways**
Tectonics; Palaeo-oceanography
Glacial History and Processes
Morphological development; Palaeoclimate; Permafrost; Technologies for Arctic Ocean Research

For the papers contained in the ICAM IV Proceedings, we strived to make sure (perhaps not completely successfully) that all of the reviewers comments were addressed by the authors and to make sure that nothing was altered or lost in the editing process. We have also strived to ensure all grammatical and spelling errors were caught and corrected and that no other errors were introduced or overlooked. However, we realize that some may have been missed and request that authors or readers report these to us and we will post them on the web at www.mms.gov/alaska/icam.

This volume of the Proceedings was produced by the volunteer efforts of many people and the editors would like to thank all who submitted papers and who patiently worked with the editors in finally achieving publication. We also thank all of our reviewers for their thoughtful and thorough work. We feel that the papers contained in the Proceedings will be valuable contributions to the Arctic Earth Science Community for years to come.

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The Arctic Ocean is a small ocean with perennial ice cover. The ocean and the surrounding landmasses are challenging regions to collect geological and geophysical data. Thus, Arctic researchers are united in their difficult working conditions. Furthermore, they all understand the importance of correlating onshore data around the Arctic Ocean and the need for onshore-offshore connections to understand the geological history.

At each of the ICAM meetings a number of topics are highlighted. One new theme for the ICAM IV was Science issues relating to United Nations Convention on Law of the Sea (UNCLOS) Article 76. Five nations (Russia, Norway, Denmark, Canada and the United States) have the potential of extending their continental shelves into the Arctic Ocean. New data collected to make the claim for additional territory will improve the scientific knowledge base of the Arctic Ocean.

A number of sessions were organized to have broad participation either by region or by scientific discipline. The Geodynamic Significance of Arctic Magmatism was a circum-Arctic theme put forward to give scientists working in geographically separated areas the opportunity to compare and contrast their findings. Vertical motions in the Arctic: Tectonic and glacial brought together a number of different disciplines with related problems. Atlantic versus Pacific tectonic regimes in the evolution of the Arctic Ocean was featured so that the Arctic Ocean plate tectonics could be set in a global prospective. The Tectonic Framework session provided an overview of Arctic specific tectonic issues. Circum-Greenland Tectonics was another unifying theme. The well studied seas adjacent Norway and east Greenland provide standards to be emulated and insights into the more difficult to access regions. New information collected on Nares Strait 2001 marine expedition could be highlighted and placed in the context of other activities in the Labrador Sea. Hydrocarbon potential and gas hydrates are of continuing interest in the Arctic and it was important to be updated on this subject.

Another new theme for ICAM conferences was Arctic Margins: Coastal and Marine Environmental Geoscience in a Changing Climate; Implications for Development. This session allowed the important geological information on changing climate and the effect on the Arctic population to be examined.

During the meeting, at the start of each day, a key note address was held. Three topics were chosen that were to both have overall appeal and to provide specific knowledge on the activities in the country where the meeting was held. Two of the presentations that addressed Canadian issues of broad interest were: Oil and Gas Potential of Canada's Northern Basins: Future Exploration Trends by Benoit Beauchamp and Implications of a 500 metre deep borehole in frozen sediments in the Canadian Beaufort Shelf presented by Steve Blasco. The data gleaned from the scientific program on US submarines called SCICEX was presented by Bernard Coakley in a talk called The Lomonosov Ridge, Top to Bottom.

At 12:10 a.m. Monday September 29, 2003, Hurricane Juan made landfall in Nova Scotia as one of the most powerful and damaging hurricanes to ever affect Canada. Hurricane Juan's maximum sustained wind speed at landfall in Nova Scotia was measured at 100 mph (160 km/h). The majority of severe property damage was concentrated in the core of the Halifax Regional Municipality where ICAM IV was held. Power lines were down, bridges were closed and roads were blocked by fallen trees. The arrival of Hurricane Juan on the day prior to the meeting (September 30 - October 3, 2003) altered all plans that had been initiated at least a year in advance.

Bedford Institute of Oceanography where the meeting was to be held was without power as well as most of the hotels in the Halifax-Dartmouth area. Claudia Currie head of the organizing committee noticed that the Ramada Hotel had power. Space for a day was booked until the power was restored at BIO. (It actually was not restored until the meeting was over). Meanwhile a low tech solution was devised to let participants know where the meeting was to be held. Volunteers walked or drove to the hotels in the region and put up hand written notices of the meeting location. In all, twenty volunteers from the Geological Survey of
Canada and Fisheries and Oceans, leapt into action. Nelly Koziel managed the additional costs that were not anticipated. With no commercial space available for a reception, Ron Macnab volunteered his home for a wine and cheese party for the 130 participants. The lobster supper was the supreme accomplishment of the volunteers who managed to find a local hall that had power, a supplier for the lobsters and a caterer to provide the rest. The field trip went as planned in unusually warm and calm conditions after the storm. The Fourth International Conference on Arctic margins closed with a successful outcome and quite a story to tell.

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Ressources naturelles Canada
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CONTENTS

Tribute to Nikita A. Bogdanov  
Viktor Khain

Tribute to Kirill V. Simakov  
Dennis Thurston

Evidence of Caledonian Orogeny in the Silurian - Devonian Successions of the Barents and Kara Shelves  
David Gee & Olga Bogolepova

Early Palaeozoic Unconformity on Severnaya Zemlya and Relationships to the Timanian Margin of Baltica  
Henning Lorenz, David G. Gee and Olga K. Bogolepova

Neoproterozoic Island Arc Magmatism of Northern Taimyr  
Vicky Pease and S. Persson

The Tremadocian Monoplacophoran Mollusc Kirengella from the Pechora Basin of Northwest Russia  
Alexander P. Gubanov and Olga K. Bogolepova

Structural, Thermal and Rheological Control of the Late Palaeozoic Basins in East Greenland  
Ebbe H. Hartz, S.N. Kristiansen, A. Calvert, K.V. Hodges, and M. Heeremans

The Nares Strait Debate: Implications from the Structural Evolution of Tertiary Outliers in Eastern Ellesmere Island  
Kerstin Saalmann

Structure of the Russian Eastern Arctic Shelf  
V.A. Vinogradov, E.A. Gusev, and B.G. Lopatin

Three-Dimensional Structural Model of the Southeast Coastal Shelf of the East Siberian Sea and the Role of Processes of Basification in their Formation (A New Gravity Interpretation)  
Yury Vashchilov, Vladimir Glotov, Tamara Zimnikova, Viktor Lyubomudrov, Olga Sakhno, and Irina Tsygankova

Global Tectonic Actions Emanating from Arctic Opening in the Circumstances of a Two-Layer Mantle and a Thick-Plate Paradigm Involving Deep Cratonic Tectospheres: The Eurekan (Eocene) Compressive Motion of Greenland and Other Examples  
Miles F. Osmaston

Regional Paleotectonic Interpretation of Seismic Data from the Deep-Water Central Arctic  
Victor V. Butsenko and Victor A. Poselov

A Seismic Model of The Earth's Crust Along The "East-Siberian Continental Margin – Podvodnikov Basin – Arlis Rise" Geotraverse (Arctic Ocean)  
Nina N. Lebedeva-Ivanova, Yury Ya. Zamansky, Aldona Y. Langinen, and Michael B. Sergeyev

Seismic Data Acquisition in the Nansen Basin, Arctic Ocean  
Øyvind Engen, Jakob Andreas Gjengedal, Olav Eldholm, and Yngve Kristoffersen
Is Grounding of an Ice Shelf Possible in the Central Arctic Ocean? A Modeling Experiment
Martin Jakobsson, Martin J. Siegert, and Mark Paton

Sedimentary Thickness Estimations from Magnetic Data in the Nansen Basin
Vladimir Glebovsky, Anton Likhachev, Yngve Kristoffersen, Øyvind Engen, Jan Inge Faleide, and Harald Brekke

Results of Density Modeling of the Major Structural Elements of the Arctic Ocean
Ekaterina Astafurova, Vladimir Glebovsky, and Vladimir Fedorov

Geological Origin of the Magnetic Anomaly Field in the Central Amerasian Basin (Arctic Ocean)
Valentina V. Verba and Vladimir I. Fedorov

Velocity Structure and Correlation of the Sedimentary Cover on the Lomonosov Ridge and in the Amerasian Basin, Arctic Ocean

Developing Outer Continental Shelf Limits in the Arctic Ocean: Geoscience Encounters UNCLOS Article 76
Ron Macnab

Experience in Applying the Geological Criteria of Article 76 to the Definition of the Outer Limit of the Extended Continental Shelf of the Russian Federation in the Arctic Ocean
Victor A. Poselov, Valery D. Kaminsky, Rinat R. Murzin, Victor V. Butsenko, and Anatoly A. Komaritsyn

40Ar-39Ar Dating of Mafic Magmatism from the Sverdrup Basin Magmatic Province
Mike Villeneuve and Marie-Claude Williamson

The Early Cretaceous Arctic LIP: Its Geodynamic Setting and Implications for Canada Basin Opening
Sergey Drachev and Andy Saunders

Correlation of Aeromagnetic Signatures and Volcanic Rocks Over Northern Greenland and the Adjacent Lincoln Sea
Detlef Damaske and Solveig Estrada

Basic Rocks of Franz Josef Land: Chemical Character and Tectonic Setting
Alexander N. Evdokimov, Nicolay M. Stolbov

Effects of Developing Iceland Plume on North Atlantic and Arctic Sedimentary Basins
Laura M. Mackay, Stephen M. Jones, Nicky White, and Robert Scott

Spatial Patterns in Circum-Arctic Coastal Storms Derived from Observed Wind Speed Data, 1950 - 2000
David E. Atkinson and Steven M. Solomon

Neutral Buoyancy Icebergs in Kane Basin between Arctic Canada and Greenland a Threat to Northern Navigation: Identifying the Source and Possible Links with Arctic Warming
Marcos Zentilli, J. Christopher Harrison, and Jonathan Crealock

The Influence of Regional Warming on the Treeline of a Subarctic Mountain Range – A First Approach to Field Research
Bernd Cyffka and Michael Zierdt
TRIBUTE TO NIKITA A. BOGDANOV
By Viktor Khain

Nikita Alexeevich Bogdanov, a prominent Russian geologist, active Arctic researcher, passed away on 14 December 2003 after serious disease. This grievous event took place only three months after his return from the ICAM IV Conference in Halifax, Canada. He felt back pain, and soon it turned out that he was affected by melanoma, the most untreatable and fulminant form of cancer.

Nikita Bogdanov was Director of the Institute of Internal and Marginal seas, Corresponding Member of Russian Academy of Sciences and Professor of Lomonosov Moscow State University. Nikita Bogdanov was born on 23 July 1931 in the family of outstanding Russian geologist Alexey A. Bogdanov, professor and dean of Geological Faculty of Lomonosov Moscow State University. Nikita Bogdanov had started his field works in Eastern Siberia, at that time poorly studied area in the Amur River basin, even before he graduated from Moscow Geological Prospecting Institute. Later on, already as a research fellow of Geological Institute of Russian Academy of Sciences, he extended his research interest to the Koryak Upland in the North-East of Russia and Wrangel Island in the Arctic. Between these fieldworks he managed to make long business trips to Australia and California.

His integrating studies of North-East of Russia, Tasmanian Foldbelt of Australia and North American Cordilleras resulted in Doctoral thesis devoted to Paleozoic geosynclines of Pacific Ring. In this work he introduced an idea of “talasso-geosynclines” initiated on the oceanic crust. With the advent of conception considering ophiolites as relics of ancient oceanic crust Nikita Bogdanov was among the first Russian geologists who, together with Academician V. Peive, began to apply this conception to folded structures of the USSR. Soon he became a Head of International Project on ophiolites studies and organized international geological excursions to Polar Urals, Middle Asia and Transcaucasia. These excursions allowed geologists from many countries to get acquaintance with ophiolites located within the territory of USSR, and one should take into account that to organize the excursions Nikita Bogdanov had to overcome the constraints on the visits of foreigners existed that time in USSR.

Nikita Bogdanov’s activity as General Secretary of Organizing Committee of 24 International Geological Congress in Moscow in 1984 brightly exhibited his managerial abilities. As participants of the Congress acknowledged, this Session was among the most successful in the history of Geological Congresses. Since that time the name of Nikita Bogdanov became widely known among the geologists from many countries. In 1979 Nikita Bogdanov was invited by Academician A. Sidorenko to take a position of Vice-Director of newly founded Institute of Lithosphere of Russian Academy of Sciences. In 1989 he was elected Director of the Institute. Nikita Bogdanov with his colleagues continued studies of geology of the North-East Russia, particularly, Kamchatka and the Arctic region. He was also a leader or active participant of the cruises carried out onboard research vessels in the Phillipine, Mediterranean and Chukchi seas. In 1992, together with S. Tilman, he published a first detailed tectonic map of the whole area of North-East of Russia. It would be no exaggeration to say that he was the best expert in geology, particularly, tectonics of this remote area of Russia. Then he initiated compilation of tectonic maps of the marginal and internal seas of Russia. Tectonic maps of the Mediterranean, Barents, Kara, Laptev, Caspian seas and Sea of Okhotsk of 1: 2 500 000 scale were published under his editorship. In 2003 this work was awarded by State Prize of RF.
During his last years he focused his attention on Arctic tectonics issues and before his unexpected death he nearly finished compilation of the entire Arctic Basin tectonic map which was intended to be demonstrated during regular Session of the International Geological Congress in Florence, but these plans could not come true. For the numerous Nikita’s friends in Russia and abroad, as well as for both Russian and international geological communities his untimely death is a heavy and irreplaceable loss.

TRIBUTE TO KIRILL V. SIMAKOV
By Dennis Thurston

Kirill Vladimirovich Simakov – Full member of the RAS, Doctor of Sciences (Geology and Mineralogy), Member of Honor of Science of the Russian Federation, Chairman of the Presidium of the North-Eastern Scientific Center of the FEB RAS (Magadan), died April 14, 2004. The North-Eastern Scientific Center hosted the second ICAM 1994 in Magadan, of which Kirill was the Chairman and organizer. He co-edited and printed the ICAM II Proceedings. He was on the organizing committee of the first ICAM meeting in Anchorage in 1992 and the third ICAM meeting in Celle, Germany in 1998. He was not able to attend to the ICAM IV meeting in Halifax.

Kirill was born in Leningrad on February 1, 1935. He was married and leaves his wife Valentina, two children and a granddaughter.

The son of a geologist, Kirill studied from 1952-57 at the Geological Department of the Leningrad State University. From 1957-70 he worked as a head of field teams and as a senior geologist of the Seimchan Geological-Exploration Team in Magadan Region. Since 1970 he worked as a scientist, a laboratory head, a senior scientist of NEISRI, Kirill defended his Candidate’s dissertation in 1970.

From 1979-84 K.V. Simakov headed a Soviet-Belgian joint investigation on a comparison of Famennian and Tournaisian reference sections of the Russian Northeast and the Franco-Belgian Basin. He is a participant in and organizer of different international geologic symposia and international geologic congresses, carboniferous stratigraphy and geology congresses, the Pacific Scientific Association Congress and others. Kirill received his Doctorate in 1985. In 1990 he became the Vice-Chairman of the NESC RAS. He was also elected a Corresponding Member of RAS in 1990 and received the title of Honored Scientist of the Russian Federation in 1991.

At the meeting of the Russian Academy of Sciences Presidium in 2000, Kirill was elected as a Full member (Academician) of the RAS.

Kirill has authored over 250 papers, including 20 monographs. He was awarded the highest award in geology of the USSR Academy of Sciences by Academician Karpinsky in 1988 for a series of his monographs published during 1984-86 which dealt with the problem of determination of chronostratigraphic boundaries and generalized worldwide knowledge on the stratigraphy of Devonian-Carboniferous boundary deposits.

Kirill had a special interest in northeastern Russian regional geology and the biostratigraphy of Devonian-Carboniferous boundary deposits. His lifetime work led him to investigate the theory of chronostratigraphy and philosophical aspects of geologic time. One of his last publications in 2001—“Origin, Development, and Perspectives of the Theory of Paleobiospheric Time”—was a book dedicated to Academician Alexander Leonidovich Yanshin for his 90-year jubilee. This book delved into Kirill’s favorite subject—the conceptual history of real
paleobiospheric time, and how changes in geological
theory and empirical data impacted the perception and
application of time in the earth sciences.

Kirill was a member of the Russian
Interdepartmental Stratigraphic Committee,
Chairman of the Russian Northeast Regional
Interdepartmental Stratigraphic Commission, a
member of several scientific councils at the RAS and
FEB RAS, and a member of the editorial board of
Pacific Geology magazine. He became a member of
the International Working Group on the Devonian-
Carboniferous Boundary in 1976, and its Vice-
Chairman in 1984; since 1992 he was a member of
the American Association of Petroleum Geologists.

Kirill liked to ski and to read detective stories,
but most of all he liked to work. According to his
colleagues he was working on his last day. Scientists
of his caliber and type are very rare and he and his
ideas are greatly missed.
EVIDENCE OF CALEDONIAN OROGENY IN THE SILURIAN – DEVONIAN SUCCESSIONS OF THE EASTERN BARENTS AND KARA SHELVES

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ABSTRACT

The Caledonide Orogen of western Scandinavia continues northwards beneath the Barents Sea to northern Eurasia's shelf-edge. Old Red Sandstone (ORS) basins occur within the hinterland of the orogen on Svalbard, in East Greenland, in the Scandes, and further south, within the North Atlantic Caledonides. On Franz Josef Land, the Early and Mid Palaeozoic successions are absent beneath an Early Carboniferous unconformity, and the deformation of underlying greenschist facies Neoproterozoic turbidites has been reported to be Caledonian in age.

Evidence of foreland basins is present in the Oslo graben and their existence is inferred further north in Scandinavia, but is less well defined. In areas to the southeast and east of the Barents Sea and northernmost Kara Sea, Devonian non-marine siliciclastic successions are present which appear to be a sedimentary response to mid Palaeozoic orogeny further west.

In northern Timan, Late Silurian (Pridolian) limestones and underlying Timanide complexes are unconformably overlain by Early Devonian (Lochkovian and Pragian) red sandstones and then Frasnian sandstones and basalts. 80 km to the northeast, on Kolguev Island, Lochkovian-Pragian (in the east) to Frasnian conglomerates, sandstones and siltstones rest, with minor angular unconformity, on Middle Ordovician siltstones and sandstones; the red beds are thickest (up to 1.5 km) in the northern part of the island, where Frasnian basalts are also reported.

On Novaya Zemlya, in eastern areas, Silurian and Devonian successions are dominated by limestones and shales. Towards the west and locally along the western coast of archipelago, the Frasnian is largely composed of thick (up to 3 km), non-marine conglomerates and sandstones with volcanics and subordinate dolomites. Coarse clastics appear first in the shales in the Wenlock and give way to an ORS facies in the Early Devonian.

On Severnaya Zemlya, thick Llandovery and Wenlock limestones pass up into the variegated marls and fine sandstones of the uppermost Pridolian, which, in turn, are overlain, with a minor stratigraphic gap, by Early Devonian (Lochkovian) sandstones. This Devonian ORS facies, up to 2 km thick, includes red and variegated sandstones, siltstones, marls and dolomites.

Thus, during Late Silurian to Devonian times, these areas of the eastern Barents Shelf, from the Kanin Peninsula to northwest Novaya Zemlya and eastwards to the Severnaya Zemlya Archipelago on the northern Kara Shelf, are characterized by regression and influx of clastic red-bed sedimentation. The successions thicken and coarsen to the west. Detrital minerals in these siliciclastic units are dominated by zircon, garnet and tourmaline, indicating the presence of metamorphic rocks and granitoids in the source areas. It is concluded that the Scandinavian Caledonides continued northeastwards into the Barents Shelf as the Barentsian Caledonides (including Svalbard and Franz Josef Land), the deformation front being located between Franz Josef Land and Novaya Zemlya.

INTRODUCTION

The Caledonian Mountains of Scandinavia strike northeastwards from northern Norway into the Barents Sea. The path of this Early-Mid Palaeozoic orogenic belt across the Barents Sea, beneath a cover of late Palaeozoic and younger sediments, is not easily defined and is therefore a subject of considerable controversy. Evidence of Caledonian deformation and metamorphism on Svalbard is widespread (Harland, 1997), overlie by Devonian Old Red Sandstones (ORS) (Friend and Moody-Stuart, 1972 and Friend and Williams, 2000); continuity into this northwestern part of the Barents Shelf is unambiguous. Further east there is no exposed pre-Mesozoic bedrock for a thousand kilometres until Novaya Zemlya, where the Early – Mid Palaeozoic succession is interrupted by minor unconformities, but not disturbed by Caledonian orogeny (Korago et al., 1992). In between, on Franz Josef Land, a single drill-hole, Nagurskaya on Alexandra Land (Dibner, 1998), provides the only direct evidence available on the character of pre-Carboniferous basement of the northern Barents Sea. Here, the pre-Carboniferous strata are metamorphosed in low greenschist facies and are represented by folded metaturbidites and phyllites. The latter are reported to contain Vendian acritarchs and have yielded a c. 360 Ma K/Ar whole rock age (Gramberg 1988, Dibner 1998).
Until recently, two hypotheses dominated interpretations of the Caledonides in the regions north of Scandinavia (Fig. 1). One of these suggested that the Caledonide Orogen influenced only the western part of the Barents Sea (Ziegler, 1990 and Nikishin et al., 1996) swinging northwards from Norway to Svalbard and then northwestwards into the North Greenland Foldbelt and the Ellesmerian Orogen.

The second hypothesis suggested that, as the Caledonides reach northwards from Scandinavia, they bifurcate into a western branch (see above) and an eastern branch, the latter existing as a deformation belt in the Eastern Barents Sea, between Novaya Zemlya and Franz Josef Land (Siedlecka, 1975; Doré, 1991; and Gudlaugsson et al., 1998). This hypothesis (Fig. 1) goes back into the Russian literature of the 1960’s, and has been favoured by many authors (Atlasov, 1964; Gafarov, 1966; Emelyanov et al., 1971; and Zhuravlev, 1972) who referred to this inferred eastern branch of the orogen as “the Norwegian – Barentsian Caledonides” or as the “Scandinavian – Severnaya Zemlya Caledonides” by Egiazarov et al. (1972, 1977), based on the distribution of non-marine clastic (ORS) successions.

The bifurcation hypothesis requires the existence of old cratonic continental crust in the northern Barents Sea, separating the Svalbard Caledonides from the eastern Barents Sea Caledonides. This (micro?)craton has been referred to as the Barents Craton (Egiazarov et al. 1972, Siedlecka 1979, Harland 1997) or Barentsia (Arkhangelsky and Shatsky, 1933; Stille, 1958; Ziegler et al., 1977; Gee, 1996; Harland, 1997; and Sharov, 2000).

A third hypothesis (Fig. 2) favours the existence of a broad continuation of the Scandinavian Caledonides north-northeastwards across the Barents Sea, with the Caledonian front c 50 - 100 km to the east of Novaya Zemlya, and Caledonian suture(s) separating Svalbard from Franz Josef Land (Gee, 2001; 2004). This hypothesis of a Barentsian Caledonides is based on three lines of evidence. Firstly, that the grade of Caledonian metamorphism and deformation on Nordaustlandet increases eastwards with extensive migmatisation as far east as Kvitoya (Teben'kov et al., 2002; Johansson et al., 2004; and Gee and Teben'kov, 2004). Secondly, that the Neoproterozoic to Ordovician successions and faunas of Nordaustlandet and Ny Friesland are unambiguously of Laurentian affinities (Swett, 1972 and Fortey and Bruton, 1973), and

Figure 1. Hypothesis for the Caledonides of the European Arctic (Late Paleozoic framework) showing the northwestern and northeastern branches of the orogen, and the inferred location of the Barents Craton (Barentsia) and the Caledonian suture (based on Gudlaugsson et al. 1998, Breivik et al. 2002)

Figure 2. The Barentsian Caledonides: A broad NNE trending orogen beneath the Barents Shelf (latest Mesozoic framework)
therefore separated from Baltica by a suture(s) located further east. And thirdly, that Svalbard’s Mesoproterozoic to earliest Neoproterozoic successions and their metamorphism and intrusions are similar in character and age to analogues in the East Greenland Caledonides (Gee and Teben’kov, 2004). That the deformation and metamorphism of the basement beneath Franz Josef Land (Dibner, 1998) is of Caledonian age is also compatible with this hypothesis.

Various lines of geophysical evidence influence these hypotheses. The aeromagnetic data (Skilbrei, 1991) indicate that the N-trending anomalies on Svalbard and the neighboring shelf are truncated to the east by a NNE-trending zone of discordance; further east and south in the Barents Sea, characteristic Timanian magnetic trends prevail. Satellite gravity data (Fichler and Sæther, 1995) indicate that the characteristic NW-trending anomalies of the Timanide Orogen extend northwards and westwards through the eastern Barents Sea to a line reaching from northernmost Norway towards Franz Josef Land. Deep seismic reflection and refraction surveys (Gudlaugsson et al., 1987; 1998) have imaged major NE-trending structural discontinuities, which were thought to be related to Caledonian basement, reactivated during late Palaeozoic and Mesozoic rifting. Recently, Breivik et al. (2002) have inferred that an important deep structural discontinuity extends north-northeastwards through the Sentralbanken High (34°E, 76°N), which they interpret to be a suture-zone. All these authors, and Roberts and Olovyanishnikov (2004), appear to favour the bifurcation model (Fig. 1), but the geophysical data are compatible with both this and the “Barentsian Caledonides” alternative, favoured here.

This paper explores the stratigraphic and sedimentological evidence for the presence of a Caledonian fold belt located northwest of northern Timan, Kolguev Island, Novaya Zemlya and Severnaya Zemlya, close enough to influence the depositional environment of the Late Silurian and Devonian strata. None of these areas are directly influenced by mid Palaeozoic orogeny. The southern regions are strongly influenced by Timanian (Vendian) Orogeny, with pre-Ordovician (locally pre-Late Cambrian) major unconformities (Bogolepova and Gee, 2004). Northeasternmost parts (e.g. October Revolution Island of Severnaya Zemlya) also provide evidence of latest Cambrian and/or earliest Ordovician deformation and low grade metamorphism (Proskurnin, 1999 and Lorenz et al., this volume); the presence of an underlying Timanian complex has been inferred (Lorenz et al., in press).

**Figure 3.** Localities with “foreland-basin” Old Red sandstones and the Late Silurian and Devonian stratigraphy across the northern Eurasian margin—Timan to Severnaya Zemlya.
TIMAN - PECHORA REGION

The Timan – Pechora region was a passive continental margin for much of the Palaeozoic era, with episodes of Ordovician and Devonian rifting (Dedeev et al., 1996). Palaeozoic Baltica extended to the west-southwest (present coordinates), and the Uralian Ocean to the east. The Early Palaeozoic basin fill changed gradually from initial siliciclastic to mainly carbonate-dominated sediments (Bogolepova and Gee, 2004).

A Devonian, non-marine siliciclastic ORS facies is present in northern and eastern parts of the Timan-Pechora Region (Fig. 3). It rests with slight angular unconformity on Silurian carbonates or older Palaeozoic strata; locally it was deposited directly on the Timanian basement.

Devonian ORS are well exposed in northern Timan and on Kanin Peninsula, where they rest on Timanian basement and, locally, on Silurian limestones, dolomites and sandstones (Fig. 3). Basal conglomerates pass up into Middle Devonian (Eifelian), red cross-bedded sandstones and Late Devonian (Frasnian) basalts and sandstones (Larionova and Bogatsky, 2002). Further southeast, along the Timan Range, the red beds give way to a Devonian carbonate facies.

In the Middle Devonian, rifting occurred in the northwestern part of the Pechora Basin, accompanied by basalt volcanism and the appearance of siliciclastics, derived locally from the west. Further east, a reef system developed bordering an internal basin where a

Figure 4. The reconstructed directions of clastic sediment provenance in Old Red Sandstone successions of the Timan-Severnaya Zemlya area (based on Gramberg, 1988 and Kurshs, 1982)
The Early Palaeozoic strata (Bondarev, 1964; 1970) on Novaya Zemlya were deposited along the northeastern (present co-ordinates) margin of Baltica (Nikishin et al., 1996 and Cocks, 2000). In general, Middle Ordovician to Silurian carbonate platform successions dominate the southwestern parts, though Wenlock-Pridolian pink polymict coarse gravels, quartz conglomerates and arkosic cross-bedded sandstones (1500 m), with subordinate shales, occur in northwesternmost parts (Bondarev, 1964; 1970). Pebbles are composed mainly of acid effusives (Andreeva in Gramberg, 1988), and a source area, supplying these clastics, has been reconstructed to the north (Gramberg, 1967). Similar lithologies occur in the Ludlow-Pridolian successions of the central areas, and in the Pridolian successions of the southwestern part of Novaya Zemlya (Nekhorosheva, 1981). In northeastern Novaya Zemlya, the Early Palaeozoic successions are dominated by turbidites.

The Early Devonian strata (up to 2500 m) are made up generally of limestones and dolomites; siliciclastics, up to a few hundreds of meters thick, are also present, locally, resting with minor stratigraphic gap on the underlying Silurian limestones and dolomites (Cherkesova et al., 1988). The Middle Devonian (700-900 m) is composed of limestones with dolomites (Cherkesova et al., 1988). The Middle Devonian strata (Eifelian) of central Novaya Zemlya (Sobolev 1982). An angular unconformity is recorded between the Middle Devonian (Eifelian) and the overlying Late Devonian (Frasnian) red polymict coarse gravels, conglomerates, quartz cross-bedded sandstones, tuffs and basalts. They pass up into the Famennian limestones and dolomites. The whole Late Devonian succession is up to 1000 m thick. Khain (2001)
compared this non-marine facies with the typical ORS of the Scandinavian Caledonides.

Studies of the mineralogical composition of the Palaeozoic siliciclastic rocks of Novaya Zemlya allowed Andreeva (1984) to distinguish a separate province, where a uniform tourmaline-zircon-ilmenite association existed through Late Silurian and Early Devonian times. A source for the detrital material was inferred to be located in the west and north (Fig. 4), in the Barents Sea, where metamorphic rocks and granites were apparently subject to deep erosion.

SEVERNAYA ZEMLYA

The Severnaya Zemlya Archipelago lies in the northeastern part of the Kara Sea, on the edge of the Eurasian continental shelf (Fig. 3). Together with the northern part of the Taimyr Peninsula, the Palaeozoic rocks have been thought to represent a separate microcontinent – the Kara Plate (Bogdanov et al., 1998). We prefer to refer to this fragment of continental crust as the North Kara Terrane, this being the area in which it exists (i.e. not the Kara Shelf as a whole).

Reconstruction of the early Palaeozoic history of this terrane is still problematic. Based on faunal data, the North Kara Terrane could have been a part of, or close to Baltica at least by until end of the Late Cambrian to earliest Ordovician (Bogolepova et al., 2001). Collision with Siberia is thought to have occurred in Late Carboniferous – Permian times (Vernikovsky, 1996).

Silurian and Devonian strata are widely distributed on the October Revolution, Pioneer and Komsomolets islands of the Severnaya Zemlya Archipelago. In general, the Silurian strata are represented by shallow water carbonates overlain by continental red siliciclastics of Devonian age (Fig. 3). On October Revolution Island, the basal beds of the Late Silurian (Pridolian) Krasnaya Bukhta Formation are red sandstones resting on Ludlovian limestones. The whole Pridolian succession (700 m in its type area to the northwest) is represented by red clayey-sandy deposits with thin interlayers of conglomerates; this unit decreases in thickness towards the south (100 m), where the beds are overlain, with a minor stratigraphic gap, by Early Devonian (Lochkovian) red conglomerates and sandstones (Khapilin, 1982).

The Early Devonian succession (up to 1000 m in westernmost areas, e.g. Komsomolets Island) is represented by red and variegated cross-bedded sandstones, conglomerates, marls, and very subordinate dolomites and limestones. The latter host a vertebrate fauna comparable with that of Svalbard (Karatautate-Talimaa, 1978). Effusives are known in its basal part. Similar sandstones on October Revolution Island (i.e. further east) are less coarse-grained and more sorted. Middle Devonian strata (up to 800 m in the western areas) are red cross-bedded sandstones, marls and minor dolomites with gypsum. The Late Devonian rocks are also continental red sandstones with remnants of wood and plants (Matukhin and Menner, 1999). The whole succession is about 3000 m thick (Fig. 3).

Nalivkin (1973) referred to this as “a typical Old Red Sandstone facies, very similar to the Devonian of Svalbard and other northern Caledonides”.

Kurshs (1982) measured current directions in the Early and Middle Devonian sandstones and studied their mineralogy; he concluded that they were derived from a source in the west and northwest (Fig. 4). The area of provenance also supplied clasts of Ordovician and Silurian carbonates (Menner, in Kurshs, 1982) and metamorphic and magmatic rocks, including granitoids. The heavy mineral associations are dominated by zircon, tourmaline and apatite, while garnet and staurolite are common.

DISCUSSION AND CONCLUSIONS

Distribution of Devonian non-marine red-bed siliciclastic successions from northern Timan and Kolguev Island, northwestern parts of Novaya Zemlya and Severnaya Zemlya provides support for the hypothesis that the Caledonian mountain front extended northeastwards from northernmost Norway through the Barents Shelf to the Eurasian continental edge. As in the foreland basin of the Oslo area, this ORS deposition can be shown in some areas to have started in the Wenlock, apparently in response to Scandian Orogeny further to the west.

The location of the Caledonian deformation front in the Barents Sea is not well constrained; it is placed (Fig. 2) midway between Franz Josef Land and Novaya Zemlya on the basis of the inferred Caledonian age of the deformation and low grade metamorphism of the basement turbidites beneath Franz Josef Land, and the lack of Caledonian deformation on Novaya Zemlya. Seismic data may eventually provide more precise control of the easternmost extension of the Caledonides beneath the Barents Seas.

Location of the Caledonian suture(s) is constrained by gravity, magnetic and seismic data in the southern Barents Sea and geological (on Franz Josef Land) and magnetic data in the north. These allow the approximate definition of a north-northeast trending suture-zone, separating the Timanian basement of northeastern Baltica, to the east, from the Laurentian basement in the west.

Continuation of the Caledonides northwards beneath the Barents Shelf suggests that, prior to the

Gee and Bogolepova
opening of the Eurasian Basin in the Tertiary, this orogen crossed Lomonosova (now the Lomonosov Ridge). Some authors (e.g. Ustritsky in Gramberg, 1988 and Ustritsky, 1990) have suggested that Lomonosova was a Late Palaeozoic orogen truncating the Caledonian and other structures of the Eurasian Arctic Margin. The existing geological and geophysical evidence from the Lomonosov Ridge, admittedly meagre, provide little support for this hypothesis and an alternative interpretation is preferred here that predicts the character of the basement beneath the Lomonosov Ridge based on the geology of the Barentsian Caledonides. The inferred locations of the Barentsian Caledonide front and suture(s) suggest that the basement of the Lomonosov Ridge is characterized by very different lithologies along its length. The Caledonian structural grain of Svalbard trends northwards, perpendicular to the Eurasian margin, and the western parts of the Lomonosov Ridge are probably composed of comparable Laurentian margin terranes, separated by major transcurrent faults. The Caledonian suture(s) coincide approximately with the bend in the Lomonosov Ridge, near the North Pole; this feature may be controlled by this fundamental discontinuity. Further east, the Lomonosov Ridge basement is probably made up of Timanian crust, overlain by Early-Mid Palaeozoic successions and subject to Caledonian folding and southeasterly-directed thrusting.

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EARLY PALAEOZOIC UNCONFORMITY ON SEVERNAYA ZEMLYA AND RELATIONSHIPS TO THE TIMANIAN MARGIN OF BALTICA

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ABSTRACT

The Severnaya Zemlya Archipelago, located to the north of the Tajmyr Peninsula in northernmost Siberia, has a bedrock dominated by a thick Neoproterozoic and Palaeozoic sedimentary succession. Neoproterozoic turbidites are succeeded by shallow marine siliciclastics in the Early and Mid Cambrian, which then give way to black shales and more turbiditic sandstones in the Late Cambrian. Folding, uplift and erosion interrupted deposition in the latest Cambrian and earliest Ordovician, resulting in an angular unconformity. Overlying this unconformity are shallow marine Early Ordovician conglomerates, sandstones and limestones. Rift volcanic rocks appear near the base of the Ordovician and, along with evaporites, reach into the Mid Ordovician, overlain by dark limestones and shales. Late Ordovician quartzitic sandstones pass up into limestones which dominate the Silurian. Overlying Devonian red sandstones were sourced from the west, suggesting a genetic link to Baltica and the Caledonide Orogen.

Recently published palaeomagnetic data for the North Kara Terrane, comprising Severnaya Zemlya, northernmost Tajmyr and large parts of the northern Kara Shelf, have suggested that it might have been an independent microcontinent in the Early Palaeozoic, but the Early Ordovician poles are poorly constrained and Mid Ordovician to Early Silurian poles follow closely those of Baltica. Cambrian faunas from Severnaya Zemlya show affinity to Baltica and also Siberia. Interpreting the North Kara Terrane as a part of Baltica from at least Vendian to Devonian time explains several features in its geological record.

In the Late Palaeozoic, the North Kara Terrane was accreted to Siberia to form the Tajmyr–Severnaya Zemlya orogen as we know it today. All major structures on the islands and on Tajmyr strike northwards towards the continental slope and are likely to be found beneath the Lomonosov Ridge and, perhaps, the margins of the Amerasia Basin.

INTRODUCTION

The Severnaya Zemlya Archipelago is located on the edge of the Kara Shelf in the Eurasian high Arctic (Fig. 1), between the Kara and Laptev seas, and is dominated by four large islands (Fig. 2), Bol’shevik (11 540 km²), October Revolution (14 200 km²), Pioneer (1547 km²) and Komsomolets (8502 km²), with generally low relief and ice–caps rising almost 1000 m above sea level. Together with the northern parts of the Tajmyr Peninsula, the archipelago represents the land–areas of the North Kara Terrane, previously referred to as the Kara Plate (Gramberg and Ushakov, 2000). We prefer the new name because the bedrock of the South Kara Basin is not obviously related to the North Kara Terrane. This terrane, a fragment of continental crust with Neoproterozoic to Cretaceous successions, may have been a small independent continent or part of a

Figure 1 – The western Eurasian Arctic: Tectonic elements and place names. Bathymetric data from Jakobsson et al. (2000).
larger entity, e.g. Baltica. It is thought to have been accreted to Siberia during Late Palaeozoic time, to form the Tajmyr–Severnaya Zemlya orogen as we know it today (Zonenshain et al., 1990).

Major structures on northeastermost Tajmyr strike northwards towards Bol’shevik Island and apparently continue through this island. On October Revolution Island, they swing northwards and are truncated by the continental slope at 82˚N. They are therefore likely to be found beneath the Lomonosov Ridge. The latter is inferred to be a sliver of continental lithosphere (Wilson, 1963 and Lawver et al., 1988) that was detached from the Eurasian shelf by the opening of the Eurasia Basin in early Cenozoic time. Possible continuations of Severnaya Zemlya structures across the Amerasia Basin remain to be defined. A good knowledge of the bedrock–geology and the tectonic history of the Arctic margins is crucial for all reconstructions of the Arctic before the opening of the Eurasia and Amerasia basins, including an improved understanding of the North Kara Terrane’s history before it became an integral part of Eurasia.

HISTORY

In 1913, B. A. Vil’kitsky’s expedition discovered Severnaya Zemlya and named it Tsar Nikolaj II Land. The archipelago was renamed to Severnaya Zemlya in 1926, and the question naturally arises whether or not it should revert to the original name. Scientific activity began with G. A. Ushakov’s and N. N. Urvantsev’s expeditions in the years 1930–1932, with the first topographical and geological studies (Urvantsev, 1933). The first geological map of the archipelago was published by Egiazarov (1957) in 1:1,500,000 scale, after fieldwork from 1948 to 1951. It was followed by 1:200,000 scale maps published in 1984 and 1985 (Markovsky et al., 1984, 1985), based on the work of the USSR State Geological Mapping Program between 1973 and 1979. Since 1967, geophysical studies, including airborne potential field surveys (magnetics and gravimetry), have been carried out over the archipelago. Since the late 1980’s, the focus of geological and geophysical work in the Severnaya Zemlya Archipelago has been on prospecting, particularly in southeastern Bol’shevik Island, where gold was mined until recently. Two major volumes, in Russian, on the geology of the Severnaya Zemlya Archipelago, edited by Kaban’kov and Lazarenko (1982) and Gramberg and Ushakov (2000) provide the main basis for our knowledge of the geology.

In 1998, 1999, 2002 and 2003, international geoscientific expeditions, supported by SWEDARCTIC, visited Tajmyr and the Severnaya Zemlya Archipelago (Gee and Pease, 1999; Gee, 2002; Gee and Lorenz, 2003) to investigate the regional tectonics. Work focused mainly on stratigraphy, palaeontology and structural geology, and collecting samples for isotope age and palaeomagnetic studies.

This paper presents results from new work on the Early Palaeozoic strata of October Revolution Island, with a focus on the occurrence of a major angular unconformity at the base of the Ordovician, and the implications for the interpretation of the history of the North Kara Terrane.

STRUCTURE OF SEVERNAYA ZEMLYA

Neoproterozoic strata dominate Bol’shevik Island and Palaeozoic successions comprise October Revolution Island, Pioneer Island and southern Komsomolets Island. Northern Komsomolets Island is almost entirely covered by Quaternary deposits, with some Devonian outcrops. The dominating structure in these rocks developed in the earliest Carboniferous, and was followed by uplift, erosion and deposition of Mid Carboniferous to Early Permian strata (Dibner, 1982). October Revolution Island is divided into two structural zones, separated by a major NNE–trending

Figure 2 – Geological map of the Severnaya Zemlya Archipelago, from Egiazarov (1967), showing the Fiordovoe Lake Fault Zone.
fault, named the Fiordovoe Lake Fault Zone (Fig. 2; Lorenz, 2004). The western zone includes central, northern and northwestern October Revolution Island, and continues westwards into southeastern Komsomolets Island, Pioneer Island and the Sedov Archipelago; the eastern zone comprises eastern October Revolution Island. Early Ordovician to Devonian strata dominate the western structural zone and are deformed into NW–trending, upright to NE–vergent close to tight folds with approximately 10 km wavelength (Figs. 2 and 3). Westwards, on Pioneer Island and westernmost October Revolution Island, only Silurian and Devonian rocks are present. They are deformed into major NW–trending, gentle folds (bedding dips in general less than 10°) with approximately 10–15 km wavelength. Further west, on the Sedov Archipelago (Figs. 2 and 3), Ordovician rocks reappear in the hinge of a tight anticline.

Cambrian to Mid Ordovician strata dominate the eastern structural zone. They are deformed by...
subhorizontal to slightly S–plunging folds which are disrupted by several NNE–trending faults. In the south, the major folds are broad and open with a wavelength of up to 10 km and show parasitic folding with 1–2 km wavelength. In the north, folding is close, upright to steeply ENE–inclined with a wavelength of up to 5 km. West of Cape Sverdlov, the major E–vergent Snezhnaya Bay Anticline folds Mid Cambrian to Ordovician strata, and the succession is vertical to partly overturned in its eastern limb.

Further to the east, on Bol’shevik Island, Neo-proterozoic successions are deformed into approximately N–trending folds; they are intruded by Early Carboniferous granites and unconformably overlain by Late Carboniferous (Dibner, 1982) sediments (Fig. 2). A major, 20–25 km wide and more than 100 km long fault zone, the Severnaya Zemlya Fault Zone (Proskurnin, 1995), has been inferred to exist in eastern October Revolution Island, between Cape Sverdlov and the Rovnaya River (Fig. 3). Marked by a concentration of volcanic rocks and a positive magnetic anomaly, it has been interpreted to be caused by Early Ordovician rifting (Proskurnin, 1999). The related igneous rocks are subalkaline to alkaline volcanics and high level intrusive rocks. The latter have been dated to 455 ± 15 Ma by the K–Ar method (Proskurnin, 1995). Kaplan et al. (2001) reported an Ar-Ar amphibole age of 434 ± 2 Ma on a gabro from eastern areas and a U-Pb zircon age of 470 ± 15 Ma for a granite from the same belt of igneous rocks in south–central parts. New work (Lorenz et al. in press) has shown that the igneous activity started in the Tremadocian (ca. 490 Ma).

STRATIGRAPHY OF SEVERNAYA ZEMLYA

In the Russian literature, the stratigraphic units are mainly referred to as "svitas". Our mapping on October Revolution Island suggests that these "svitas" usually correspond either to formations or to groups in the generally accepted international lithostratigraphic nomenclature; however, pending remapping of lithostratigraphic units, we prefer in this paper to retain the Russian terminology.

Outcrops of the Proterozoic bedrock have been thought to be restricted to Bol’shevik Island (Fig. 2), but unfossiliferous sections beneath the Cambrian of October Revolution Island may also be Neoproterozoic in age (see below). On Bol’shevik Island, an about 2100–2500 m thick turbidite dominated succession has been described (Shul’ga, 2000; Kabankov et al., 1982), which is composed of ve svitas, metamorphosed up to lower greenschist facies. The youngest of these has been dated to late Vendian using acritarchs; relationships to the Cambrian are not known.

**Figure 4 – Stratigraphic units of the Severnaya Zemlya Archipelago (based on Shul’ga 2000).**
Figure 5 – Geological interpretation of southeastern October Revolution Island, based on fieldwork in key-areas and satellite image-interpretation (modified from Lorenz, 2004).
The Palaeozoic successions (Fig. 4) composing the other islands of the Severnaya Zemlya Archipelago were described by Shul’ga (2000) and Markovsky and Makar’ev (1982). Cambrian strata (about 2500 m thick) dominate the eastern parts of October Revolution Island and are composed of four units (Fig. 4). The lowermost exposed strata (Nekrasov Svita) are composed of unfossiliferous turbidites and are conformably overlain by fossiliferous Early Cambrian strata (Kabankov et al., 1982). The unfossiliferous lower part of the unit has been regarded as Cambrian by Lazarenko (1982), but Proskurnin (1999) suggested a Vendian age because of similarities in lithofacies and metamorphism to the Neoproterozoic rocks on Bol’shevik Island. Shales and siltstones dominate the overlying Early Cambrian Marat Svita, passing up into sand- and siltstones of the Mid Cambrian Universitet Svita. In the Late Cambrian the basin deepened and Kurchavaya Svita dark and black shales were deposited; sandstones, often graded, appear in the upper part of this unit.

Four svitas (Fig. 4) compose the Ordovician succession, which is up to 1700 m thick. All four svitas are developed in central and northern October Revolution Island; however, in southeastern areas, only the lower two units are present, and on northwestern October Revolution Island and southwestern Komsomolets Island, only the upper three. Early Ordovician rocks of the Kruzhilikha and Ushakov svitas overlie the Late Cambrian strata unconformably. The Kruzhilikha Svita is dominated by sandstones and limestones. The Ushakov Svita, introduced by Markovsky and Makar’ev (1982) as the unit conformably overlying the Kruzhilikha Svita, consists of multicoloured sandstones, siltstones and gypsumiferous marls with limestones and volcanics. They defined the lower boundary of the Ushakov Svita as the base of the multicoloured sandstones and/or the top of light grey sandstones and limestones.

Mid Ordovician strata (Ozernaya Svita) consist mainly of dark shales and carbonate rocks, which commonly are gypsumiferous. In the Late Ordovician (Stroynaya Svita), sandstones are overlain by fossiliferous limestones, which continue into the Silurian. Silurian strata, up to 2500 m thick, are present in central and western October Revolution Island, western Pioneer Island and southwestern Komsomolets Island. They have been divided into five svitas (Fig. 4), dominated by carbonate rocks, including organic and detrital limestones, stromatolitic limestones and dolomites. Red marls appear near the top of the Silurian succession.

Devonian red sandstones are up to 2300 m thick and are widespread in central and western October Revolution Island and cover Pioneer Island and large parts of Komsomolets Island. They rest with a hiatus or minor unconformity on the Late Silurian succession. Fish and plant fossils are reported (Shul’ga, 2000). The Devonian sandstones contain subordinate intercalations of mudstones and some thin carbonate units.

Early Carboniferous to Permian rocks overlie a major unconformity, but are rarely preserved. On northern Bol’shevik Island, Early Carboniferous (Dibner, 1982) sandstones, siltstones and conglomerates occur in Akhmatov Fjord in the north of the island (Fig. 3) overlying the folded Riphean and Vendian strata (Gramberg and Ushakov, 2000). On October Revolution Island, Late Carboniferous (Dibner, 1982) sandstones, siltstones and conglomerates overlie folded Early Ordovician strata. On northwestern Komsomolets Island, Early Permian (Dibner, 1982) sandstones with subordinate mudstones and conglomerates rest on Early

Figure 6 – Kruzhlkhka River area. For location see Fig. 5.
Figure 7 – The Early Ordovician unconformity at the Kan’on River. a) High-angle contact in the western limb of the Kan’on River Syncline. Black shales with graded sandstone beds of the Kurchavaya Svita are unconformably overlain by the basal conglomerate of the Kruzhilikha Svita. b) Contact in the eastern limb of the Kan’on River Syncline. Black shales of the Kurchavaya Svita are tightly folded with axial planes perpendicular to the bedding of the unconformably overlying, now subvertical conglomerates and sandstones of the Kruzhilikha Svita. Photography is taken northwards along the about 40 m high cliffs of the Kan’on River. c) Basal conglomerate of the Kruzhilikha Svita. d) A massive green tuff layer, here in a about 4 m high exposure, is a marker horizon throughout the region, mostly in form of scree. e) Section of the Kruzhilikha Svita in the eastern limb of the Kan’on River Syncline at locality A (Fig. 8). The red sand-and siltstones belong to the overlying Ushakov Svita.
and Mid Devonian strata.

**THE EARLY ORDOVICIAN KAN’ON RIVER UNCONFORMITY**

Interpretation of the nature of the contact between the Cambrian and Ordovician on October Revolution Island has been controversial. Egiazarov (1957) concluded that it was an angular unconformity, which was supported by Proskurnin (1999). Makarev et al. (1981) argued for a thrust fault because they observed striations on the contact plane; Markovsky and others did not exclude displacement, but along the weakness zone of an angular unconformity (from Proskurnin, 1999). This study demonstrates, that the disputed contact is an angular unconformity; the name “Kan’on (canyon) River Unconformity” is introduced, based on the well exposed locations in the lower reaches of this river.

The Cambrian–Ordovician contact is exposed in several places in the eastern part of October Revolution Island (Figs. 2 and 3). The unconformity is folded in open (in the south) to close (in the north), upright to E–vergent and slightly S–plunging folds of up to 10 km wavelength (Fig. 5). NNE–striking faults are common, displacing the contact and locally cutting out parts of the section.

The stratigraphic unit underlying the unconformity throughout the area is the Kurchavaya Svita. According to Makarev et al. (1981) this unit is about 800 m thick. Our mapping has indicated a thickness of at least 2000 m in the Kurchavaya River type area, where it is developed as a unit of black and dark grey shales. Its lower part includes a few layers of siltstone and sandstone; up-section the number and thickness of sandstone beds increases and channel deposits with pebbly to coarse grained sandstone have been observed. In areas further north, another Late Cambrian siltstone and sandstone dominated facies has been mapped. It is best seen south of Ostroia Lake, between Vos’merka Hill and Kan’on River (Figs. 3 and 5). Here, the black and dark shales give way to medium grey shales and siltstones and further up-section, grey sandstones. Both these facies occur directly beneath the unconformity. All strata directly overlying the unconformity have been assigned to the Early Ordovician Kruzhilikha Svita (e.g. Gramberg and Ushakov, 2000), which is 100–250 m thick (Shul’ga, 2000) and consists of conglomerates, sandstones, fossiliferous limestones and a few tuffs. In the course of this study, some of the units immediately overlying the unconformity in the northwest have been inferred (based on lithology) to be basal parts of the Ushakov Svita (600–1200 m thick), with its dominance of multicoloured and tuffaceous sandstones, gypsiferous marls (Shul’ga, 2000) and volcanic rocks. Elsewhere, the Ushakov Svita conformably overlies the Kruzhilikha Svita.

In the following text, the relationship between the Cambrian and Ordovician strata is described from four key–areas which have been studied during recent eldwork (Fig. 5): Kruzhilikha River, Lake Fiordovoe, Kan’on River and Kurchavaya River–Snezhnaya Bay areas.

**KRUZHILIKHA RIVER AREA**

The Kruzhilikha River and its tributaries drain the area south of Lake Fiordovoe into the Ozernaya River (Figs. 3 and 5). Along the riverbanks, the Late Cambrian and Early Ordovician successions are well exposed. The strata are folded into an anticline–syncline pair with several higher–order folds. At the best exposed locality visited by us (Fig. 6, locality A; N 79.16270, E 097.28930), documented in Proskurnin (1999), Gee and Pease (1999) and Bogolepova et al. (2001), the Kurchavaya Svita is developed in dark shale facies with interbedded centimetre to decimetre thick beds of brownish graded sandstone. Molluscs, trilobites, brachiopods and acritarchs have been reported from these Cambrian units (Bogolepova et al., 2001; Rushton et al., 2002). They dip steeply (60–80°) northwards and are unconformably overlain by very gently SW–dipping pale grey conglomerates and calcareous sandstones. These basal beds of the Kruzhilikha Svita, a few metres thick, pass up into sandy limestones with brachiopods and lingulids and include subordinate, thin, organic–rich black shales. In the stream section of a tributary owing from the southeast into the Kruzhilikha River at locality A (Fig. 6), the basal beds occur 1.5 km to the southeast, where they are overlain by a thin (20 cm), bright green, acid volcanic unit and then by green and red cross– and ripple–bedded sandstones. The latter, a few tens of metres thick, are overlain by green siltstones and white rhyolitic volcanic units, above which there occur various volcanioclastic sandstones and conglomerates.

At a second locality (Fig. 6, locality B) along the Kruzhilikha River, upstream from locality A, the unconformity is also well exposed. Only the basal brownish sandstones, which contain brachiopods and molluscs, and the conglomerates of the Kruzhilikha Svita are exposed, overlain by greenish and white sandstones. Downstream from locality A, the unconformity is exposed again (Fig. 6, locality C), where the contact is nearly concordant. Basal calcareous sandstones and conglomerates (including a thin black shale) pass rapidly upwards into green and red sandstones. Variegated, mostly red and green sandstones and marls with tuffs and other volcanic
rocks, including basalts and rhyolites, are characteristic of the Ushakov Svita in the area of Fig. 6, as elsewhere on October Revolution Island. In the Kruzhilikha River area, discontinuity of exposure prevented a detailed study of the relationships between the Kruzhilikha and Ushakov svitas.

KAN’ON RIVER AREA

Meltwater from the Universitet Glacier feeds westwards into the Kan’on River and then northwards into Lake Ostroia (Fig. 3 and 5). Excellent sections through the Early Ordovician unconformity are exposed from Lake Ostroia southwards along the steep canyon sides of Kan’on River and its western tributaries (Fig. 7). The unconformity is exposed in both limbs of the N–trending Kan’on River Syncline, a close fold in which the western limb dips approximately 45° E and the eastern limb is subvertical to vertical, or even slightly overturned westwards. From Lake Ostroia, the fold continues at least 25 km to the south, its hinge partly cut by a N–trending fault zone.

There are many good sections through the unconformity from Lake Ostroia southwards to where the Kan’on River bends northwards and a tributary enters from southwest (Fig. 8, locality A; N 79.23087, E 097.96237). At this locality in the tributaries, and throughout the western limb of the syncline, Kurchavaya Svita dark shales and thin sandstones dip about 45° W, approximately perpendicular to the E–dipping unconformity (Fig. 7a, N 79.25477, E 097.99515). The eastern limb of the syncline is well exposed in the main river section, where the dark shale dominated Kurchavaya Svita, dipping at moderate to low angles eastwards, is in contact with unconformably overlying, sub-vertical conglomerates. Remarkably, at one location in the western cliff of the Kan’on River (Fig. 8, locality B; N 79.23855, E 097.98052), the Kurchavaya dark shales are folded with axial surfaces at–lying and perpendicular to the unconformity (Fig. 7b).

In the Kan’on River section, the unconformity is overlain by conglomerates with a clast size from a few centimetres up to several decimetres (Fig. 7c); they occur in several beds interlayered with coarse and pebbly sandstones. A few metres up–section, the first calcareous sandstone and sandy limestone with brachiopods, molluscs and trilobites occurs. These lithologies become more common higher in the sequence. A bright green tuff in a 2–3 m thick layer (Fig. 7d), located approximately 60 m above the unconformity, is a stratigraphical marker in the region. The first fossiliferous (brachiopods and trilobites) limestones appear a few metres below the tuff bed. Sandy limestones, and often fossiliferous limestones (oysters) constitute the overlying 60–70 m of the Kruzhilikha Svita.

Figure 8 – Kan’on River area. For location see Fig. 5.
The boundary with the overlying Ushakov Svita is marked by a lithological change from limestones to red and green sand- and siltstones, which occupy the hinge of the syncline (Fig. 7e).

The Kan’on River Syncline continues south of the bend in the Kan’on River (Figs. 5 and 8), but exposures are poor and limited to small, west-owing creeks crossing the syncline. Marker lithologies (e.g. the green tuff) can commonly be followed in the scree. About 3 km south of the section described above (Fig. 8, locality C), the main difference observed is a variation of the basal unit which is composed of very coarse grained quartz sandstone. At a creek about 5 km south of the Kan’on River bend (Fig. 8, locality D; N 79.25477, E 097.90059), the basal units are again conglomerates and the stratigraphic thickness below the green tuff layer is estimated to have increased to 90 m.

About 2 km farther to the south (Fig. 8, locality E; N 79.16993, E 097.89186), in the western limb of the syncline, the basalt pebbly sandstones are underlain, with a small angular discordance, by green mudstones, sandstones and shelly limestones, at least a few tens of metres thick. These are inferred to overlie the Kurchavaya Svita with major unconformity, as in the locations further north, although the contact was not seen.

LAKE FIORDOVOE AREA

Lake Fiordovoe (93 m asl., Figs. 3 and 5) is the largest lake in southeastern October Revolution Island; it is separated from Lake Ostroia, in the east, only by ice from the Karpinski Glacier, which borders both lakes to the north. Along the shore of Lake Fiordovoe, Cambrian to Silurian bedrock is exposed. The unconformity is exposed both close to the lake and 8 km further south.

Along the long and narrow fjord which forms the southern end of the lake (Fig. 9), the unconformity is folded into the slightly N-plunging Fiordovoe Lake Syncline which is cut by a fault in the eastern limb (Figs. 9 and 10a). Both the silt-/sandstone dominated and the dark/black shale facies of the Kurchavaya Svita occur in this area. The former underlies the W-dipping unconformity in the central and eastern parts of the syncline, while the latter is present close to Lake Fiordovoe.

A section across the unconformity can be studied on the slope above the southern end of the lake (Figs. 9 and 10a, locality C; N 79.31193, E 097.61795). Here the W-dipping Kurchavaya Svita consists of approximately 50% dark shales interlayered with graded sandstone beds up to a few centimetres thick (Fig. 10b). The high-angle contact (>70°; Fig. 10b) with overlying brown pebbly sandstones, conglomerates and very coarse grained sandstones has a rough surface. A strong high-angle cleavage is present in both the Cambrian shales and the Ordovician sandstones/conglomerates, indicating the importance of younger deformation. Above a 5 m thick conglomerate unit, the section is not well exposed, but the presence of a red and a bright green tuff or tuffaceous sandstone can be recognised in the scree. Up-section, a thin unit of brown, coarse-grained sandstones and red sandstones with thin bands of green tuff and joints of orange gypsum occurs. In general, grain-size decreases up-section, the sandstones are overlain by a unit of evaporites (Fig. 10c), consisting of many thin (up to a few metres) cycles with conglomerate, gypsum and salt-pseudomorphs. The unit is heavily deformed and

Figure 10 – The Early Ordovician unconformity along the southern shore of Fiordovoe Lake. a) General view looking northwards (Fiordovoe Lake in the background): conglomerates, sandstones and evaporites unconformably overlie shales, silt- and sandstones of the Kurchavaya Svita. The unconformity, here folded in a slightly northwards plunging syncline, is marked by the line. b) The unconformity along the steep slope close to Fiordovoe Lake: cleavage cross the unconformity. c) Evaporites above the unconformity. No exposures are present, but the cyclic nature is still visible in the scree, where it is possible to trace the gypsum (foreground) and conglomerate beds.
eroded, prohibiting a thickness–estimate (Fig. 10c). Along this arm of Lake Fiordovoe, the evaporites are bordered by the unconformity in the west and south, Lake Fiordovoe in the west and north and a fault–contact to silt- and sandstone dominated Kurchavaya Svita rocks in the east. In general, the evaporite–bearing strata are common along the southern shore of Lake Fiordovoe (Fig. 5), west of the described location, with fault–contacts to the Late Cambrian strata; they continue from the western termination of Lake Fiordovoe in a thin band farther northeastwards into central October Revolution Island. Beds of multicoloured tuff are present within the evaporites. The upper contact is exposed in two locations: on the northernmost peninsula in the above mentioned fjord (Fig. 9, locality D; N 79.35126 , E 097.77799 ) and to the west of the fjord, towards the outlet into the Ozernaya River (Fig. 9, locality E; N 79.34421 , E 097.61603 ). Here limestones, some with stromatolites, overlie the evaporitic succession which shows an increasing volcanic in ux in the multicoloured tuff towards the contact; in the western location an about 0.5 m thick bed of basic volcanic rock has been observed.

The combination of multicoloured sandstones in the lower part and various volcanic rocks and limestones higher up in the section implies that this unit should be assigned to the Ushakov Svita. The extent of the unit, which can be traced at least 30–40 km by its spectral signature on satellite imagery (Fig. 3), suggests a status as an independent stratigraphic unit. It is of particular interest that it rests directly on the unconformably underlying Cambrian strata; the Kruzhilikha Svita is apparently absent in this area, unless represented by the basal unfossiliferous conglomerate.

The area between the Kruzhilikha River and Lake Fiordovoe (Fig. 5) consists mainly of tundra and boggy tundra, drained by small streamlets which do not incise into bedrock. These collect in an unnamed tributary to the Kruzhilikha River, which exposes a little bedrock along its banks. In general, it follows the Cambrian–Ordovician boundary, which is interpreted to be part of the Fiordovoe Lake Fault Zone (Lorenz, 2004).

About 12 km north of the locations in Kruzhilikha River (Fig. 6), and about 8 km SSE of southernmost Lake Fiordovoe, the above mentioned stream cuts a section through red sandstones, referred to here as the “Red Section” (Figs. 5 and 9, locality B). Two kilometres south of the Red Section, about 100 m east of the river, red conglomerates and pebbly sandstones are the lowermost exposed strata (Figs. 9, locality A, and 11a; N 79.39144 , E 097.26477 ). Interpreting the scree, they are directly overlying the silt- and sandstone dominated facies of the Kurchavaya Svita; the contact itself is not exposed. The conglomerate and pebbly sandstone beds are some tens of centimetres to 1.5 m thick and folded in a local syncline. The Cambrian–Ordovician contact is interpreted as the Early Ordovician unconformity in its northernmost location close to where it is cut out by the Fiordovoe Lake Fault Zone. Only a few tens of metres of basal Ordovician strata are preserved at this location in the Red Section (Fig. 9, locality B), upstream the junction with the tributary (N 79.27929 , E 097.40554 ) and 500 m to the west, the Ordovician bedrock is well exposed (Fig. 11b).

About 200 m of alluvial red sandstones crop out, but the lower contact is not exposed. Sandstones of variable grain–size have bed thicknesses of up to a few decimetres, show cross–bedding and asymmetric sinuous ripple marks of 5–10 cm wavelength (Fig.
After a few metres of pale sandstones interbedded with red–ne–grained sand- and siltstones, a prominent layer of green sandstone marks the beginning of volcanioclastic sedimentation which is topped by multicoloured tuff (Fig. 11d), consisting of red, white, brown and violet layers, similar to the multicoloured tuffs in the evaporites of the Lake Fiordovoe area. Uppermost in the exposed section are white to pale-grey sandstones. The multicoloured tuff and white to pale–grey sandstones are recognisable in the western limb of the syncline which folds the described section. However, no bedrock is exposed and only the volcanics can be traced in the scree.

The section above the unconformity lacks carbonate rocks and the calcareous sandstones typical of the Kruzhilikha Svita; it contains, however, the multicoloured and tuffaceous sandstones and volcanics which are characteristic for the Ushakov Svita.

Located about 6 km east of the locations at Lake Fiordovoe described above, Vos’merka Hill (261 m asl.; N 79.29950 , E 097.86977 ; Fig. 9) is 110 m higher than the surrounding tundra. The eastern limb of a syncline forms the hill and exposes Late Cambrian and Early Ordovician strata on its eastern slope. Bedding, where exposed, is approximately parallel to the gradient of the western slope of the hill. The western limb and the closure of the slightly S–plunging syncline can be traced in the tundra west and north of the hill.

In the valley to the east of Vos’merka Hill (Fig. 9, locality F), the Kurchavaya Svita is represented by dark and black shale facies and shows deformation in form of small folds (tens of metres wavelength). Following the exposures in the streams from Vos’merka Hill, the sand/shale ratio increases; cross–bedded channel–sandstones indicate that the section is the right way up. Along a small creek farther north on the hillside (Fig. 9, locality G; N 79.30795 , E 097.91351 ), the silt- and sandstone facies of the Kurchavaya Svita overlies the dark and black shale facies, as it is observed on the next hill to the east (P 257 m asl.). Grey sandstones are interlayered with subordinate grey shale; conglomerates in the base of channels are present throughout the approximately 75 W dipping succession. Unfortunately, the upper contact of the siliciclastic unit is not exposed. A massive, grey, sandy limestone, also dipping westwards, is the next unit exposed, after 100 m of scree. It has a distinctive orange–brown weathering colour and forms outcrops in the form of small ridges and cliffs along the hillside and farther north. The ridge and summit of Vos’merka Hill is made up of a thick bright green tuff. It is underlain by a layer of red tuff (along the ridge) or sandstone (at the summit). Near the top of the hill, the scree contains boulders of grey, yellow weathering, pebbly sandstone and conglomerate.

The Cambrian–Ordovician contact is not exposed on Vos’merka Hill. Structural observations lead to the conclusion that no major angular discordance exists; however, it is likely that the conglomerates in the scree originate from the basal conglomerates above the Kan’on River Unconformity. The siliciclastic, volcanic and calcareous units which overlie the grey silt- and sandstones belong to the Kruzhilikha Svita.

**KURCHAVAYA RIVER–SNEZHNAYA BAY AREA**

Along the southwestern margin of the Universitet Glacier, from the vicinity of the Kurchavaya River and eastwards (Figs. 3 and 5), the Cambrian and Ordovician sections are exposed in a major S–plunging, E–vergent fold, named here the Snezhnaya Bay Anticline. In the core of this fold, grey green–ne–grained sandstones and highly bioturbated mudstones of the Mid Cambrian Universitet Svita (Makarev et al., 1981) are overlain by black shales with some limestone concretions of the Kurchavaya Svita. Higher in this unit, beds of graded...
sandstone and channel sandstones appear; they increase in thickness and frequency upwards. Acidic porphyry sills (and occasional feeder dykes) intrude the Kurchavaya Svita rocks in the vicinity of Snezhnaya Bay (Fig. 12), where the geological map of Markovsky et al. (1984) shows a large intrusion with related contact metamorphism. The porphyries are gently dipping sheets and they occur now as isolated bodies which form topographic highs above underlying eroded black shales.

The contact between the Kurchavaya Svita and overlying Ordovician successions was not observed by us in the section west of the Kurchavaya River where the lowermost exposures are composed of gently W–dipping grey and red sandstones, mudstones and subordinate limestones.

In the eastern limb of the Snezhnaya Bay Anticline, both the Cambrian black shales and turbidites and the Ordovician red sandstones and volcanics strike north–south and dip approximately vertically. The units overlying the black shales and sandstones of the Kurchavaya Svita are dominated by acid to intermediate volcanics with subordinate basic rocks; intercalations of conglomerates, sandstones and siltstones occur. Neither the western nor the eastern limbs of the Snezhnaya Bay Anticline contain strata characteristic for the Kruzhilikha Svita and it is probable that the Ushakov Svita rests directly on the Kurchavaya Svita.

**DEFORMATION OLDER THAN THE KAN’ON RIVER UNCONFORMITY**

As described above, the Kan’on River Unconformity shows, in most exposures, an angular relationship between the underlying and overlying units of about 90°. Despite this, the Early Ordovician strata apparently everywhere overlie the same stratigraphic unit, the Late Cambrian Kurchavaya Svita.

Only one area, in the vicinity of the Kan’on River, yields evidence for pre-Ordovician folding. Here the poles to cleavage from rocks below the unconformity approximate a great circle (Fig. 13a) while those from rocks above the unconformity approximate a pole (Fig. 13b). Although the available number of measurements is low, these results indicate the presence of pre-Ordovician cleavage; elsewhere, cleavage is clearly of Late Palaeozoic age.

Unambiguous evidence of folding prior to the Ordovician has been found only in the eastern limb of the Kan’on River Syncline (Fig. 8), where the axial surfaces of the folds in the Kurchavaya Svita are approximately perpendicular to the unconformity. It can be concluded from this tight folding, the generally high angle orientation of the Cambrian strata prior to Ordovician deposition and the occurrence of Late Cambrian units directly beneath the unconformity.

![Figure 13 – Cleavage data from the Kan’on River area plotted on lower hemisphere, equal area nets (Schmidt Net). a) Data from Late Cambrian rocks below the Kan’on River Unconformity. b) Data from Early Ordovician rocks above the Kan’on River Unconformity.](image)
throughout southeastern October Revolution Island, that this episode of Early Palaeozoic deformation involved regional shortening largely by upright folding of short wavelength and amplitude. Further work on the internal structure of the Kurchayava Svita and evidence for way-up will be necessary to better understand the character of this deformation.

**AGE OF THE UNCONFORMITY AND CORRELATION**

Strata beneath the unconformity in the Kruzhilikha River (locality A, Fig. 6) yield the trilobites *Kujandaspis ketiensis*, *Protopeltura holdtedahli*, *Maladiodella aff. abdita* and the acritarchs *Impluviculus multiangularis*, *Leiofusa stoumoensis*, *Lusatia dramatica*, *Veryhachium dumontii*, indicating a Late Cambrian age (Agnostus and Peltura minor zones; Bogolepova et al., 2001, Rushton et al. 2002). Acritarchs from the dark shales of locality A in Fig. 8 confirm this age (Raevskaya and Golubkova, in press). Sandstone lenses below the unconformity in the Kruzhilikha River (locality A, Fig. 6) yield the brachiopods *Finkelnburgia*, *Orusia* and *Euorphisca*; the last of these has not previously been reported below the Tremadocian (Rushton et al., 2002). Markovsky and Makar’ev (1982) recorded the problematic fossils *Angarella lopatini* and *Angarella sp.* in the basal beds of the Kruzhilikha Svita. These forms also occur in the Arenig–Llanvirn of Siberia, Tajmyr, Polar Urals, Paj–Khoj and Novaya Zemlya. Carbonate nodules above the unconformity in the Kan’on River (locality A, Fig. 8) yield Mid Tremadocian to Early Arenig conodonts (T. Tolmacheva, pers. comm.) and acritarchs with a stratigraphic range from Late Cambrian to the Early Ordovician (Raevskaya and Golubkova, in press).

Higher up in the section, zircons from a tuff layer yielded an U-Th-Pb ion-microprobe age of c. 490 Ma (Lorenz et al. in press). These data limit the time of deformation, erosion and therewith development of the unconformity to the latest Cambrian (Landing et al., 2000).

Although the strata above the unconformity vary in the area of investigation, several stratigraphic markers are present. Most prominent is the bright green tuff layer (Figs. 7d, 7e and 14), which reaches a thickness of up to 5 m in the Kan’on River section and on Vos’merka Hill. In both areas this layer can be traced for long distances in exposure and scree. A thin layer of similar green tuff, only 0.2 m thick, has been observed in the Kruzhilikha River area (Fig. 1).

All the strata overlying the unconformity were regarded as part of the Kruzhilikha Svita by Markovsky and Makar’ev (1982), which Egiazarov (1970) introduced as a unit dominated by sandstones and carbonate rocks with its incomplete type section in the middle reaches of Kruzhilikha River. A second section has been described above, in the eastern limb of the Kan’on River syncline (Fig. 7e), which exposes a complete section between the unconformity and the Ushakov Svita. These sections can be regarded as

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**Figure 14** – Stratigraphic sections across the Early Ordovician unconformity. For geographic location see Figs. 5 and 9.
PALAEOGEOGRAPHIC RELATIONSHIPS

Many of the Early Palaeozoic successions of October Revolution Island are fossiliferous. Lazarenko (1982) suggested a close relationship between the Middle and Upper Cambrian of Severnaya Zemlya and Novaya Zemlya on the basis of a detailed study of trilobites from both areas. Bogolepova et al. (2001) summarized published data, information from reports and their own data for the whole Cambrian. On the basis of the biogeographical distribution of the trilobite genera and species, they favoured the independent existence of the North Kara Terrane, but in proximity to both Baltica and Siberia throughout the Cambrian. The same authors reported Late Cambrian acritarchs with an affinity to Baltica. However, acritarchs from the Upper Cambrian of Siberia are not known. Rushton et al. (2002) identified new species of trilobites and brachiopods from the Late Cambrian Kurchavaya Svita. The observed trilobite species prevails on the Siberian platform; brachiopods, however, are endemic to Severnaya Zemlya and suggest that it was not part of Siberia during the Late Cambrian.

The only palaeomagnetic study on the Severnaya Zemlya Archipelago has been published by Metelkin et al. (2000). They interpret the North Kara Terrane (their Kara Plate) as an independent plate from the time of their oldest pole, 500 Ma, until it collided with Siberia in the Late Carboniferous. The data for the oldest pole are from samples both below and above the Kan’ on River Unconformity, from bedrock of Mid Cambrian to Mid Ordovician age (Fig. 12). This casts doubt on the significance of the pole in question. The Late Ordovician and Late Silurian poles (ca. 450 and ca. 420 Ma) follow the Apparent Polar Wander Path of Baltica (from Torsvik and Rehnström, 2003), apparently, with a small temporal offset.

Pease (2001) presented the first results of a detrital zircon study of turbidites from Tajmyr’s northern belt. These strata have been interpreted to be deposited on a continental slope during the Neoproterozoic and Cambrian. They are inferred to continue northwards into the succession exposed on Bol’shevik Island. Ages between 555 and 570 Ma (Pease, 2001) dominate the two samples, a signature not typical for northwestern Siberia, but characteristic of northeastern Baltica (Pease, 2001), where subduction related granites intrude the Pechora Basin basement (Gee et al., 2000).

Another line of evidence favouring the affinity of the Severnaya Zemlya Neoproterozoic and Palaeozoic successions to Baltica is to be found in the occurrence of Old Red Sandstones and their provenance from the west. It has been argued elsewhere (Gee and Bogolepova, this volume) that these non-marine siliciclastic successions were derived from the Barentsian Caledonides of Baltica.

The age of deformation and erosion related to the Kan’ on River Unconformity on October Revolution Island is known in the Caledonides of Baltica (Roberts and Gee, 1985). The deformation on October Revolution Island could be related to deformation in the hinterland of the Barentsian Caledonides.

CONCLUSIONS

The Palaeozoic successions of October Revolution Island are influenced by at least two major episodes of deformation, the earlier one close to the Cambrian–Ordovician boundary, related to the Kan’ on River Unconformity, and the later in the latest Devonian and/or earliest Carboniferous. An important facies change from deep marine Vendian turbidites to shallow marine sandstones, conglomerates and carbonate rocks occurs in the Early to Mid Cambrian. This change may also be accomplished by a significant hiatus, perhaps related to Timanian orogeny elsewhere on Baltica (Gee and Pease, 2004).

The major Kan’ on River Unconformity is present throughout southeastern October Revolution Island. Underlying and overlying rocks generally have a high–angle relationship which can be observed along the Kruzhilikha and Kan’ on rivers and along the southern shore of Lake Fiordovoe. The underlying shales, siltstones and sandstones belong, apparently without exception, to the Late Cambrian Kurchavaya Svita, including the Agnostus and Peltura minor zones. The Peltura scarabaeoides and Acerocare zones have not been recorded.

In general, conglomerates, sandstones and limestones of the Kruzhilikha Svita directly overlie the Kan’ on River Unconformity. However, in the vicinity of Lake Fiordovoe, red sandstones, evaporites and volcanic rocks were deposited on the unconformity. Here, and in the southeast of October Revolution Island, the Kruzhilikha Svita may be absent.

The oldest fossils that so far have been found in the lower part of the Kruzhilikha Svita is mid–Tremadocian to Arenig in age and zircons from a tuff bed have yielded an U-Th-Pb ion-microprobe age of c. 490 Ma. These age data suggest a latest Cambrian to earliest Ordovician period of deformation and erosion for the development of the unconformity. The Old Red Sandstone facies, if related to the Caledonian orogeny of Baltica (Gee and Bogolepova, this volume), also
provides evidence that Severnaya Zemlya was a part of Baltica.

Palaeomagnetic data (Metelkin et al., 2000) for the Early Ordovician are not reliable, because measurements were made on samples varying in age from Mid Cambrian to Mid Ordovician. These data cannot be used as evidence for an independent North Kara Terrane in the Early Palaeozoic. Taken in relation to the other data referred to above, it is probable that the North Kara Terrane was part of Baltica from at least Vendian through Devonian time.

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NEOPROTEROZOIC ISLAND ARC MAGMATISM OF NORTHERN TAIMYR

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ABSTRACT

We present geochemistry and geochronology of extrusive and intrusive rocks from the Primet Mountain region of northern Taimyr, Arctic Russia. The intrusive complex comprises a tholeiitic olivine gabbro cumulate. U-Pb zircon ion-microprobe dating of a pegmatitic phase of the gabbro provides a crystallization age of 692 ± 10 Ma (95% confidence), in general agreement with the Sm-Nd whole rock isochron age for the gabbroic suite of 627 ± 48 Ma (2σ). Isotopic signatures (Sm-Nd and Rb-Sr) lie along the mantle array, chondrite-normalized rare earth elements (REE) are slightly light REE enriched (relative to heavy REE), and high field strength element (HFSE) concentrations are low.

Volcanic rocks include greenschist facies basalt, basaltic andesite, andesite, and dacite, with rare rhyolite, and their tuffaceous equivalents. U-Pb zircon ion-microprobe dating of a dacitic tuff yielded a crystallization age of 703 ± 11 Ma (95% confidence), within error of the age obtained for the pegmatitic gabbro. Metavolcanic isotopic signatures (Sm-Nd and Rb-Sr) lie on the mantle array, light REE enrichment (relative to normal mid-ocean ridge basalt, NMORB) is proportional to large ion lithophile element enrichment, and NMORB-like HREE are unfractionated. These combined chemical characteristics are consistent with magma genesis in an ocean island arc tectonic setting. The volcanic suite was derived from a NMORB-like mantle source, later enriched in LREE. HFSE concentrations define negative Nb and Ta anomalies.

This is the first documented 700 Ma island arc volcanism in Northern Taimyr. Though the time is roughly consistent with ophiolite genesis documented in Taimyr’s Central belt (700-750 Ma), volcanic correlatives of the same age are as yet unrecognized elsewhere in Taimyr.

INTRODUCTION

Three tectonostratigraphic domains are recognized within the south vergent fold and thrust belt of the Taimyr Peninsula in northern Siberia (Uflyand et al., 1991). These are referred to as the North, Central, and Southern Belts (Fig. 1). In the Northern Belt, there is a unique occurrence of little studied igneous rocks near the Arctic coast. The aerial distribution of these rocks is debated (Malitch et al., 1999; Bezzubtsev et al., 1986). Their age, inferred on the basis of lithologic and stratigraphic correlations, is poorly constrained but thought to be early to mid-Neoproterozoic (late Riphean). The tectonic setting associated with their formation is not known. Consequently, petrogenetic correlations with other igneous rocks of the Taimyr Peninsula are speculative.

These enigmatic rocks were the primary target of the SWEDARCTIC 2002 Taimyr expedition (Pease, 2004). Field work was undertaken in the Primet Mountain region of northern Taimyr to determine i) the regional extent of these igneous rocks, ii) their age and petrogenesis, iii) their tectonic significance, and iv) to assess their possible correlation to lithologies of the Central Belt. We present initial results from our petrogenetic and geochronologic investigations and discuss their tectonic implications. This research is part of a larger program aimed at understanding circum-Arctic allochthonous terrane distribution.

REGIONAL GEOLOGIC SETTING

Two of the three tectonostratigraphic domains recognized in Taimyr represent allochthonous terranes. The Northern and Central Belts were accreted during the late Paleozoic and Neoproterozoic, respectively (IUGS classification used throughout; Remane et al., 2002).
These Belts are separated by major south-verging thrust faults with estimated horizontal displacements between tens and several hundreds of kilometers (Urvantsev, 1949; Bezzubtsev et al., 1986).

The Southern Belt (Fig. 2) is composed of unmetamorphosed Ordovician-Permian carbonates and marine clastic sediments, and Late Permian-Early Triassic volcanogenic deposits representing the platform successions of the Siberian craton (Urvantsev, 1949; Bezzubtsev et al. 1986; Uflyand 1991). This Belt experienced Paleozoic-Mesozoic folding and thrusting which shortened and displaced originally autochthonous stratigraphy. Though displaced, these rocks are not exotic to Siberia. The intensity of folding and faulting decreases to the south (Inger et al., 1999). These sedimentary and volcano-sedimentary deposits are intruded by Permo-Triassic 245 Ma granites and syenites (Vernikovsky et al., 2003).

The Central Belt (Fig. 2) is structurally complex, lithologically diverse, and regarded as an accretionary terrane (Zonenshain and Natapov, 1987; Zonenshain et al., 1990; Uflyand et al., 1991). Paleozoic strata unconformably overlie Neoproterozoic volcanoc-sedimentary successions which include fragmented continental crust, ophiolites, and island-arc volcanic rocks (Vernikovsky et al., 1994; Pease and Vernikovsky, 2000). Neoproterozoic rocks are generally greenschist facies and associated with late Mesoproterozoic/early Neoproterozoic amphibolite facies metasedimentary units intruded by ca. 900 Ma granites (Pease et al., 2001). It has been demonstrated locally that the Neoproterozoic successions lie unconformably on these higher grade complexes. Though previous authors (e.g. Bezzubtsev et al. 1983, 1986) inferred that the latter are of probable Archean or Paleoproterozoic age and related to the Siberian craton exposed 600 km to the south, metamorphism and granite genesis of latest Grenvillian age, imply that continental fragments within the Central Belt are unrelated to the Siberian Craton (Pease et al., 2001).

Wide spread, high-grade metamorphism (600-700°C and 6-9 kbar) occurred in the Central Belt at about 600 Ma (Vernikovsky, 1995). Unmetamorphosed Neoproterozoic III (Vendian) to Early Carboniferous sediments unconformably overlie the metamorphosed basement (Bezzubtsev et al., 1986). Consequently, accretion of the Central Belt to the Siberian craton must have occurred prior to deposition of the late Neoproterozoic and younger sediments. The age of regional metamorphism associated with the Central Belt may reflect the time of this accretionary event.

The Northern Belt (Fig. 2) is dominated by rhythmically interbedded sandstones, siltstones, and pelites. These sediments, inferred from fossils (acritarchs and anabarites; Bezzubtsev et al., 1986) to be late Neoproterozoic (Vendian)-Cambrian in age, are interpreted to represent turbidites formed on a continental slope. Late Paleozoic deformation resulted in regional greenschist to amphibolite facies metamorphism of the Northern Belt. Thermobarometry suggests that temperatures associated with garnet to sillimanite grade metamorphism vary from 460 to 650°C, at pressures of 3-6.5 kb (Vernikovsky, 1995). The structure of the Northern Belt is further complicated by extensive migmatization associated with intrusion of syn- and post-
tectonic Carboniferous-Permian granites (Vernikovsky et al., 1995; Pease, 2001).

PRIMET MOUNTAIN REGION

An igneous complex(es) of gabbroic and metavolcanic rocks is exposed in the Northern Belt east of the Mikaelova Peninsula (Fig. 2). The rocks occur along the Goose River from the region of Primet Mountain to the Kara Sea (Fig. 3). The most recent geologic map (Malich et al., 1999) restricts the occurrence of these igneous rocks to east of the Goose River, while earlier mapping shows these rocks also cropping out west of the Goose River (Bezzubtsev et al., 1983). Detailed mapping and sampling west of the Goose River was not undertaken, but reconnaissance of the region confirms that volcanic rocks interbedded with sediment occur west of the river. This is in agreement with the mapping of Bezzubtsev et al. (1983) and a wider regional extent for the unit(s).

On existing maps (Malich et al., 1999; Bezzubtsev et al., 1983; 1986), there is a fault contact between rocks north of Goose River and plutonic rocks south of the river. Indeed, a low-angle fault dipping 10-15° SSE with a well-developed mylonite to ultramylonite zone is exposed within these igneous rocks along the Goose River (Fig. 3). The gently undulating mylonite zone is about three meters thick. The fault can be traced along the Goose River northwards until exposure is finally lost (Fig. 3). The direction of motion during mylonite formation could not be determined (the thin-section collection was lost during shipping from Russia). Previous workers consider most faults in the region to represent south-directed thrusting (Malich et al., 1999; Bezzubtsev et al., 1983, 1986). To the north and east, the contact between these rocks and overlying sediments (Leniven Formation) of the Northern Belt is suggested to be conformable (Bezzubtsev et al., 1986), however the contact is poorly exposed.

A strong schistosity is developed in the volcanic rocks along the contact with the gabbroic complex of Primet Mountain. To the north, from the estuary to the Kara Sea, this deformation gives way to upright, small amplitude (less than tens of meters), open folding of the volcanic rocks in which bedding and graded bedding are preserved. Flat-ramps visible along the Goose River near Basecamp (Fig. 3) and at the Kara Sea, as well as NE-SW striking fold axes/parasitic fold axes exposed at the estuary, are consistent with NNW-SSE compression. S-C fabrics, porphyroclasts, and bedding/cleavage relationships in volcanic rocks at the estuary (near VP02-033) and the Kara Sea, indicate a top-to-north sense of relative motion.

Malich et al. (1999) and Bezzubtsev et al. (1983, 1986) correlate the volcanic rocks, on the basis of lithologic and stratigraphic similarities, to the Laptev or Borzov Formations of the Central Belt. In the Shrenk-Faddey region of the Central Belt, the Laptev Formation is almost 1000 m thick with a lower section of variably metamorphosed mafic volcanic rocks (vesicular basalt, andesite, tuff, and conglomerate) and an upper section of epidotized rhyolite, tuff, ignimbrite, and minor sandstone. It is poorly constrained to early-middle Neoproterozoic (late Riphean) in age, i.e.- 1000-650 Ma. The Borzov Formation from the same area represents about 1200 m of metamorphosed basalt and andesite intruded by dikes. Its age is not known, but is also inferred on geological relationships to be late Mesoproterozoic to early Neoproterozoic (late Riphean).
<table>
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<th>CaO</th>
<th>Na$_2$O</th>
<th>K$_2$O</th>
<th>TiO$_2$</th>
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<td>0.13</td>
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**La, Ce, Pr, Nd, Sm, Eu, Gd, Tb, Dy, Ho, Er, Tm, Yb, Lu**

**Notes:** Major and trace (including rare earth) elements analyses were obtained from ACME Analytical Laboratories, Canada. Major element oxides were determined by lithium metaborate fusion using 0.2 g of whole rock powder and analyses were made using the ICP-ES technique reported in weight percent (wt.%). Trace elements (including the rare earth elements) were analysed by ICP-MS and are reported in parts per million (ppm); <, less than analytical detection limit.
ICAM IV Proceedings

35


SAMPLE RATIONALE

Samples were collected in order to determine the geodynamic setting and age of these igneous rocks. Geochemical sampling reflects the compositional diversity seen in the field, which conveniently divides into two groups: the gabbro complex of Primet Mountain and the volcanic suite north of Primet Mountain (Fig. 3). From the gabbro complex, six samples were collected for geochemical analyses and one sample for geochronology (VP02-049). From the volcanic suite, seventeen samples were collected for geochemical analyses and one sample for geochronology (VP02-071). All 23 samples were collected for geochemical analyses and one sample for geochronology (VP02-049). 

Geochemical analytical methods

Major and trace (including rare earth) elements analyses were obtained from ACME Analytical Laboratories, Canada. Major element oxides were determined by lithium metaborate fusion using 0.2 g of whole rock powder and the ICP-ES technique (results reported in weight percent). Trace elements (including rare earth elements, REE) were analysed by ICP-MS and are reported in parts per million (ppm). Calibrations were made using reference samples and international standards. Relative standard deviations are ~ 1% for SiO₂ and ~ 2% for the other major elements, except MnO and P₂O₅ (~0.01%) and K₂O (~0.005%). Relative standard deviations for trace elements are generally ~ 5%. Elements whose abundance is below analytical detection are not plotted (c.f. Table 1). The determination of structurally bound H₂O was made via loss on ignition (LOI).

Additional analyses of the six gabbro samples were obtained from Activation Laboratories (Canada) for improved detection limits of high field strength (HFS) elements. Analytical procedures are similar to those described above, but in addition to reference samples and international standards for analytical calibrations, duplicate analyses of VP02-046 were also included. Lower limits of detection were improved from 0.05 ppm to 0.005 for Tm, 0.01 to 0.002 ppm for Lu, and 0.05 to 0.01 ppm for the other heavy REE (Ho, Er, Yb). Lower limits of detection for the HFS elements were also improved (from 0.5 to 0.1 ppm for Hf, from 0.5 to 0.2 ppm Nb, and from 0.1 to 0.01 ppm for Ta). These analyses are used in the following discussion and figures.

Chemical separation and mass spectrometric analyses of Rb, Sr, Sm, and Nd were carried out at the Laboratory for Isotope Geology, Naturhistoriska Riksmuseet, Stockholm. Typically ~ 0.3-0.5 g of whole rock powder was mixed with a δ¹⁴⁷Sm/¹⁵⁰Nd spike solution and dissolved with HF and HNO₃ in Teflon bombs. Sr, Sm, and Nd were separated using standard chromatographic ion-exchange procedures. Isotopic ratios were determined on a Finnegan MAT261 mass spectrometer equipped with multiple Faraday cups.

Rb and Sr concentrations (ppm) were determined by ICP-MS (see above). Precision for Sr and Rb concentrations is ~ ± 3% (2σ) and ~ ± 5% (2σ), for ~ 50 ppm and ~ 5 ppm respectively. Samples were not spiked for Sr and the δ⁸⁷Rb/δ⁶⁸Sr ratios were calculated using the measured δ⁸⁷Sr/δ⁸⁶Sr ratios and determined concentrations of Rb and Sr. Within-run mass fractionation was corrected by normalisation to an δ⁶⁸Sr/δ⁸⁸Sr ratio of 0.1194. Error associated with δ⁸⁷Rb/δ⁶⁸Sr ratios is ~ ± 5% (2σ).

Replicate analyses of unspiked USGS BCR1 standard yielded δ⁸⁷Sr/δ⁸⁶Sr ratios indistinguishable from the accepted value of Gladney et al. (1990) and no further correction was applied. External precision of δ⁸⁷Sr/δ⁸⁶Sr ratios is ~ ± 0.01%. Initial δ⁸⁷Sr/δ⁸⁶Sr ratios (δ⁸⁷Sr/δ⁸⁶Sr) were calculated by correcting for the amount of δ⁸⁷Sr produced by δ⁸⁷Rb decay since the formation of the rock. Decay constants follow the convention of Steiger and Jäger (1977) and Lugmair and Marti (1978).

Sm and Nd concentrations (ppm) were determined by isotope dilution. Within-run mass fractionation was corrected by normalisation of Sm to a δ¹⁴⁷Sm/¹⁵²Sm ratio of 0.51686 and Nd interference was monitored at mass 146; Nd was normalised to a δ¹⁴⁶Nd/δ¹⁴⁴Nd ratio of 0.7219, with Sm interference monitored on mass 149. Error on δ¹⁴⁷Sm/¹⁴⁴Nd ratios is less than 0.1%. Error on δ¹⁴⁶Nd/¹⁴⁴Nd ratios is less than 0.002%. Replicate analyses of La Jolla standard yielded δ¹⁴⁴Nd/δ¹⁴⁴Nd ratios within error of the accepted value of 0.511854. The εNd parameter was calculated relative to CHUR (δ¹⁴⁷Nd/δ¹⁴⁴Nd).
**Table 2. Rubidium-Strontium and Samarium-Neodymium isotopic results**

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<th>$^{87}\text{Sr}/^{86}\text{Sr}$</th>
<th>2σ (%)</th>
<th>$\varepsilon_{\text{Sr}}$</th>
<th>$^{143}\text{Nd}/^{144}\text{Nd}$</th>
<th>2σ (%)</th>
<th>$^{147}\text{Sm}/^{144}\text{Nd}$</th>
<th>$\varepsilon_{\text{Nd}}$</th>
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<td>2.159</td>
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Notes: Isotopic ratios were determined on a Finnegan MAT261 mass spectrometer equipped with multiple Faraday cups. Chemical separation of Rb, Sr, Sm, and Nd used standard chromatographic ion-exchange procedures. $^{87}\text{Rb}/^{86}\text{Sr}$ ratios were calculated from their concentrations and $^{147}\text{Sm}/^{144}\text{Nd}$ ratios; consequently error in this ratio is about 5%. Decay constants follow the convention of Steiger and Jäger (1977) and Lugmair and Marti (1978). $\varepsilon_{\text{Nd}}$ parameters were calculated relative to CHUR (143Nd/144Nd = 0.512638; 147Sm/144Nd = 0.1966; Jacobsen and Wasserburg, 1984). Isotopic ratios recalculated to initial ($i$) values at time of crystallization (700 Ma).

$= 0.512638; \, ^{147}\text{Sm}^{144}\text{Nd} = 0.1966; \, \text{Jacobsen and Wasserburg, 1984}$. 

**GEOCHRONOLOGICAL ANALYTICAL METHODS**

The separation of zircons from whole rock samples was performed using conventional magnetic and heavy liquid mineral separation techniques. U-Pb analyses of zircons from these samples were performed using a Cameca IMS1270 ion-microprobe at the Nordsel Facility, Swedish Museum of Natural History, Stockholm (Table 3). Analytical procedures are similar to those described by Whitehouse et al. (1997, 1999). Geostandard 91500 with an age of 1065 Ma (Wiedenbeck et al., 1995) was used for internal U/Pb ratio calibration. Due to sufficiently low $^{204}\text{Pb}$ count rates, correction common Pb was not required. The ion beam was oval and ca. 25 μm in its long dimension. Ion-beam spot placement and size was verified using a scanning electron microprobe after ion-probe analyses and is indicated by the oval in the cathodoluminescence (CL) images. The data are plotted on inverse concordia (Tera-Wasserburg) diagrams ($^{207}\text{Pb}/^{206}\text{Pb}$ vs. $^{238}\text{U}/^{206}\text{Pb}$). Concordia ages (Ludwig, 1998) are reported at the 95% confidence level and are indicated by grey ellipses.

**RESULTS**

**PETROGRAPHY**

**Gabbro complex**

The gabbro complex of Primet Mountain is heterogeneous with regards to composition and grain size. It is generally medium to coarse grained, holocrystalline, and inequigranular, but a pegmatic phase also occurs. Finer grained varieties are hypidiomorphic to allotriomorphic granular, while coarser grained varieties display mesocumulate textures with pyroxene poikilitically enclosing both olivine and plagioclase (Fig. 4). It is an olivine gabbro according to its modal composition (average): ca. 46% sub- to anhedral twinned plagioclase (An$_{40}$) with inclusions of pyroxene and olivine, 27% anhedral interstitial clinopyroxene (good exsolution lamellae of orthopyroxene and occasionally twinned) with inclusions of plagioclase and olivine, 26% an- to subhedral olivine phenocrysts (rarely glomeroporphyritic) with inclusions of plagioclase, and 2% hornblende. Early (included) plagioclase phenocrystals are rounded, suggesting that they are resorbed. hornblende as a primary phase is brown and occurs interstitially; secondary green hornblende pseudomorphs after pyroxene. Both pyroxene and olivine lack obvious compositional zoning, while plagioclase sometimes exhibits concentric zoning. Sub- to anhedral oxides also occurs as an accessory phase(s). The gabbro complex is unfoliated and textural relationships indicate plagioclase and olivine crystallized...
developed contact coronas. When both plagioclase and pyroxene, having well-crystallographically controlled twin planes. Sericite and chlorite (20%) are present as secondary alteration products.

**Geochronology sample VP02-049, pegmatitic gabbro**

In contrast to most of the gabbro complex, the pegmatic phase is pervasively altered under low greenschist facies conditions, probably due to late stage deuteric alteration. It is predominantly plagioclase + zoisite/clinozoisite. Anhedral plagioclase (44%) has albite and carlsbad twins. Sub- to anhedral zoisite/clinozoisite (40%) has blue birefringence, yellowish pleochroism, and shows both twinning and sector zoning. The presence of primary sub- to anhedral brown hornblende (5%) is consistent with pegmatite genesis in which high water contents are associated with late-stage melts. Hornblende is partially altered to actinolite. Relict sphene (1%) is recognized on the basis of crystal habit and the exsolution of oxides along crystallographically controlled twin planes. Sericite and chlorite (20%) are present as secondary alteration products.

**Volcanic suite.** The volcanic rocks north of Primet Mountain (Fig. 3) comprise interbedded mafic to more felsic metavolcanic rocks, including well preserved green tuffaceous equivalents with fiamme up to 10 cm in length (Fig. 4). Fiamme are unwelded to welded. Sample pairs document flows of different composition found interbedded at a single location (e.g. VP02-034a and c, VP02-035a and b, VP02-040a and b).

 Petrographically, basaltic samples include plagioclase-phyric, amygdaloidal trachybasalts and plagioclase-phyric felsites and crystal-lithic tuffs. Plagioclase is variably saussuritized. Small roundish grains, now an amorphous brown alteration product, may represent thoroughly altered pyroxene/olivine phenocrysts. Spherical amygdules are composed of secondary epidote and chlorite. Tuffaceous equivalents have large amounts of devitrified glass. Dacitic samples are predominantly tuffaceous, with occasional plagioclase-phyric felsites. Normative mineralogy indicates that some of the lowest silica samples (e.g. VP02-042, VP02040b, VP02-035b) could have had olivine in the crystallising assemblage, while the majority are quartz, orthoclase, albite, anorthite, diopside, and hypersthene normative (Table 1). Approximately 30% of these samples are corundum

### Table 3. U-Th-Pb ion-microprobe analytical data

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<tr>
<th>Sample</th>
<th>U</th>
<th>Th</th>
<th>Pb</th>
<th>Th/U</th>
<th>f206</th>
<th>207Pb/206Pb ± 1σ</th>
<th>206Pb/238U ± 1σ</th>
<th>Age estimates (Ma)</th>
<th>Disc</th>
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<td>ppm</td>
<td>ppm</td>
<td>%</td>
<td></td>
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Notes: Analyses were performed on a high-mass resolution, high-sensitivity Cameca IMS 1270 ion-microprobe at the NORDSIM facility in Stockholm, Sweden. Analyses from cores denoted by “a” and from rims by “b”. Errors in age estimates are quoted at 1σ. All ages are calculated using the decay constants of Steiger and Jäger (1977). Th/U ratios calculated from 206Pb/208Pb and 206Pb/238U ages; f206(%), fraction of common Pb calculated from 206Pb; parentheses indicate no common Pb correction made. Disc. (%) refers to the degree of discordance at the 2σ level; reverse discordance is indicated by positive numbers.
normative, indicating a relative excess in alumina (sodium/potassium loss?). These rocks record equilibrium greenschist facies metamorphic conditions.

**Geochronology sample VP02-071, welded dacitic tuff**

This sample was collected for geochronology, but due to pervasive alteration it was not selected for isotopic work. Though there is no thin section for this sample, similar samples (e.g. VP02-035, VP02-036) represent green crystal, lithic, vitric tuffs. Pumice fragments are welded to various degrees, defining a eutaxitic texture. Recrystallized feldspar phenocrysts (seriate to glomeroporphyritic, twinned) are thoroughly saussuritized. The groundmass is composed of devitrified glass and/or felsite.

**GEOCHEMISTRY**

**Gabbro complex.** Primet Mountain comprises a tholeiitic complex (classification scheme of Cox et al., 1979) of olivine gabbro of limited compositional range (Fig. 5). The combination of primary amphibole and/or petrographic alteration (see Petrography) results in LOIs up to 3 wt.% (Table 1). Modal mineralogy is in good agreement with normative mineralogy (Table 1), however, reflecting the slow-cooling plutonic setting and a lack of significant open-system mobility. Gabbro chondrite normalized REE patterns are weakly light REE enriched relative to heavy REE ([La/Lu]N = 1.2-2.8), with total abundances up to 3x chondrite (Fig. 6). HREE are unfractinated ([Ho/Lu]N = 1.0-1.3). Positive Eu anomalies (Eu/Eu* 1.5-2.0) are also typical. NMORB normalized trace elements define irregular patterns (Fig. 6). LIL element abundances (e.g. Sr, Rb, Ba) peak at Sr and Rb, while K defines weak negative anomalies. With the exception of K, absolute LIL element concentrations are similar, or slightly enriched, relative to NMORB. HFSE such as Ce, P, Zr, and Ti are about an order of magnitude depleted relative to NMORB, with low absolute concentrations. Ta, Nb, and Hf concentrations (when above analytical detection limits) are also significantly depleted relative to NMORB (Table 1). The gabbro is isotopically juvenile, with measured $^{143}\text{Nd}/^{144}\text{Nd}$ of 0.51294 to 0.51301 and measured $^{87}\text{Sr}/^{86}\text{Sr}$ of 0.7026 to 0.7030 (Table 2).

**Volcanic suite.** The volcanic rocks represent a calc-alkaline suite of fine grained, massive to amygdaloidal basalt, basaltic andesite, andesite, dacite, and rare rhyolite (classification of LeMaitre et al., 2002 and Rickwood, 1989). However, these

**Figure 4. Textural characteristics of intrusive and extrusive rocks in the Premet Mountain region.** A) Photomicrograph showing cumulate texture of gabbro (VP02-050; plan polarized light). B) Same view but with crossed-polarizers. Clinopyroxene (cpx) intercumulus enclosing plagioclase (pl) and olivine (ol). Note the inclusions of rounded plagioclase within olivine. C) Out-crop view of tuffaceous andesite with flame at the estuary (VP02-033a). Notebook (for scale) is 21 cm in long dimension.
Chondrite normalized REE patterns of basaltic group samples are relatively smooth. Chondrite normalized light REE of the basaltic samples are enriched relative to heavy REE ([La/Lu]N = 1.9-2.4) and to N-MORB (Fig. 6). Heavy REE have similar concentrations to N-MORB and are unfractionated ([Ho/Lu]N = 0.9-1.2). N-MORB normalized trace element patterns of the basaltic samples define somewhat smoother patterns than the gabbroic samples (Fig. 6). Surprisingly, as LIL element mobility was noted previously, LIL elements increase smoothly in abundance to a peak at Ba (ca. 10x N-MORB). Ce and P are generally enriched relative to N-MORB with an average Ce concentrations of 28 ppm. Ta concentrations (when above analytical detection limits) are low and similar to N-MORB (Table 1). Ta, Nb, ± Ti define small negative anomalies, with an average Nb/Ta ratio of 31 and Zr/Nb of ca. 35. The basaltic samples are isotopically juvenile with respect to measured 143Nd/144Nd ratios (0.51284 to 0.51292) over a wide range of measured 87Sr/86Sr values (0.7031 to 0.7044; Table 2).

Chondrite normalized REE patterns of the dacitic group samples are relatively smooth, with enriched light REE relative to heavy REE ([La/Lu]N = 1.7-2.5) and to N-MORB (Fig. 6). Heavy REE have N-MORB-like absolute concentrations or are slightly enriched relative to N-MORB (VP02-035a) and are unfractioned ([Ho/Lu]N = 0.9-1.2). Ta, Nb, and Ti concentrations (when above analytical detection limits) are low and generally define negative anomalies with an average Nb/Ta ratio of 31 and an average Zr/Nb of ca. 37 (Table 1). Ce is generally enriched relative to N-MORB, with a average Ce concentration of 37 ppm. Dacitic samples show weak negative Eu anomalies (Eu/Eu* = 0.7 to 0.9). LIL elements increase in abundance to a peak at Ba (ca. 30x N-MORB), though as noted previously LIL element mobility may have occurred. The dacitic rocks are isotopically juvenile with respect to a narrow range of measured 143Nd/144Nd ratios (0.51281 to 0.51285) over a wider range of 87Sr/86Sr from 0.7035 to 0.7047 (Table 2).

**GEOCHRONOLOGY**

**Pegmatitic gabbro (VP02-049)**

Sampling of the gabbro complex, a mafic peraluminous melt in which zircon is not expected to be abundant, was restricted to a pegmatitic
(late, coarse grained) phase of the gabbro complex. The sample had a good yield of large (200–500μm in length), euhedral to subhedral, stubby (1:1 to 3:1 aspect ratio), inclusion free, dark pink zircon. CL images of zircon show pronounced oscillatory growth zoning with no resorption (Fig. 7). All analyses yielded concordant U-Pb ages and a concordia age of 692 ± 10 Ma (95% confidence, n=8) was calculated (Fig. 7).

Welded dacitic tuff (VP02-071)

This sample was collected for geochronology because, as one of the more evolved (felsic) volcanic samples, it was most likely to contain zircon. The sample had a good yield of stubby (2:1 aspect ratio), small (<100 μm in length), clear to light pink, sub-to euhedral zircon. Inclusions are common, but were easily avoided with the small ion beam size used during analysis. CL images show pronounced oscillatory growth zoning with no resorption (Fig. 8). Most analyses yielded concordant U-Pb ages, though one spot (no. 8) was inadvertently located on a crack and is omitted from further discussion. Using the remaining analyses, a concordia age of 703 ± 11 Ma (95% confidence, n=10) was calculated (Fig. 8).

DISCUSSION

Age of the gabbro complex and the volcanic suite

Concentric, oscillatory zoning in zircon is evidence of an igneous origin, i.e.- crystallization from a melt (Hanchar and Miller, 1993). The age of the gabbro complex and the volcanic suite overlap at the 95% confidence limit, all of the U-Pb geochronological data was generated in the same analytical session, and the same standard calibration was used for both samples. Additionally, the gabbro complex defines a Sm-Nd whole rock isochron (albeit only four points) age of 627 ± 48 (95% confidence; Fig. 9), also in general agreement with its U-Pb age. Therefore the ages of both the gabbro complex and the volcanic suite are taken to be ca. 700 Ma and to represent the time of crystallization. Depleted mantle model ages (Table 2) are in good agreement with this time of crystallization (especially for the volcanic rocks), indicating their juvenile nature and the lack of significant crustal contamination.

Mantle source of the gabbro complex

Major and trace element analyses of the gabbro complex are well-grouped and define relatively coherent trends (Figs. 5 and 6); this, combined with petrographic evidence of the extent and type of alteration (e.g.
sericitization of plagioclase), suggests that at least the less mobile isotopes, REE and HFS elements can be used to investigate the petrogenetic evolution of the gabbro complex. In this discussion, isotopic data are expressed in epsilon notation and recalculated to initial ratios at the inferred time of crystallization, i.e.- 700 Ma (Table 2).

The gabbro complex has an average initial $\varepsilon_{Nd}$ of $+8$ and an average initial $\varepsilon_{Sr}$ of $-17$. These values lie near the mantle array (DePaolo and Wasserburg, 1979; Fig. 10); they indicate a relatively juvenile, depleted character for the mantle source of the gabbro complex and little crustal contamination. This composition is consistent with mantle sources for MORB, OIB, or OIA (Staudigel et al., 1984; Arculus and Powell, 1986). Normalized gabbro REE patterns show modest LREE enrichment relative to HREE, but HREE are unfractionated. The unfractionated HREE suggests a non-garnet bearing (e.g. shallow) mantle source. They also have absolute REE concentrations less than those associated with NMORB (Fig. 6). Consequently, given that the gabbro samples preserve cumulate textures (Fig. 4), it seems likely that low absolute REE concentrations of the gabbro complex indicate that it represents a cumulate residue, rather than a melt composition. In fact, TiO$_2$/Al$_2$O$_3$ values <0.03 are similar to those for gabbroic cumulates from oceanic crust (Seifert et al., 1997; Neumann et al., 2000). Samples of the gabbro complex with greater modal abundances of plagioclase also have more pronounced Eu anomalies. Plagioclase incorporates Eu$^{3+}$ and the LREE into its crystal structure, suggesting that the positive Eu anomalies and the LREE/HREE variability of the gabbro complex may reflect heterogeneity of plagioclase modal abundances associated with plagioclase accumulation. It is, however, surprising that the geochemical analyses (majors, traces, and isotopes) are so well grouped since significant compositional variation might be expected among cumulate rocks in which different phases accumulate in different proportions. In summary, the gabbro complex likely
represents the cumulate residue of a melt generated from a shallow depleted mantle source.

**Mantle source of the volcanic suite**

Linear covariation of HFSE (Fig. 11), combined with relatively coherent REE trends (Fig. 6), indicates that the HFSE and REE, in contrast to LIL elements, are immobile and unaffected by the greenschist facies metamorphism associated with the volcanic rocks. Therefore, these elements can be used to investigate their petrogenesis. The metavolcanic suite has initial $^{143}$Nd/$^{144}$Nd ratios similar to the gabbro complex: average initial $\varepsilon_{Nd}$ of +7.5. This lies within the mantle array (Fig. 10) and is consistent with either MORB, OIB, or OIA mantle sources (Staudigel et al., 1984; Arculus and Powell, 1986). Initial $\varepsilon_{Sr}$ ratios for the volcanic suite, however, vary from ca. -9 to -20 for a relatively constant $\varepsilon_{Nd}$ value (Fig. 10). Samples with the lowest initial $\varepsilon_{Sr}$ values (ca. -20) lie on the mantle array, consistent with the Nd data. The range in $^{87}$Sr/$^{86}$Sr suggests disturbance of the Rb-Sr isotopic system. Indeed, the wide range in Rb, Ba, and Sr concentrations for a given SiO$_2$ content suggests that all of the K-group elements have been disturbed.

Both the basaltic and dacitic metavolcanic rocks show a more pronounced enrichment of LREE relative to HREE than the gabbro (Fig. 6) and the metavolcanic source has absolute LREE concentrations greater than the gabbro or NMORB. Absolute HREE concentrations in the metavolcanics, however, are more NMORB-like and unFractionated.

Small variations in the metavolcanic HREE concentrations can be explained by fractional crystallization processes involving a mineral assemblage in which REE are incompatible, increasing REE concentrations without changing REE patterns. Unfractionated metavolcanic HREE indicate a non-garnet bearing (e.g. shallow) mantle source. Enriched LREE combined with unFractionated and NMORB-like HREE suggests enrichment of a previously depleted mantle source.

Metabasaltic rocks have Ta concentrations less than Nb (when determinable; Table 1), producing pronounced negative Nb and Ta anomalies (Fig. 6). Nb, Ta and Ti anomalies are also well developed in the dactitic samples, suggesting the involvement of either i) a subducted slab component (e.g. Drummond and Defant 1990; Drummond et al., 1996), or ii) mixing with partial melts of continental crust with a residual Nb, Ta, and Ti bearing mineral phase. However, the volcanic rocks have a notably high average Nb/Ta of 27.5 – significantly higher than average continental crustal values (ca. 10) and
higher than average modern MORB ($>16.5$) (Kamber Collerson, 2000; Kamber et al., 2002). In addition, they have Ce concentrations (average 33 ppm) lower than continental crust (Wilson and Davidson, 1984). Consequently, the Nb/Ta and Ce concentrations, combined with the relatively juvenile isotopic signatures, favour a subducted slab (OIA) component over the involvement of a significant continental component in the genesis of these rocks (Wilson and Davidson, 1984; Kamber et al., 2002). The relative concentrations of immobile major and trace elements in the metavolcanic rocks of the Primet region also support an OIA genesis and these samples plot in transitional OIA fields according to most tectonic discrimination diagrams (e.g. Wood, 1980; Bailey, 1981; Pearce, 1982; Fig. 12).

The intimate association of the basaltic and dacitic rocks, their similar REE abundances, similar REE patterns, similar concentrations and ratios of HFS elements, and similar isotopic values, suggests that these rocks are co-genetic. The presence of a Daly Gap and the positive covariation of incompatible elements suggests that they are related by fractional crystallization/melting processes. In summary, the metavolcanic rocks north of Primet Mountain have been derived from a shallow, depleted NMORB-like mantle source and were subsequently enriched in LREE. This, combined with the geochemical indicators discussed above, suggests an OIA setting for these rocks.

**Towards a unified geodynamic interpretation**

The preceding discussion has shown that NMORB and/or continental crust were not involved in the genesis of the volcanic suite. Thus, the following discussion attempts to distinguish between OIB and OIA mantle sources for these rocks. A plume-type source associated with OIB derives from deeper asthenospheric mantle. Mantle plume geochemical signatures usually include strong fractionation of LREE/HREE and fractionation within the HREE – both the result of partial melting of deeper, garnet-bearing asthenospheric mantle which is less depleted than NMORB (Wilson, 1989). HFSE behave incompatibly and are preferentially concentrated (relative to NMORB) in plume (OIB) mantle sources: Zr/Nb is characteristically low in OIB ($<10$) and Nb and Ce higher than NMORB (Basaltic Volcanism Study Project, 1981). The volcanic suite of the Primet Mountain region lack these geochemical indicators common to plume/OIB mantle sources (Table 4).

The dominant source for OIA magmas is generally considered to be suprasubduction zone mantle wedge peridotite. Melting of the mantle wedge is the result of metasomatism by H$_2$O and/or CO$_2$-rich fluids derived from prograde dehydration of the subducted ocean lithosphere, with or without additional components derived from arc lithosphere and subducted sediment (Arculus, 1994; Ayers, 1998). Nb, Ta, and Ti are retained in the slab while it dehydrates and/or melts during subduction, resulting in low concentrations of Nb, Ta, and Ti in the mantle wedge overlying the dehydrating subducting slab. This generates negative Nb, Ta, and Ti anomalies in melts derived from the metasomatized mantle wedge, which are often interpreted as a geochemical fingerprint of dehydrated or melted subducted oceanic basaltic crust (e.g. Drummond and Defant 1990; Drummond et al., 1996). Consequently, metasomatism of the mantle wedge overlying the dehydrating subducting slab typically results in LILE enrichment of the mantle wedge above the slab, since LIL elements are highly mobile in the hydrous fluids derived from the dehydrating slab. The enrichment of LIL elements is approximately equal to the enrichment of LREE (Wilson, 1989).

The distinguishing geochemical characteristics of the Primet Mountain metavolcanic suite are consistent with an island arc setting (Table 4). Their isotopic and non-fractionated HREE signatures indicate a juvenile, shallow mantle source. The enriched (relative to NMORB) LREE and LIL elements, combined with low HFSE (Ce, Nb, Ta) concentrations, are consistent with extraction from the metasomatized mantle wedge. Mantle metasomatism from dehydration of the slab leads to enriched LIL elements, and dehydration or partial melting of the slab explains their very low Ta and Nb concentrations. The weak fractionation of LREE/HREE in the metabasaltic samples suggests fractional crystallization processes.

![Figure 11. Plot of ‘immobile’ elements (Hf and Zr). Correlation coefficient of the volcanic rocks indicated. Note the positive linear correlation.](image-url)
rather than actual partial melting of the slab. In summary, the metavolcanic rocks can be derived from an island arc setting. This requires that the metavolcanic rocks are derived from a metasomatized mantle wedge, enriched via slab dehydration.

Island arcs develop on a foundation of oceanic crust. They evolve through the combined effects of volcanism and plutonism, which thickens the volcanic pile. The transitional to calc-alkaline nature of the metavolcanic rocks (Fig. 5) is consistent with a thickening crust in an evolving OIA. As the crust of the OIA thickens, it cools and sinks, and seawater circulates through the crust. The interaction with seawater adds Sr to the system, enriching the mantle 
\(^{87}\text{Sr}/^{86}\text{Sr}\) signature by driving it to the right of the mantle array, towards the composition of seawater (Jacobsen and Wasserburg, 1979). Such hydrothermally altered rock provides an enriched 
\(^{87}\text{Sr}/^{86}\text{Sr}\) reservoir and can explain the 
\(^{87}\text{Sr}/^{86}\text{Sr}\) variation seen in the metavolcanic rocks (Fig. 10).

The spatial and temporal association of the Primet Mountain gabbro complex and volcanic suite suggests that they might be cogenetic. Indeed, there is geochemical evidence in support of this hypothesis (Table 4). Both units have \(\varepsilon_{\text{Nd}} = +7.5\) to +8 and unfractinated HREE, indicating that both units are derived from a shallow (non-garnet bearing) depleted mantle source. Both units have low HFSE concentrations and show slight to modest enrichment of normalized LREE/HREE (though, as expected, absolute concentrations in the gabbro cumulate are much lower), consistent with derivation from the subsequently metasomatized mantle wedge. This supports a cogenetic relationship if the gabbro complex is the cumulate residue to the melt which produced the volcanic suite. More work (electron microprobe mineral analyses and geochemical modelling) is needed, however, to establish a strong genetic link between the gabbro complex and the volcanic suite.

### Relationship of volcanic suite to the Central Belt

Most workers accept that Taimyr’s Northern Belt is thrust over the Central Belt (Vernikovsky and Venrikovskaya, 2001, and references therein). The low-angle, mylonitic fault exposed in the southern part of the study area (Fig. 3) permits this interpretation, though the details of it’s structural evolution may be debated, i.e.- the north vergent indicators in the Primet Mountain region could represent back-thrusts of a south-vergent thrust system. In this case, the Primet Mountain region may represent a tectonic widow to the underlying Central Belt. This is one explanation for lithologic correlations between volcanic rocks of the Northern and Central Belts.

The volcanic suite north of Primet Mountain with its basalt, basaltic andesite, andesite, dacite, rare rhyolite and tuffaceous equivalents, is compositionally more similar to the Laptev Formation of the Central Belt than the Borzov Formation of the Central Belt. Additionally, the Borzov Formation is characterized by the intrusions of mafic dikes, which are rare in the northern Primet area. The crystallization age of ca. 700 Ma determined for the Primet volcanic suite is also consistent with the poorly constrained middle to late Neoproterozoic (late Riphean) age of the Laptev Formation.

There are no published U-Pb ages for the Laptev Formation and recent single zircon U-Pb secondary ion mass spectrometry (SIMS) age data for volcanic and ophiolitic rocks come from the eastern part of Taimyr’s Central Belt. Here plagiogranite and gabbro associated with opholite fragments yield ages of ca. 730 to 755 Ma (Vernikovsky et al., 2004); a metarhyolite from Laptev Point thought to represent the Laptev Formation gives an age of 599 ± 4 Ma (concordia age, U-Pb zircon); a felsic metavolcanic from the Klazma River region (possibly correlative with the Kunar-Mod volcanic suite, see below) gives an age of 625 ± 4 Ma (2\(\sigma\), U-Pb zircon); a volcaniclastic sample from the Prodlonaya River region

### Table 4. Summary of geochemical data from Primet Mountain area samples

<table>
<thead>
<tr>
<th>Gabbroic</th>
<th>Basaltic</th>
<th>Dacitic</th>
</tr>
</thead>
<tbody>
<tr>
<td>tholeiitic</td>
<td>calc-alkaline</td>
<td>calc-alkaline</td>
</tr>
<tr>
<td>HREE &lt; NMORB</td>
<td>HREE &gt; NMORB</td>
<td>HREE &gt; NMORB</td>
</tr>
<tr>
<td>HREE unfractinated</td>
<td>HREE unfractinated</td>
<td>HREE unfractinated</td>
</tr>
<tr>
<td>LREE &lt; NMORB</td>
<td>LREE &gt; 2x NMORB</td>
<td>LREE &gt; 3x NMORB</td>
</tr>
<tr>
<td>Nb &lt; NMORB</td>
<td>Nb &gt; NMORB</td>
<td>Nb &gt; NMORB</td>
</tr>
<tr>
<td>Ta &lt; NMORB</td>
<td>Ta &lt; NMORB</td>
<td>Ta &lt; NMORB</td>
</tr>
<tr>
<td>Nb/Nb 16.7 (13-20)</td>
<td>Nb/Nb 34 (32-40)</td>
<td>Nb/Nb 37 (27-45)</td>
</tr>
<tr>
<td>Ce 2.5 ppm (2-4)</td>
<td>Ce 28 ppm (18.5 - 43)</td>
<td>Ce 37 ppm (27 - 40)</td>
</tr>
<tr>
<td>Eu/Eu* = 1 - 2.0</td>
<td>Eu/Eu* = 0.8 - 1.0</td>
<td>Eu/Eu* = 0.7 - 0.9</td>
</tr>
<tr>
<td>(\varepsilon_{\text{Nd}}(700 \text{ Ma}) = +8.0)</td>
<td>(\varepsilon_{\text{Nd}}(700 \text{ Ma}) = +7.5)</td>
<td>(\varepsilon_{\text{Nd}}(700 \text{ Ma}) = +7.5)</td>
</tr>
<tr>
<td>(\varepsilon_{\text{Sr}}(700 \text{ Ma}) = -17)</td>
<td>(\varepsilon_{\text{Sr}}(700 \text{ Ma}) = -11) to -20</td>
<td>(\varepsilon_{\text{Sr}}(700 \text{ Ma}) = -9) to -18</td>
</tr>
</tbody>
</table>
gives an age of $662 \pm 9$ Ma (concordia age, U-Pb zircon; Pease, unpublished data). These data from the eastern Central Belt are not easily correlated with the volcanic suite of the Primet Mountain area.

The Kunar-Mod volcanic suite in the easternmost Central Belt comprises greenschist facies tholeiitic basalts, andesites, and rhyolites. Geochemically the suite represents an island arc formed in a marginal sea basin, the result of intraoceanic subduction (Vernikovsky et al., 1996). A metarhyolite from the Kunar-Mod volcanic suite yielded a U-Pb zircon SIMS age of $627 \pm 7$ Ma (weighted mean, 95% confidence), and a metagabbroic dike representing the latest stage of mafic magmatism associated with the Kunar-Mod suite yielded a single highly discordant analysis of similar age (ca. 615 Ma; Pease and Vernikovsky, 2000). These ages, however, are too young to correlate with the rocks of the Primet Mountain region.

Consequently, the igneous rocks of the Primet Mountain region do not correlate with well-dated rocks of the Central Belt. At present, this includes the Kunar-Mod volcanic rocks, as well as the volcanic rocks exposed along Laptev Point, in the easternmost Central Belt. Future work evaluating possible correlations with the central region of the Central Belt is important for understanding the stratigraphy and tectonic evolution of Taimyr.

CONCLUSIONS
1) A tholeiitic olivine gabbro complex from the Primet Mountain region of Taimyr’s Northern Belt is ca. 700 Ma and lacks significant crustal contamination. It represents the cumulate residue from melting of a NMORB-type source.
2) A volcanic suite of metabasalts through metadacites from north of Primet Mountain are ca. 700 Ma. They represent transitional tholeiitic to calc-alkaline magmatism of an evolving island arc. Geochemical indicators suggest that they were derived from a metasomatized mantle wedge, the consequence of dehydration of subducting oceanic crust.
3) The volcanic suite and the gabbro complex may be cogenetic if the later represents the cumulate residue to the melt which generated the volcanic suite. More work is needed to confirm this hypothesis.
4) At present, these rocks have no known correlatives in Taimyr’s Central Belt. This, however, may reflect the paucity of data from the more westerly region of the Central Belt. It is important to explore possible correlatives in order to understand the stratigraphic and tectonic evolution of the region.

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Figure 12. Tectonic discrimination diagrams for basaltic compositions. A) La/10-Y/15-Nb/8 (after Cabanis and Lecolle, 1989). TH, tholeiitic; CA, calc-alkaline; Cont., continental. B) MnO-TiO2-P2O5 after Mullen (1983). MORB, mid-ocean ridge basalt; OIT, ocean island tholeiite or seamount tholeiite; OIA, ocean island alkali basalt or seamount alkali basalt; IAT, island arc tholeiite; CAB, island arc calc-alkaline basalt, and Bon, boninite. Diamonds, basaltic samples; triangles, dacitic samples; circles, gabbroic samples (for reference only). Samples with LOIs > 2 wt% represented by unfilled symbols. Analyses at the limits of detection plot on the tie-line, e.g. P2O5 and Nb.
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REFERENCES


THE TREMADOCIAN MONOPLACOPHORAN MOLLUSC KIRENGELLA FROM THE PECHORA BASIN OF NORTHWEST RUSSIA

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ABSTRACT

A limestone bed within the terrigenous sequence of the core of the Bol’shepul’skaya 1 drill-hole in the Pechora Basin, northwest Russia, has yielded the abundant Tremadocian (Early Ordovician) monoplacophorans Kirengella kultavasaensis, previously known from the southern Urals. This find facilitates stratigraphic correlation of the Pechora Basin with the southern Urals, Siberia and Laurentia, and supports a sub-tropical/tropical, rather than a sub-polar, position of Baltica closer to these continental blocks during Early Ordovician time.

INTRODUCTION

In the Pechora Basin of northeastern Baltica, the Bol’shepul’skaya-1 drill hole (Fig. 1) penetrated coarse-grained sandstones, unconformably overlying Riphean (Late Proterozoic) metamorphic rocks at a depth of 1587 m. Higher in the drill-hole (interval of 1246-1252 m), siltstones are intercalated with limestones with the abundant well-preserved monoplacophoran mollusc Kirengella kultavasaensis Doguzhaeva, 1972 and rare acrotretid brachiopods and conodonts which suggest a late Tremadocian age for these strata (Belyakova, 1988). The first Kirengella (K. ayaktica Rozov, 1968) was described from the Late Cambrian to Early Ordovician transitional beds on the Kirenga River in southern Siberia (Rozov, 1968). This species was later also found in northwestern Siberia (the Kulyumbe River) (Sokolov, 1982). Kirengella kultavasaensis was originally described from the Early Ordovician Kidryasovo Formation of the southern Urals (Doguzhaeva, 1972). Kirengella is also known from the uppermost Cambrian Minaret Formation of Antarctica and the Early Ordovician strata of the Ozark Uplift of Missouri. Kirengella has been reported from the Severnaya Zemlya Archipelago as well (Rozov, 1968), but our studies do not support this assignment.

Figure 1. The main geographic elements (A) and a generalised tectonic map (after Bogolepova and Gee, 2004) (B) of the Urals and Timan-Pechora Region, showing localities with Kirengella kultavasaensis Doguzhaeva, 1972.
In general, large monoplacophoran molluscs became one of the dominant groups of shallow-water fauna during the Late Cambrian-Early Ordovician and had a worldwide distribution. Such faunas are known from Australia, Antarctica, North China, North America, Baltica and Siberia. Non-molluscan fossils are rare in these shallow-water sediments, and monoplacophorans are therefore of considerable value for biostratigraphic correlation and palaeobiogeographic reconstruction.

**GEOLOGICAL AND STRATIGRAPHIC SETTINGS**

The metaturbidites of the Timan Range extend northeastwards beneath the Pechora Basin, and are intruded by ca 560 Ma granites (Gee et al., 2000). They are flanked to the northeast by a greenschist facies volcano-sedimentary association of calc-alkaline affinities (Belyakova and Stepanenko, 1991). A lower Palaeozoic succession transgresses this basement, with Lower Ordovician quartzites usually separating an Ordovician-Silurian carbonate platform facies from the underlying basement.

Locally, red beds separate the Ordovician basal siliciclastic formations from the volcano-sedimentary basement; these have been interpreted to be of Late Vendian to Cambrian age (Bogatsky et al., 1996). Similar lithologies have also been reported from east of the Timanides in the Izhma-Omra area, where polymict red sandstones and conglomerates, varying in from 200 to 1000 m (Belyakov, 1991), occur above the unconformity. This basal succession passes up into the (?) Middle Cambrian-Lower Ordovician Sed’iol’skaya Formation (100-200 m), which comprises pale coloured quartz sandstones with interbeds of shales (Timonin, 1998). The ages of the rocks above the unconformity are poorly constrained.

At one locality in the Pechora Basin (the Malaya Pera-11 drill-hole) shales unconformably overlying granites at 3169 m depth have yielded the acritarchs *Celtiberium dedalinum*, *Cristallinum cambriense*, *Eliasmum ilaniscum*, *Leiosphaeridia ssp.*, *Leiovalia* sp., *Microhystridium (= Heliosphaeridium) obscurum*, *Ovulum lanceolatum*, *Solisphaeridium flexipilosum* and *Timofeevia phosporitica* (Jankauskas in Belyakova, 1988). These acritarchs indicate a Cambrian age. Higher in the drill-hole, Tremadocian acrotretid brachiopods have been found (Belyakova, 1988). The Bol’shepul’skaya-1 drill-hole (Fig. 2; location 1 on Fig. 1B) penetrated a thick series of conglomerates, coarse-grained sandstones and siltstones unconformably overlying Riphean metamorphic rocks at a depth of 1587 m. At about 50 metres above these unfossiliferous sandstones and siltstones, intercalated mudstones and limestones (interval of 1246-1252 m) yield the monoplacophoran mollusc *Kirengella kultavasaensis* Doguzhaeva, 1972, rare acrotretid brachiopods and the conodonts *Acodina* aff. *lrata*, *Drepanodus* ex. gr. *lineatus*, *D*. aff. *simplex*, *Oneotodus* aff. *gracilis*, *O*. aff. *erectus*, and *Paltodus* (?) aff. *bassleri*, which suggest a late Tremadoc – early Arenig

**Figure 2.** Stratigraphic column of Precambrian-Devonian succession penetrated by the Bol’shepul’skaya 1 drill-hole (modified after L. Belyakova, unpublished data).
age (Belyakova, 1988). *Kirengella* is known from the Upper Cambrian-Lower Ordovician of Siberia (Rozov, 1968; Sokolov, 1982), the Upper Cambrian-Lower Ordovician of the Ozark Uplift of Missouri (Stinchcomb and Angeli, 2002) and the Upper Cambrian of Antarctica (Webers et al., 1992).

In contrast to the Early-Mid Cambrian microscopic helcionelloid monoplacophorans, the large tergomyan monoplacophorans became one of the dominant groups in shallow-water fauna during the Late Cambrian-Early Ordovician. Increase in high-energy shallow areas during that time stimulated a development of ecologically specialized organisms adapted to a high energy environment. One of the most successful groups was tergomyans with a large, thick, carbonate shell and a strong foot that allowed clamping against the substrate to withstand wave action in the nutrient-rich, shallow environment. One of the most successful groups was tergomyans with a large, thick, carbonate shell and a strong foot that allowed clamping against the substrate to withstand wave action in the nutrient-rich, shallow environment. Multiple muscle scars are normal condition in molluscs that hold the shell down on the substrate (Harper and Rollins, 2000). These large-size monoplacophorans were obviously slow-moving, herbivorous animals. Abundant tergomyan monoplacophorans in the Ozark region of southern Missouri and northern Arkansas are associated with stromatolites (Stinchcomb, 1980) and were probably grazing on them.

The presence of *Kirengella* in the Tremadocian of both the Pechora Basin and the southern Urals suggests that this eastern margin of Baltica was probably located much closer to the equatorial Laurentia and Siberia with the similar Tremadocian monoplacophoran fauna, than is drawn on the prevailing palaeogeographic reconstructions (McKerrow et al., 1994; Cocks and McKerrow, 1993; and Dalziel et al., 1994). Distribution of the undoubted tropical gigantic gastropod mollusc *Maclurites*, which appeared in the Arenigian, also supports the position of Baltica closer to the equator (Gubanov and Tait, 1998).

**CONCEPT OF MONOPLACOPHORA**

The Class Monoplacophora was formally proposed for Palaeozoic fossil molluscs with a bilaterally symmetrical shell by Knight (1952), though the name was informally introduced by Wenz (1940). It was assumed that the soft parts responsible for the shell formation in fossil fauna were also bilaterally symmetrical. The presence of multiple paired muscle scars on the internal surface of the shell was the main evidence for the dominant bilateral symmetry of the soft body that was later demonstrated by Lemche (1957), who described the living monoplacophoran *Neopilina galatheae* Lemche, 1957. Although the Monoplacophora was generally accepted as a class of Mollusca, there is a controversy about its content, particularly in connection with the Bellerophontoidea as well as with the Early Cambrian coiled microscopic molluscs. Moreover, some living and fossil gastropods develop a secondarily bilaterally symmetrical shell. The key monoplacophoran molluscs *Pilina unguis* (Lindström, 1880) and *Tryblidium reticulatum* Lindström, 1880 from the Silurian of Gotland, were described by Lindström (1884) as patelliform gastropods. In contrast, another well-known fossil *Metoptoma* Phillips, 1836, from the Lower Carboniferous of England, was referred to the patelloidean gastropods because it bears a horseshoe-shaped muscle scar (Knight et al., 1960). Although the Patellogastropoda is considered as the most primitive of all Gastropoda and assumed to be an ancient lineage (Lindberg et al., 1996), its geological and biological relationship with other gastropods and monoplacophoran molluscs is still very obscure. Despite the presumed primitiveness of the limpet shell form (Haszprunar, 1988), the geological record of patellogastropod molluscs is first proven in the Early Mesozoic (Peel and Horný, 1999) and the oldest known

![Figure 3. Reconstruction of muscle scars of Kirengella kultavasaensis. 1-3, BP1-06, apical view; 4-6, BP1-03, lateral view; 7-9, BP1-05, apical view.](image-url)
molluscs appeared with a coiled, not limpet-like shell (Khomentovsky et al., 1990).

Multiple paired muscle scars are considered to be the prime feature of the Class Monoplacophora. However, many fossil taxa that are generally accepted as monoplacophoran molluscs, as well as the shells of modern monoplacophorans, do not show multiple muscle scars. According to Webers et al. (1992) the presence of muscle scars reflects, in part, possession of a thick shell and, in part, life in a specialised clinging environment. Following study of living Neopilina, in which the apex lies anteriorly, all monoplacophoran molluscs with paired muscle scars have been interpreted as exogastrically coiled. In contrast, the early Cambrian genera Latouchella Cobbold, 1921, Helcionella Grabau and Shimer, 1909 and their relatives have been interpreted as endogastrically coiled (Yochelson, 1978; Geyer, 1986; 1994; and Peel and Yochelson, 1987). The exogastric and endogastric monoplacophoran molluscs were later referred by Peel (1991a, b) to the two Classes Tergomya and Helcionelloida, with monoplacophora only being employed as an informal term (Wingstrand, 1985 and Peel, 1991b).

**TERGOMYA**

The Class Tergomya has been defined as containing generally bilaterally symmetrical molluscs in which the calcareous shell is usually planispirally coiled through about half a whorl (Peel, 1991a). The shell is often cap-or spoon-shaped with an anterior apex, which may vary from sub-central to overhanging the anterior margin. The aperture is planar or slightly arched in lateral view. Muscle scars are grouped into a ring on the dorsal, supra-apical surface; the apex lies outside of this ring. Peel (1991a) emphasised that the Class Tergomya contains untorted univalved molluscs, which are coiled exogastrically. Thus, the sub-apical surface is considered to lie anteriorly while the supra-apical surface is posterior. This definition of Tergomya is very close to the undefined term Tryblidiida of Wingstrand (1985) and corresponds to the Sub-class Tergomya of Horný (1965a, b). The confirmed record of Tergomya started with the Late Cambrian Proplina Kobayashi, 1933 and can be traced up to the Recent Neopilina.

Hypseloconelloidea with an unusually tall and often septate shell are also considered to be tergomyans (Peel, 1991a). Although, tergomyan tryblidioides are considered to be the ancestral group for the other conchiferous molluscs (Lauterbach, 1983; Wingstrand,
1985), it is difficult to trace this group to the Early Cambrian where the first radiation of Mollusca probably happened. Tergomyans have considerably large size when compared with helcionelloids and their relationship to the latter group is still not clear.

SYSTEMATIC PALAEONTOLOGY
Class TERGOMYA Peel, 1991a 
Order KIRENGELLIDA Rozov, 1975 
Family KIRENGELLIDAE Starobogatov, 1970 
Genus Kirengella, Rozov, 1968

TYPE SPECIES: Kirengella ayaktchica Rozov, 1968 
DIAGNOSIS: Small to middle size (up to several cm high), bilaterally symmetrical univalve mollusc with variable low to high, cone-shaped, straight or slightly curved, moderately expanding, laterally compressed shell with concentrically arranged belt of six to eight par of muscle scars.


Kirengella kultavaensis Doguzhaeva, 1972 
(Figs 3.1-3.9, 4.1-4.21)

SYNONYMY: Kirengella kultavaensis Doguzhaeva, 1972, pl. 4, fig. 1a-c, p. 24-28.

HOLOTYPE: MGU (Moscow State University) No 172/1, Kidryasovo Formation, Lower Ordovician (Tremadocian), Ebita River, Kultavasai Creek, Southern Urals.

DIAGNOSIS: Small to middle size, bilaterally symmetrical univalve molluscs with variable, moderate to high, cone-shaped, straight or slightly curved, moderately expanding, laterally compressed shell with concentrically arranged belt of eight pairs of muscle scars. Third and fourth muscle scars as well as eighth pair of muscle scars are often fused.

MATERIAL: Over 40 specimens, often with partly preserved shell layer.

DESCRIPTION. The shell is middle size, up to 13 mm long, with length/height ratio of about 1.2 and length/width ratio about 1.3; cone-shaped, bilaterally symmetrical, barely coiled on early stage and almost straight on later stages. The apex is smooth and rounded. The protoconch is not distinct from the rest of the shell. The aperture is oval, with slightly wider supra-apical part and narrower, flat or arched upward sub-apical part. The external surface of the shell is rather smooth with faint comarginal growth lines and folds and radial striation. Some specimens (Figs. 4.1-4.8) show more rugose comarginal folds and growth lines that approach the morphology of K. ayaktehca. Muscle scars arranged in circumapical ring, with the two smallest pair of rounded scars on the supra-apical slope closer to the apex. The third to fifth (from the medium line of the supra-apical slope) pairs are tightly fused. The sixth to eighth elongated pairs are approximately the same size.

VARIATION: Shell form is very variable, ranging from low cone with length/height ratio 1.73 to high cone with length/height ratio 0.85. Most of the specimens show a slightly coiled initial shell (ca 3 – 4 mm high) following by orthoconic shell with the angle of expansion from 15° to 86° for the adult shell with the overage angle is about 45°. Initial shell is less variable and changes between 45° and 60°. Aperture outline varies from circular (length/width ratio 1) to oval (maximal length/width ratio 1.5) with overage length/width ratio is about 1.25. The specimen with maximal length/width ratio 1.9, recovered from the contact of fossiliferous limestone lens and siltstone, was probably slightly deformed due to the sediment compaction.

COMPARISON: Described specimens of K. kultavaensis are half the size of specimens from the type locality. This is probably due to the different collecting condition. Molluscs from the southern Urals were collected from a natural outcrop along the Kultavasa Creek, while the described specimens were extracted from a piece of the borehole core about 8 cm in diameter and 2 cm thick with little chance of finding any large specimen. Though most of shells are small, one cone-shaped cavity filled with calcite crystals, was probably the incomplete mould of a large shell with a size exceeding 35 mm. Due to the smaller size of molluscs from the Pechora Basin, muscle scars are weaker than in specimens from the southern Urals. They are preserved only in a few specimens, though the pattern and number of muscle scars is the same. The two pairs of smallest muscle scars (seventh and eighth pairs in Doguzhaeva, 1972 sense) on the supra-apical slope of the shell are clearly preserved only on the
largest specimen in which the third pair (sixth pair of Doguzhaeva) is obviously fused with fifth and fourth pair (Figs. 3.1-3.3). The presence of six, narrow, radial, smaller than others, excavations (traces of migrated muscle scars) on the supra-apical slope of another specimen (Figs. 3.7-3.9) shows that all of these three pairs of muscle scars were present but obscured by poor preservation in other specimens.

Comparison of *K. kultavasaensis* with other species of *Kirengella* (Fig. 5) is difficult because the determination of all erected species is based on different morphological features. The type species was studied in detail on the basis of 40 specimens with description of external and internal morphology including description of the muscle scars (Rozov, 1968). *K. kultavasaensis* is represented by four internal moulds and was established on the different from the type species pattern of muscle scars (Doguzhaeva, 1972). Other species have been established mostly by using shell morphology without detailed study of the intraspecific variation. Most of the molluscs from the Ozark Uplift, Missouri are lacking the muscle scars due to the coarse preservation of silicified material (Stinchcomb and Angeli, 2002). The very variable morphology of the described specimens overlaps with the morphology of all known species of *Kirengella*, excepting *K. pyramidalis* from the Upper Cambrian of West Antarctica (Webers et al., 1992). The latter has a sub-apical slope about half the length of the supra-apical slope, with the apex distinctly shifted toward the shorter slope.

**Figure 5. World map showing the distribution of *Kirengella*.** 1, Pechora Basin; 2, Southern Urals; 3-5, Siberia; 3, Kulyumbe River; 4, Kirenga River; 5, Chopko River; 6, Ozark Uplift; 7, Ellsworth Mountains, West Antarctica.

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STRUCTURAL, THERMAL AND RHEOLOGICAL CONTROL OF THE LATE PALEOZOIC BASINS IN EAST GREENLAND

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ABSTRACT

Traditional tectonic models of the Carboniferous-Permian basins of East Greenland emphasize rifting and half graben formation particularly along the Post Devonian Main Fault (PDMF)—a north to northeast trending fault that delimits the present-day western boundary of the basin. Structural analysis of new and published data suggests, however, that some and perhaps most of the stratigraphic offset along this fault is post-Eocene, and that key localities of the inferred rift fault display no stratigraphic offset at all. This is supported by thermal modeling of K-feldspar, which suggests that at least locally the footwall of the PDMF in Carboniferous-Early Permian time was buried and heated synchronously with formation of the exposed basin in the hanging wall. Structural data and thermal and rheological modeling further suggest that: 1) the Fjord Region Detachment Zone, a major scoop-shaped extensional fault, controlled the late Paleozoic basin geometry, and partially controlled Mesozoic basin geometries due to isostatic, topographic and compactional effects; 2) rotations and unconformities within these basins reflect folding by E–W shortening, rather than rifts guided by E–W extension and 3) the analyzed crust below the Devonian-Carboniferous basins was substantially weakened by high heat flow; and thus vulnerable to folding even low tectonic stresses (far field or internal).

INTRODUCTION

By Carboniferous time, the Caledonides formed the spine of the growing craton, which was eventually to become the supercontinent Pangea (e.g. Ziegler, 1988). After the Silurian collision of Avalonia, Baltica and Laurentia, East Greenland was situated in the centre of the craton, but nevertheless continued to deform throughout the Late Paleozoic (Fig. 1). The Carboniferous to Mesozoic geologic record of the interior of Pangea, is today mostly covered by water, so tectonic models for central East Greenland, where basins of this age are superbly exposed (Fig. 2), often strongly influence grand-scale tectonic models for the North Atlantic region (e.g. Ziegler, 1988 and Brekke et al., 2001). Widely accepted models suggest that late Paleozoic to Mesozoic sediments deposited in west-tilted half-grabens, east of the inferred 600 km long E to SE dipping Post Devonian Main Fault (PDMF), and that regional mid-Carboniferous and mid-Permian angular unconformities indicate erosion of rotated extensional fault blocks (e.g. Vischer, 1943; Koch and Haller, 1971; Haller, 1971; Surlyk, 1990; Stemmerik et al., 1992; Kreiner-Møller and Stemmerik, 2001; Surlyk, 2003; and Seidler et al., 2004) (Fig. 1).

The present study focuses on the Carboniferous to Early Permian history, but also has indirect implications for the proposed Late Permian to Mesozoic rifting in the region. First, we analyze timing of folds and offset across the faults; secondly, the burial and uplift of rocks in the hanging and footwall blocks of the PDMF dated by 40Ar/39Ar analysis; and thirdly rheological modeling is applied to discuss the structural effect of thermal history.

REGIONAL GEOLOGY

The Caledonian metamorphic domes extend along the inner fjords of East Greenland and represent the basement as well as the structural foundation for subsequent basin evolution (Koch and Haller, 1971). These metamorphic core complexes are bounded by large extensional faults (The Fjord Region Detachment Zone (FRDZ)), which juxtaposes exhumed high-grade domes against low-grade pre-Caledonian sediments (Hartz and Andresen, 1995 and Andresen et al., 1998). The collapsed hanging wall of the detachment was covered by thick continental sedimentary and bimodal volcanic rocks of Middle– to Late Devonian age (Bütler, 1959 and Friend et al., 1983), and Early and Late Carboniferous (locally Lowermost Permian) continental deposits (Vischer, 1943; Haller, 1971; Piasceki et al., 1990; and Stemmerik et al., 1992) (Figs. 1, 2, 3 and 4). These deposits were overthrust and folded around approximately N–trending axes before the Mid Permian (Bütler, 1959; Haller, 1971; and Hartz, 2000). The pre-mid Permian deposits are unconformably overlain by late Permian to Mesozoic mainly marine deposits traditionally regarded to have
Figure 1. a) Map of the geology of East Greenland, simplified from Koch & Haller (1971) MOF = Moskusoksefjord, G = Gastisdal; M = Margrethe Dal, KS = Kap Stosch. b) approximate Late Paleozoic position of East Greenland in relation to the Norwegian continental margin (reconstruction by Torsvik et al., 2001, model 1). The Norwegian structures are adapted from Braathen et al. (1999) and Osmundsen et al. (2002). c) Reference map (modified from Torsvik et al., 2001, model 1) showing Greenland's position in Pangea, and illustrating how foldbelts formed along the long amalgamated Caledonian suture.
Figure 2. Photographs of areas discussed in the text. a) Moskusoksefjord Inlier viewed towards the north. Notice that both gneisses and sediments are folded. The left summit is c. 2km above sea-level; b) West dipping Carboniferous sediments on southern Traill Ø, typically interpreted to represent block rotation above extensional faults, but also situated on the eastern flank of a syncline. View to the Northwest, the summit is approximately 1 km above sea level; c) Mid Permian angular unconformity on south-eastern Traill Ø, representing a typical example of Late Carboniferous to mid Permian block faulting, which alternatively can reflect folding. View to the northeast, the basal conglomerate is approximately 10 m thick; d) Devonian to Upper Permian angular unconformity in Koralkloft, notice that there is no evidence for faulting along the contact, and that the contact dip 27° towards the east.
Figure 3. Logs through the Upper Permian and Lower Triassic deposits in Koral Kløft at Margretheadal, modified after Kristiansen (2003). The Permian deposits show relatively thin layers of conglomerate, with clast derived from the east (figure 2d) alternating with evaporites, limestone and black shales, typical for the section (Stemmerik et al. 1998). The Triassic section shows alternating shales, silstones and sandstones interpreted as offshore turbidite deposits (Kristiansen 2003). There is no evidence for blocks of slides of the Devonian deposits situated farther west, thus considerable topography (e.g. a fault scarp) probably did not exist directly to the west.

Eocene basalts cover the Mesozoic basins, and mark the opening of the North Atlantic Ocean (Upton et al., 1995 and Jolley and Whitham, 2004). The basalts and associated sills are important structural markers as a major phase of block faulting occurred after their emplacement (e.g. Vischer, 1943; Koch and Haller, 1971; and Price and Whitham, 1997).

STRUCTURAL CONTROLS ON THE LATE PALEOZOIC BASINS

There are essentially three types of structures that have been proposed to affect the Late Paleozoic basins of East Greenland: 1) The PDMF delimits the basins towards the west, and is traditionally suggested to control development of these basins (Up. Cit.). 2) The Fjord Region Detachment Zone (FRDZ) is a major low angle extensional fault that previously has been shown to be active from the mid Silurian to Mid Carboniferous (e.g. Hartz et al., 2000). 3) North-South trending folds deform the pre mid Permian deposits gently regionally, and intensely locally (Bütler, 1959). The timing and effect of these structures are evaluated below.

Post Devonian Main Fault (PDMF)

The PDMF was originally suggested to have formed a Carboniferous-Permian half-graben in the Clavering Ø–Hold With Hope area, where sedimentary rocks of this age thicken and coarsen towards the fault (Vischer 1943) (Figs. 1, 4a). Similar geometries elsewhere (e.g. Traill Ø and Geographical Society Ø, Figs. 2b and c, and 4b) later led to extrapolation of the PDMF from Hold with Hope to Scoresby Sund (Fig. 1) (e.g. Piasecki et al. 1990; Stemmerik et al. 1992; Suryk et al. 1981, 1986; Clemmensen 1980; Piasecki et al. 1990; Suryk 1990; Stemmerik et al. 1992; Dam and Suryk 1998; Kreiner-Møller and Stemmerik 2001; Suryk, 2003; and Seidler et al., 2004).

Bütler’s (e.g. 1955) structural interpretations may be valid regionally suggesting that the PDMF have little to do with the Late Paleozoic to Early Mesozoic deposits. This notion is supported by a series of regional profiles (Fig. 4), and a detailed study along Kejser Franz Joseph Fjord. The four main arguments for this interpretation are:
Figure 4. Profiles across the PDMF and associated settings. All profiles illustrate how Devonian to lowermost Permian sediments appear folded before the Late Permian, and that there is no obvious evidence of Late Paleozoic passive rifting (e.g. block rotation) along the structure. The profile across Jameson Land (figure 3g), suggest that Late Paleozoic detachment controlled basins guided younger deposition. The simplified map marks the position of the profiles.
1) Where the PDMF interferes with Cenozoic sills, it offset the sills the same amount as the older sediments. This has been demonstrated for the Trail Ø to Ymer Ø transect by Bütler (1955). Mapping along the Jameson Land basin by Haller (1971) and Henriksen et al. (1980) suggest the same relationship for this section of the fault.

2) Syn-sedimentary faults inside terrestrial basins generally lead to erosion of the uplifted footwall block. The footwall of the PDMF, however, contains uppermost Devonian to lowermost Carboniferous sediments, either as complete outcrops, or local outliers, with no evidence of regional pre-Mesozoic erosion (Figs. 1 and 4). These deposits must have been protected from erosion by overlying Late Paleozoic or younger deposits until recently.

3) West-dipping wedges unconformable overlain by Late Paleozoic deposits east of the PDMF are often regarded as half graben hanging wall deposits, generally explained by Late Paleozoic rifting and block rotation (Fig. 2b,c) (e.g. Vischer, 1943; Kempter, 1961; Haller, 1971; Collinson, 1972; Piasecki, 1984; Surylyk et al., 1981; 1986; Larsen, 1988; Surylyk, 1990; Stemmerik et al., 1991; 1992; and Kreiner – Møller and Stemmerik, 2001). This relationship, however, only occurs on the eastern flank of synclines, and the typically setting of the direct hanging wall of the PDMF is eastwards dipping deposits (Fig. 4).

4) Late Paleozoic half graben geometries are suggested based on coarse basin fill which thins and wedges out east of the PDMF, as documented at Traill Ø (Piasecki et al. 1990; Stemmerik et al. 1992; Vigran et al. 1999) (Figs. 2b,c), or in the Jameson Land basin (Clemmensen, 1980; Surylyk et al., 1981; 1986; Larsen, 1988; Surylyk, 1990; Stemmerik et al., 1991; 1992; and Kreiner – Møller and Stemmerik, 2001). Although graben-geometries could produce these relationships, they could just as well synformal infilling, without the influence of active faults.

**Late Paleozoic folding**

Post-Devonian folding along roughly N-S trending axis is well documented in the older literature (Bütler, 1935; 1954; 1959; Maync, 1949; and Haller, 1961). Reference to the Carboniferous folding has essentially disappeared from the literature after Haller (1971). These structures are here re-emphasized and divided into three groups bounded by unconformities, which have been dated by modern stratigraphic analysis.

**Ymer Ø phase**

During the Ymer Ø phase the Late Devonian to Earliest Carboniferous Celsius Bjerg Groups were folded (Bütler, 1959) (Figs. 4 and 5). There no obvious break in sedimentation between the Late Devonian (Famennian) and Early Carboniferous (Tournaisian and Visean) deposits (Marshall et al., 1999 and Vigran et al., 1999), and all of these deposits are folded locally (Bütler, 1955; 1959; and Olsen and Larsen, 1993), indicating that folding occurred during or after Early Carboniferous times.

An angular unconformity below the Lower Westphalian marks a mid Carboniferous hiatus (Stemmerik et al., 1991), and may represent the end of the Ymer Ø phase, which is here attributed to Early Carboniferous (Tournaisian to Namurian) folding.

**Gastisdal phase**

The Gastisdal phase reflects folding of deposits in Gastisdal, north of outer Kejser Franz Joseph Fjord (Bütler, 1935 and Maync, 1949) (Figs. 4c and 5), dated as Westphalian in age (Vigran et al., 1999), and probably corresponds to the “Scoresbyland phase”, a term which Haller (1961 and 1971) used for folding of Late Carboniferous to earliest Permian deposits farther south in Jameson Land (e.g. Schuchert Dal) (Figs. 1 and 3).

The post-unconformity Late Permian sediments in the region were not affected by the folding, and overlie deeply eroded older structures, suggesting that folding ended well before Late Permian times (Figs. 2c, d, 4, 5). The phase thus includes Late Permian – Early Permian folding, and is here regionally referred to as the Gastisdal phase, in reference to Maync’s (1949) and Bütler’s (1955, 1959) detailed documentation of these folds in this region.

Collectively this suggests that the Mid to Late Devonian Hudson Land phases, the Carboniferous Ymer Ø phase (Bütler, 1959 and Olsen and Larsen, 1993) and the Gastisdal phase reflect continuous (or semi-continuous) folding, with an eastward shift in fold-locus. This interpretation suggests that the basins formed during decreasing amounts of E–W shortening from the Mid Devonian to the mid Permian.

It has been previously suggested that wrenching caused Devonian to Permian folding in East Greenland associated with left-lateral (Friend et al., 1983; Larsen and Bengaard, 1991; and Olsen, 1993) or right-lateral displacement (Coffield, 1992). Such models would infer that hundreds of kilometers long folds-hinges rotated 45° into parallelism with N-S trending strike-slip faults. Paleomagnetic studies cannot confirm such large rotations (Hartz et al., 1997). Strain partitioning between E-W shortening and potential minor deep seated parallel strike-slip faults (some of which may not cut the overlying basins), possibly mixed with minor wrenching, is thus an alternative (Hartz, 2000).

**Magrethedal Phase**

Folding of Late Paleozoic deposits is not always constrained upwards in age, and interpretation of such...
Figure 5. a) Steeply west-dipping latest Devonian sediments, from the eastern flank of a tight synform, the hingeline of which follows the proposed trace of the PDMF. View to the north. Location just east of Magrethedal, along Kejsers Franz Joseph Fjord; b) Late Devonian to Permian unconformity. The PDMF is typically mapped in the canyon (e.g. Escher 2001), but there is no evidence for faulting. Notice that the beds dip towards west in the Devonian sediments, and east in the Permian sediments, and that the Permian paleosurface extends above the Devonian sediments. Koral Kloft, just east of Margrethedal. View to the northeast. c) Faults on the western side of Obruchews Bjerg. The faults cut a complex pattern of folds in the Devonian sediments; however, the displacement of the Celsius Bjerg Group is about 700 m down to the west. The cliffs are approximately 1 km high; view to the NNE. d) Faults on the eastern side of Obruchews Bjerg. The faults displace cross-cutting folds with right-lateral motion. The Upper Permian deposits to the east could well have extended above the Devonian deposits towards west. The cliff to the left is approximately 1 km high. The view is to the NNE. e) Map of the Margrethedal area modified from Koch and Haller (1971) and Kristiansen (2003). K = Koral Kloft.
structures is thus enigmatic. The Magrethedal phase is a new term applied to differentiate Cenozoic folding from the Paleozoic events. The best example of this is the close (30°–50° dip of beds) and large amplitude folds in Paleozoic and Mesozoic sediments and Paleocene basalts and sills at Magrethedal and Kap Franklin (Figs. 4 and 5) (Bütler, 1955). Post Cretaceous folding can also be inferred from mapping in other regions (e.g. Traill Ø; Koch and Haller, 1971 and Price and Brodie, 1997) and Scoresby Land (Grasmück and Trumpy, 1969) but is generally hard to differentiate from older folding, block-rotation and thermal sagging.

**Fjord Region Detachment Zone (FRDZ)**

Several lines of evidence suggest that the FRDZ played a major role in the late Paleozoic basin formation. In the inner fjords far west of the Carboniferous-Permian basins, the footwall of the detachment was exhumed during the Early Carboniferous and possibly later (Hartz et al., 2000). Farther east, the detachment is cut by several younger faults, although the segmented structure can be traced eastward in scoop-shaped depressions, separated by up-doming footwall rocks (Hartz et al., 2001). A recent study comparing mica ages across the detachment, with detrital mica in the overlying deposits in the Hold With Hope region, suggests that the FRDZ in this region was active into the Carboniferous (Hartz et al., 2002). These isotopic data are supported by sediment distribution as Devonian to Visean deposits occurs only in the hanging wall of the detachment, whereas post-Visean deposits cover both foot and hanging wall rocks (Hartz et al., 2002; dating by Vigran et al. (1999)).

Seismic sections across the Jameson Land basin show and overall sag-geometry, where the lower deposits (Devonian to Early Permian), are cut by large extensional faults, which appear not to cross the Late Permian angular unconformity (Fig. 4g) (Larsen and Marcussen, 1992).

**THE GAUSS HALVO TRANSECT OF THE PDMF**

The northern shore of Kejser Franz Joseph Fjord represents a key transect across the PDMF, as it is the only place north of Kong Oscar Fjord where Upper Permian and younger sediments occur near the fault that supposedly controlled their deposition (Fig. 1). The transect was studied in detail in regard to the Devonian stratigraphy and folding (Kristiansen, 2003), however, several important features pertaining to the younger events along the PDMF are noted here.

The most important findings are that the contact between the Upper Permian and Devonian deposits is not a fault, and that there is no fault in Koralkløft directly west of the Permian deposits (Figs. 4c, 5). Two faults occur west of the post-Devonian outcrops and thus could serve as alternative PDMF-candidates. These faults laterally offset Early Cenozoic dikes and sills, but show no down to the east stratigraphic offset or striations (Kristiansen, 2003). Late Paleozoic to Mesozoic down-to-the east normal (rift) faulting is difficult to advocate given both the outcrop pattern along the fjord and the regional map pattern (Fig. 5) (Hartz et al., 2002), and can only be inferred by invoking the unlikely hypothesis that Carboniferous thrusting occurred along the exact same fault planes.

The Permian and younger rocks are folded during the Cenozoic Margrethedal phase and dip approximately 30° (locally 50°) to the east into a syncline, and thus extrapolate well above the Devonian deposits father west (Figs. 4c, 5d). The Late Permian and Early Triassic sediments directly west of the proposed PDMF represent calm deposition, without evidence (e.g. slumps) of major fault controlled topography to the East (Fig. 3). In addition, clasts in the basal Permian deposits are dominated by gneisses and rhyolites that are absent towards west, but common in outcrop towards the East, suggesting that these deposits were not associated with the PDMF.

It is also clear that the geology along Kejser Franz Joseph Fjord cannot reflect major Late Paleozoic extension, as the Devonian-Carboniferous Harder Bjerg formation and all older sediments are closely folded, locally with steep dips (even overturned, when post-Permian rotation is back rotated (Fig. 5a)).

Collectively these data suggest that the only faults that can be linked with the regional PDMF predominantly are post- (or syn) Early Cenozoic lateral faults (Hartz et al., 2002 and Kristiansen, 2003). Minor Late Paleozoic displacement along the same structures cannot directly be disproved, but in such a case faulting would not relate to E-W extension or rifting. There is 100 % exposure in 1–2 km high coastal cliffs for more than 50 km west of the post-Devonian deposits at Gauss Halvo, so it is considered unlikely that a hard linkage of the proposed Northern and Southern PDMF could be hidden (Fig. 1).

**THERMOCHRONOLOGICAL DATING AND MODELING**

Timing of displacement along both the PDMF and the FRDZ is among the main problem of the Late Paleozoic geology of East Greenland. This issue is tested here through 40Ar/39Ar analyses K-feldspar, which will reflect cooling and heating (exhumation and burial). Two samples were analyzed, one from the detachment-footwall, and PDMF-hanging wall (at Hold
Incremental heating of slowly cooled K-feldspar grains often shows complex release spectra (e.g. Lovera et al., 2002) and these spectra are interpreted to reflect volume diffusion of radiogenic argon at low temperatures (<400°C). Diffusion modeling of K-feldspar data is based on the theory of Lovera et al. (1989) that alkali feldspars have a distribution of diffusion domain sizes and that conventional step-heating experiments can be modeled to extract the diffusion parameters and reconstruct quantitative temperature-time histories. This technique uses Arrhenius data to construct a log (r/r₀) plot (Richter et al., 1991), which yields information about diffusion domain sizes independent of heating schedule. Once diffusion domain sizes are modeled, age spectra can be synthesized by iterative calculation of different cooling histories (Lovera, 1992) and matched to incremental heating age spectra. Temperature-time curves calculated from these model age spectra are presented below.

Figure 6. Diffusion parameters and modeling results for HWH-K1 K-feldspar collected at Stille Ø (a and b). Diffusion modeling routines are an adaptation of the Lovera (1992) autoarr program. Log R/R₀ plot is after Richter et al. (1991). c) Age spectrum and 1σ error boxes with representative model spectra (dotted lines) superimposed. Fifty spectra were obtained for each sample, but less are shown for clarity. D) Modeled cooling histories obtained using Lovera’s (1992) program autoage-free.f.
40Ar/39Ar K-feldspar Geochronology

40Ar/39Ar analyses were performed at the Massachusetts Institute of Technology on pure K-feldspar separates. Samples were encapsulated in an Al-disk shielded with Cd-foil and irradiated with flux monitor MMHb-1 hornblende (520.2 Ma, Samson and Alexander, 1987) at the McMaster Nuclear reactor in Hamilton Ontario, Canada. Incremental heating took place in a double-vacuum resistance furnace controlled to within c. 5 K using a W-Re thermocouple in contact with the outside of the crucible. Analytical blanks during the furnace experiments varied as a function of temperature, but were typically in the range of 10^{-15} to 10^{-16} moles M/e40. The uncertainties for all apparent ages include propagated contributions for analytical errors, blank errors and errors in the J-value. Further description of the lab and statistical methods can be found in Hodges et al. (1994).

Gneisses in the Post Devonian Main Fault hanging wall, at Hold With Hope.

Sample E98-HWH1 was collected from Stille Ø, a small island immediately to the north of Hold With Hope (Fig. 1). The island consist of the southernmost exposed basement dipping below the Carboniferous to Cretaceous basins of Clavering Ø and Hold With Hope. The crystalline rocks project below the Late Permian on the Island (Vischer, 1943) (Fig. 4a).

The basement rocks primarily are migmatitic pelites. K-feldspar leucosomes that have distinct compositional layering, that follows the original layering in the garnetiferous paleosome but occasionally cuts the layering in the form of N–S striking veins of mm to tens of m thickness. The gneisses are deformed by N–S trending, open folds. The dated sample is an undeformed vein of fine-grained, two-mica granite that cuts the coarse, red

Figure 7. Similar to figure 5 but for the sample collected at Moskusoksefjord.
migmatitic layering. White mica and biotite from the same sample were dated by single-grain laser fusion and have an error weighted mean age of 401.2 ± 3 and 389.6 ± 3 Ma, respectively (Hartz et al., 2002).

Incremental heating of K-feldspar from sample E98-HWH1 released 65% of the gas released at high temperature increments with a near-plateau $^{40}$Ar/$^{39}$Ar age of 335 Ma and younger (down to 310 Ma) steps duplicated during thermal recycling (Fig. 6a). The lower temperature increments show a more complex pattern where ages do not duplicate by thermal recycling, suggesting that the oldest of each temperature pair reflects excess Argon. A qualitative assessment of the age spectrum suggests well

Figure 8. Synthesis of thermal history for the Moskusoksefjord Inlier (top) and Stille Ø (bottom) based on U-Pb and $^{40}$Ar/$^{39}$Ar data (Hartz et al. 2001), apatite fission track data (Johnsen and Gallagher, 2000), stratigraphic data (Bütler, 1959 and Vigran et al., 1999) and thermal modeling presented here. Timescale used is that of Gradstein and Ogg (1999).
constrained rapid cooling near 335 Ma followed by somewhat slower cooling to 310 Ma and poorly constrained low-T cooling (Fig. 6d).

In contrast to the complicated low-T history in the age spectrum, \(^{39}\)Ar release was well modeled using three diffusion domains and an Activation Energy, \(E = 41.18\) kcal/mol (Figs. 6a-c). The modeled age spectra match the actual data well and the corresponding cooling histories show substantial cooling from 350 to 310 Ma (Fig. 6d). The subsequent thermal history is not well constrained due to complexities of the age spectrum but the models suggest modest reheating between 260 to 230 Ma, followed by renewed cooling. There is no clear evidence for younger thermal events in the age spectra, although several low-T steps yielded ages with large uncertainties between 180 – 270 Ma.

**Figure 9.** Rheological modeling of the lithospheric strength (a) and thermal profiles (b) across Late Paleozoic non-Caledonized Laurentian craton in the foreland of the Caledonian thrusts (continuous line); regions between the FRDZ and the hot thin crust and collapse basins (thick dotted line); and settings represented the hot collapsed crust, and overlying Late Paleozoic basins (thin dotted line); c) simplified profile across East Greenland based on Hartz et al. (2001) and Schlindwein and Jokat (2000).
Gneisses in the Post Devonian Main Fault footwall, at Moskusoksefjord.

The Moskusoksefjord Inlier is a remarkable basement high situated within the Devonian basins, 25 km west of the PDMF (Fig. 1). The inlier forms an asymmetrical antiform (Fig. 2a), flanked by folded Devonian deposits with internal, high angle unconformities at both flanks (east and west) (Bütler, 1959). Structural mapping and fault kinematic analysis show that the inlier formed by synsedimentary folding (N–S trending axis) and extensional faulting (E-W trending faults), reflecting mid to late Devonian N–S extension and E–W shortening (Hartz, 2000).

The Caledonian thermal history was studied through U-Pb monazite analysis and shows that prograde metamorphism was ongoing by c. 435 Ma, and resulted in anatectic melting by c. 427 Ma. Cooling of the high-grade rocks is dated by \(^{40}\text{Ar}/^{39}\text{Ar}\) white mica analysis to c. 413 Ma (Hartz et al., 2001). Exhumation of the basement gneisses is not well constrained. The gneisses are onlapped by the Vilddal and Kap Kolthoff Groups, which are non-fossiliferous around the inlier but suggested to be Mid to early Late Devonian based on regional correlations (Friend et al., 1983 and Olsen and Larsen, 1993). The succession is topped by the fossiliferous Kap Graah Group dated as Famennian (Friend et al., 1983 and Olsen and Larsen, 1993). Burial of the Inlier thus took place during Late Devonian.

\(^{40}\text{Ar}/^{39}\text{Ar}\) incremental heating dated a sample of K-feldspar from red granitic gneiss. Most of the gas (65%) was released at high-temperature increments with a near-plateau at between 375 to 360 Ma. The lower temperature increments show a more complex history with steps climbing from 290 to 350 Ma from low to moderate temperatures (Fig. 7).

Diffusion modeling fits the \(^{39}\text{Ar}\) release well with seven diffusion domains and \(E = 54.17 \text{ kcal/mol}\) (Figs. 7a-c). The modeled age spectra match the actual data well, and the corresponding cooling histories show substantial cooling from 370 to 360 Ma (Fig. 7d). The sample was then modestly reheated until about 300 to 290 Ma. Renewed cooling began sometime between 280 to 250 Ma (Fig. 6d). There is little evidence for younger thermal events in the age spectra.

Tectonothermal synthesis

Modeling the thermal history from both samples indicates marked Late Devonian cooling suggesting exhumation, but the cooling is far more drastic in the sample from Moskusoksefjord Inlier (Fig. 8). Generally, this corroborates the onlap of Late Devonian sediments upon the inlier (Fig. 2a), while the sample from the Hold With Hope region that was still hot and thus at greater depth.

Thermal modeling for the Moskusoksefjord sample shows a short-lived temperature minimum at the Devonian–Carboniferous boundary reflecting the end of Devonian onlap. The modeled high temperatures at the Devonian–Carboniferous transition when the sample was at shallow crustal levels probably reflect that exhumation and burial was very rapid, and that the region had a very high geothermal gradient (Fig. 8). The high heat flow is also documented by the nearby synchronous bimodal volcanic activity.

In contrast to the synchronous cooling during the Late Devonian exhumation of both samples, the Carboniferous thermal histories of the two samples are opposite: The Moskusoksefjord Inlier was heated 200°C or more, while Stille Ø cooled 200°C or more (Fig. 8). Clearly, the rocks in Moskusoksefjord were buried, while the rocks at Stille Ø was exhumed. This has several implications for basin geometries. The most prominent is the 200°C heating of the Moskusoksefjord Inlier that, even with a high geothermal gradient would require deep late Devonian to Early Permian burial west (in the footwall of) the PDMF, thereby corroborating evidence for a lack of major Late Paleozoic stratigraphic offset along the structure. A cover of about three km of Lower to lower Upper Carboniferous deposits, similar to the Traill Ø Group exposed father east in the hanging wall of the PDMF (Vigran et al., 1999), would explain the thermal history modeled for that age, and it is thus suggested that there was no major difference in Carboniferous deposition across the PDMF.

The Stille Ø area, cooled rapidly through the Early Carboniferous, and somewhat slower during the Late Carboniferous (Fig. 8). The sample is collected in the FRDZ footwall, and the shift in cooling rate coincides with the previously suggested cessation of displacement along the detachment (Hartz et al., 2000). By the latest Carboniferous, the area had cooled below the resolution of Ar-dating. This corresponds well with formation of the Late Carboniferous (Westphalian) basins unconformable on gneisses farther west on Clavering Ø.

Thermal information beyond the Early Permian is not well constrained by data, however the modeling suggests that the samples from Moskusoksefjord cooled sometime after the mid Permian, and that the Stille Ø sample was slightly heated during the Late Permian to mid Triassic. There is no local sedimentary data to compare to the Moskusoksefjord data, but the modeling fits the regional observation of a mid Permian erosion and hiatus. The Stille Ø data also corroborates the mid–Permian hiatus and subsequent cover of thick Upper Permian to Lower Triassic cover recorded just.
south of Stille Ø, and furthermore suggests that these Triassic basins continued to thicken into the mid Triassic. These data also agree with apatite fission track data from the nearby Clavering Ø, suggesting substantial cooling (erosion) of the Carboniferous basins during the mid to late Permian, succeeded by late Permian to Triassic burial (Johnson and Gallagher, 2000) (Fig. 8).

RHEOLOGICAL MODELING

One of the most striking findings of the thermochronological modeling is the high temperatures at times when there was a relatively thin cover above the dated rocks (Fig. 8). This is best illustrated in the Moskusoksefjord Inlier gneisses that remained relatively hot, even when the rocks was at shallow crustal levels during exhumation and subsequent burial.

The geothermal gradient can not directly be determined from these data, however, there is no reason to assume that the now eroded Carboniferous cover above the Inlier was thicker that the sections measured elsewhere (i.e. east of the PDMF, Vigran et al., 1999), and the mid Carboniferous geothermal gradient thus locally appears to have been as high as 40°C/km (or even higher). The high geothermal gradient was not regional, because K-feldspar dated from clasts in pre-Caledonian sediments west of the Devonian basins shows no sign of major Paleozoic heating (Hartz et al., 2001). In these cool regions, Devonian deposits rest relatively flat on folded pre-Caledonian deposits (Bütler, 1955 and Koch and Haller, 1971), implying that Late Paleozoic folding in these areas was minor, and that a high geothermal gradient and folding are spatially associated. The effect of such localized zones of high heat flow is investigated through modeling of the lithospheric strength.

Methods

In order to model the lithospheric strength we used a simple three-layered model to construct geothermal gradients and strength envelopes. It is beyond the scope of this study to treat the concepts of these models in detail, and we refer the interested reader to Turcotte and Schubert (1982), Kirby (1983), and Rannali (1995).

In order to limit the number of variables and isolate the thermal effects, the lithospheric model is made as simple as possible. Modeling, thus, does not quantitatively the exact lithospheric strength at each considered setting, but rather illustrate variations in differential stress dependent on the thermal gradient implementing a constant strain rate. The lithologies are kept constant as follows: supracrustal basins (young or old) as quartzite; an upper crust as granite; a lower mafic crust of diabase; and a lithospheric mantle of dunite (Fig. 9a). Rheological parameters of these lithologies are listed in Table 1. The temperature at the base of the lithosphere is set to be 1300°C and the surface temperature is assumed to be 0°C (Fig. 9b). Thicknesses of each layer in the substrata are adapted from the crustal scale profiles across central East Greenland based on seismic and gravity data (Schlindwein and Jokat, 2000) (Fig. 9c).

Modeled results

Three different scenarios were modeled: 1) an old thick craton in the foreland of the orogen with crustal and lithospheric thickness of 45 and 120 km respectively, 2) a section through the thinned but cold crust between the FRDZ and the Devonian basins with a crustal and lithospheric thicknesses of 30 and 120 km, and 3) a region with late Paleozoic basins and high heat flow with crustal and lithospheric thickness of 35 and 80 km (Fig. 9).

Non-orogenic regions

The foreland of the East Greenland Caledonides is little affected by Late Paleozoic tectonism. Following the profile presented by (Schlindwein and Jokat, 2000) and regional mapping (Koch and Haller 1971), the modeled crust is divided into 1 km of sediment cover, 26 km of upper crust, and 18 km of lower crust, placing the moho at 45 km (Fig. 9a). The low geothermal gradient is low and the surface heat flow is set to 45 mW·m⁻² (Fig. 9b). Modeling of the lithospheric strength illustrates a strong lithospheric mantle and upper crust and a relatively weaker lower crust.

Cold orogen scenario

East Greenland’s crust is considerably thinned east of the FRDZ (Schlindwein and Jokat, 2000) (Fig. 9c). In this scenario it has a 2 km thick sediment cover and a 28 km thick granitic crust overlying the lithospheric mantle. The surface heat flow is set to 60 mW·m⁻². The low geothermal gradient (Fig. 9b), is verified by the pre-Caledonian Ar- ages from this setting (Hartz et al.,

<table>
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<th>Density (kg·m⁻³)</th>
<th>Conductivity (W·m⁻¹·K⁻¹)</th>
<th>Capacity (J·kg⁻¹·K⁻¹)</th>
<th>Heat Production (W·m⁻³)</th>
<th>Skin depth (km)</th>
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</table>

Table1. Thermal parameter values used in the rheological modelling

Hartz et al.
Modeling of the lithospheric strength illustrates a weakened lithosphere with respect to the craton, although the upper crust and lithospheric mantle still show considerable strength (Fig. 9a).

**Hot, late Paleozoic basins**

The crust below the Devonian basins of East Greenland is modeled with a 4 km thick sediment cover and a 31 km thick granitic crust overlying the lithospheric mantle (Fig. 9c). The surface heat flow is set to 70 mW·m$^{-2}$, and geothermal gradient is high (Fig. 9b). The heat significantly weakens the middle crust (Fig. 9a), despite that the strength profile is modeled with the present day relatively thin middle crust probably caused by Late Paleozoic or later thinning. If we had attempted to restore this thinning, the thicker middle crust would represent even more extreme weaknesses in the heated regions.

**Late Paleozoic intracratonic soft-zones**

The rheological modeling of the three different lithospheric settings illustrates how elevated heat flow below the late Paleozoic basins strongly influences strength properties. The simplified models presented here should not be taken as absolute measures of lithospheric strength, particularly since rock properties may vary locally. The profiles nevertheless illustrate how strength varies between the modeled sections as a function of crustal properties and thermal gradient.

The most prominent finding of the rheological modeling is that the Late Paleozoic basins formed above thermally weakened lithosphere. This has direct influence on tectonic models of the area. Firstly the findings show that any (extensional and contractional) deformation of the rocks would require relatively little stress, thereby explaining why impressive structures such as the large folds (Fig. 2a), could form in the middle of a craton (Fig. 1). Secondly, the data suggest that the middle crust would have flowed in response to tectonic forces. Such mid crustal flow will directly explain mid crustal thinning below the deep late Paleozoic basins (Figs. 4g, 9c).

**DISCUSSION AND REGIONAL TECTONIC IMPLICATIONS**

The new and published structural, stratigraphic, thermochronological and rheological data discussed above both independently and collectively call for revised post-Caledonian basin models.

The geology along Kejser Franz Joseph Fjord (Figs. 2, 5) and regional profiles (Fig. 4) suggest that the northern and southern segment of the PDMF is primarilay, and perhaps even exclusively, a Cenozoic fault, thereby making major Carboniferous riftting difficult to reconcile. This observation is corroborated by the thermal data in the PDMF footwall (Fig. 8), interpreted to illustrate thick Late Paleozoic burial, similar to that observed in the PDMF hanging wall today, and Carboniferous to mid Permian folds (e.g. Margreathedal (Fig. 5)), in areas that were supposedly extended at that time. The history of the central PDMF is more difficult to evaluate due to the limited exposure and lack of nearby younger sediments to evaluate timing of the offset. The linkage of such fault with the Mesozoic faults at Hold With Hope and farther North (Vischer, 1943; Surlyk, 1990; Kelly et al., 1998; Larsen et al., 2001; and Hartz et al., 2002) is presently unclear (Fig. 1).

In contrast to the PDMF the FRDZ was active the Carboniferous time (Hartz et al., 2000), inherited topography, thermal subsidence and basin compaction in the hanging wall generated accumulation space from Permian and well into Mesozoic time.

The Carboniferous and Early Permian appears to have been a period of considerable E-W shortening of the uppermost crust, as for example illustrated by the intense folding of the Devonian–Carboniferous deposits in Margreathedal (Fig. 5a) (Bütler, 1959 and Kristiansen, 2003) and elsewhere (e.g. Haller, 1971 and Kempter, 1961). Such high contractual strains appears surprising so long after the mainly Late Silurian orogeny in the region (e.g. Haller, 1971), yet the thermally weakened middle crust documented here suggest that the dramatic strain reflects weak rocks rather than dramatic stress. At present it is unclear if these low stresses reflects a remnant of kinematic energy from the pre-Caledonian WNW drive of Baltica relative to Laurentia, or an far-field effect of the synchronous Variscan and Uralian orogenies (e.g. Ziegler, 1988) transferred across the stiff Baltic plate (Fig. 1c).

East Greenland basin models are traditionally extrapolated into the submerged continental margins between Greenland and Norway (e.g. Brekke et al., 2001). Our general notion is that such extrapolations should be done with caution, as parallel regions; even within East Greenland, tell a highly variable story. We nevertheless note that the detachment-controlled Late Paleozoic basins and synchronous folding may be a regional feature, since rapid Carboniferous cooling and exhumation of the FRDZ footwall match the thermal history of the detachment footwall in the Scandinavian Caledonides (Eide et al., 1999), and that the overlying Devonian basins also there are folded (Osmundsen et al., 2000), broadly synchronous with major Late Paleozoic to Earliest Mesozoic detachment faulting in and orogen parallel extension north-central Norway (Broaathen et al., 1999 and Osmundsen et al., 2002). East Greenland detachments may thus directly compare or even link up with the synchronous detachments on
and offshore Norway (Fig. 1b). The thermal weakening of a middle crust here documented through K-feldspar analysis may, elsewhere along the Caledonides, be diagnosed by Late Paleozoic folding evidence from the British Isles to Svalbard (Vogt, 1936 and Harland, 1965).

CONCLUSIONS

East Greenland's Late Paleozoic history has long been reflected in rift models emphasizing the PDMF (Vischer, 1943; Surlyk, 1990; and Stemmerik et al., 1992). This view is challenged by thermal and field data contrasting Late Paleozoic rifting along the PDMF, and instead emphasizes thinning of a weak middle crust along the extensional detachment as the main control of the Carboniferous to middle Permian basins. It is furthermore suggested that subsequent thermal subsidence of the regions above the thinned crust, and compaction of the thick Late Paleozoic deposits in the detachment-hanging wall, continued to control the Late Permian to early Mesozoic basin geometries.

In the new model, tilting of Late Carboniferous to Early Permian sediments is explained by folding (E–W shortening) rather than by block rotation of half grabens by E–W extension. The extreme thermal weakening modeled for the crust below the Late Paleozoic basins could have facilitated this folding, suggesting that the impressive strain recorded in the folds does not necessarily reflect large stresses. The source for these intracratonic post-orogenic stresses is presently unclear, but could have resulted either from the remnant kinematic energy Baltica's pre-Caledonian westwards drift, or far-field stress reflecting Variscan and Uralian orogenies acting on the margins of the cold and stiff Baltic plate.

The contrasting views of tectonic models clearly reflect that the Late Paleozoic geology of central East Greenland still is enigmatic, and that none of these models should be extrapolated uncritically regionally (e.g. to the offshore areas) prior to further structural and thermochronological investigations.

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THE NARES STRAIT DEBATE: IMPLICATIONS FROM THE STRUCTURAL EVOLUTION OF PALAEOGENE OUTLIERS IN EASTERN ELLESMERE ISLAND

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ABSTRACT

Palaeogene sediments are exposed on the Judge Daly Promontory (east Ellesmere Island) in elongate SW-NE striking fault-bounded basins cutting folded Early Palaeozoic strata. These basins are characterized by broad synclines cut by wrench faults. Fold axes strike NE-SW at an acute angle to the border faults indicating left-lateral shear during contraction. Weak deformation in the interior of the outliers contrasts with locally intense deformation at the border faults. Subordinate faults and shear planes in the footwall are predominantly SW-NE oriented NW-dipping reverse faults associated with SSW-NNE to SW-NE striking sinistral and WNW-ESE striking dextral faults. The degree of deformation varies along-strike the northwestern faults, and there is also a contrast in degree of deformation between the northwestern and southeastern border faults. A two-stage structural evolution is suggested: (1) The fault pattern on the Judge Daly Promontory is the result of sinistral strike-slip faulting, during which the Palaeogene Judge Daly basin formed by a pull-apart mechanism (2) the strike-slip faults were reactivated during oblique compression. Sinistral motion along the Wegener Fault is linked with coeval spreading in the Labrador Sea, Baffin Bay and Eurasian Basin. Plate reconstructions for Late Cretaceous to Oligocene times (e.g. Srivastava, 1985) infer true sea-floor spreading in the Labrador Sea and Baffin Bay and large-scale left-lateral strike-slip displacement in Nares Strait in the range of 120 to 320 km (e.g. Johnson and Srivastava, 1982; Srivastava and Tapscott, 1986). Other models are based on on-shore geological data which imply only 0-50 km displacements (Dawes and Kerr, 1982b; Higgins et al., 1982; Johnson and Srivastava, 1982; Newman, 1982; Higgins and Soper, 1989; Okulitch et al., 1990).

INTRODUCTION

The Nares Strait is a narrow marine channel that separates eastern Ellesmere Island from north Greenland (Fig. 1). The “Nares Strait problem” represents a debate about the existence and magnitude of left-lateral movements along the proposed Wegener Fault within this seaway (see Dawes and Kerr, 1982a). Its plate tectonic significance has been a continuing topic of debate because it is unclear whether the Wegener Fault represents a major plate boundary and transform fault, and how much (sinistral) lateral motion occurred along this fault during the opening of the Labrador Sea, Baffin Bay and Eurasian Basin. Plate reconstructions for Late Cretaceous to Oligocene times (e.g. Srivastava, 1985) infer true sea-floor spreading in the Labrador Sea and Baffin Bay and large-scale left-lateral strike-slip displacement in Nares Strait in the range of 120 to 320 km (e.g. Johnson and Srivastava, 1982; Srivastava and Tapscott, 1986). Other models are based on on-shore geological data which imply only 0-50 km displacements (Dawes and Kerr, 1982b; Higgins et al., 1982; Johnson and Srivastava, 1982; Newman, 1982; Higgins and Soper, 1989; Okulitch et al., 1990).

GEOLOGIC SETTING

In the Canadian Arctic Islands, Cenozoic Eurekan tectonics overprinted the Early Carboniferous Ellesmerian Fold Belt, and only a few regions with preserved post-Ellesmerian strata exist which permit the investigation of unequivocal Eurekan structures. The study of Cenozoic tectonics on the Judge Daly Promontory, adjacent to the Nares Strait, can provide constraints on the geometry, evolution and kinematics of Eurekan deformation. These data may also shed light on the nature and kinematics of offshore structures as well as on the amount of displacements along the faults.

Scattered exposures of Palaeogene sediments occur on Judge Daly Promontory at the northeastern coast of Ellesmere Island (Fig. 1), these are the Cape Back outlier, the Pavvy River outlier and the “Triangle Tertiary”. Palaeogene deposits are exposed in basins bounded by SW-NE striking faults.

STRATIGRAPHY

Previous stratigraphic divisions for Palaeogene strata on Judge Daly Promontory (e.g. Miall, 1982, 1986, 1991) comprise four units or members with the conglomerates resting with an angular unconformity on the three lower members. Field studies, however, do not confirm the supposed age relationships, and imply that lateral as well as vertical interfingering of the different lithologies occurs. Therefore, in this study a division of Palaeogene strata into five facies units (A to E) is preferred (Fig. 2), in part corresponding to the
members sensu Miall (1982), though without implications for age relationships. Only facies unit A could represent a relative time marker for correlation because it consists predominantly of detritus derived from basaltic volcanics. These deposits occur in all Palaeogene outcrops on the Judge Daly Promontory and can even be traced offshore to the Kennedy Channel in seismic sections (Damaske and Oakey, 2003). The deposits were probably derived from the same volcanic source, and thus mark a distinct erosional and depositional event, probably of the same age. Facies units C, D and E are restricted to the Cape Back outlier. Coarse conglomerates and breccias, up to several hundreds-of-metres thick, of facies unit C are interpreted as proximal deposits adjacent to the basin border faults. The conglomerate fan thins out to the NW and greatest thickness at the Cape Back Fault suggests sedimentary input from a source located to the SE. Deposition was triggered by syn-depositional normal or extensional oblique-slip faulting.

Figure 1. Map of the Judge Daly Promontory with main faults and position of the Palaeogene exposures. Location of Judge Daly Promontory study area in eastern Ellesmere Island adjacent to Nares Strait.
thrusting deposition of the conglomerates is indicated, for example, by interfingering of the conglomerates with sandstones of facies unit D as well as by basal conglomerate beds overlain by facies units E and A.

**PALAEOGENE (EUREKAN) STRUCTURES**

The Palaeogene exposures are preserved as SW-NE trending fault-bounded basins (Fig. 1) striking sub-parallel or slightly oblique to Kennedy Channel and Nares Strait. The border faults are major lineaments which can be traced in the Palaeozoic units. They strike at an acute angle to the pre-Cenozoic (i.e. Ellesmerian) structures in the Palaeozoic units and cut km-scale upright Ellesmerian folds.

The Cape Back outlier is bounded to the NW by the Mt. Ross Fault which can be regarded as the northern continuation of the Rawlings Bay Thrust (Mayr and de Vries, 1982) at Cape Lawrence. In its northern part hangingwall it exposes Palaeozoic platform carbonates on Palaeogene sediments, whereas the hangingwall in its southern parts contains Palaeozoic deep water deposits (Danish River Formation). The southeastern fault is named the Cape Back Fault. Thrusting fault is the northwestern border fault of the Pavy River outlier and mostly Early Paleozoic basin sediments are thrust on Palaeogene sediments. The steep Judge Daly Fault in the SE separates Palaeogene siliciclastics from Palaeozoic platform deposits.

The Triangle Tertiary is exposed in the acute angle between two converging sinistral faults striking NNE-SSW and NE-SW (Fig. 1).

**Basin interiors**

The overall geometry of the Cape Back outlier is a broad open syncline (Fig. 3). The fold axis strikes NE-SW to ENE-WSW at an acute angle to the border faults. The syncline is truncated and displaced by a series of approximately NNE-SSW oriented sinistral faults (Fig. 3), which do not cut the border faults. This contrasts with some WNW-ENE trending faults which are believed to be younger structures, although they may also be old faults previously associated with the main deformation of the Palaeogene strata that were later reactivated. Subordinate folds are also developed, such as a NW-vergent anticline on a tens of m scale to the NW of the syncline in the northern parts of the Cape Back outlier (Fig. 3), and subordinate synclines in the southern part of the Cape Back outlier (Fig. 3, section A) and within conglomerates southeast of Mt. Ross (Fig. 3, section C). The Palaeogene beds were tilted to the NW and SE at the northwestern and southeastern bounding faults, respectively, leading to drag folds due to oblique reverse movements on these faults.

The deformation of the Palaeogene deposits in the interior of the Pavy River outlier is remarkably weak. The beds are nearly undeformed apart from local folding and tilting, and shear planes are rare. The most conspicuous structure is the broad syncline exposed in the central eastern part of the outcrop (Fig. 4). The bedding planes dip to the NW and SE and the SW-NE oriented fold axes strike slightly oblique to the border faults.

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**Figure 2. Top: Stratigraphic table of Palaeogene sediments in Ellesmere Island by Miall (1982, 1991). Bottom: Subdivision of Palaeogene strata in the Cape Back area into 5 facies units and their stratigraphic relationship used in this paper. Facies units A and B also occur in the Pavy River outlier and "Triangle Tertiary".**

<table>
<thead>
<tr>
<th>Stratigraphic unit</th>
<th>Lithology</th>
<th>facies assemblage</th>
</tr>
</thead>
<tbody>
<tr>
<td>Member Te4</td>
<td>Cape Lawrence Fm</td>
<td>boulder conglomerate</td>
</tr>
<tr>
<td>Member Te3</td>
<td>Cape Back Fm</td>
<td>dark grey siltstone, thin argillaceous units, beds of sideritic mudstone</td>
</tr>
<tr>
<td>Member Te2</td>
<td></td>
<td>thinly interbedded, fine- to very fine-grained sandstone, siltstone and mudstone</td>
</tr>
<tr>
<td>Member Te1</td>
<td>Mokka Fjord Fm</td>
<td>cyclic succession of laminated dark fine- to coarse-grained sandstone and pebbly sandstone, local thin conglomerate beds</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>unit A</th>
<th>unit B</th>
<th>unit C</th>
<th>unit D</th>
<th>unit E</th>
</tr>
</thead>
<tbody>
<tr>
<td>dark brown sandstone, medium to coarse grained with pebbles, black shale; proximal beds are conglomeratic; pebble: &gt; 90% basal volcanic</td>
<td>light brown fine to medium grained sandstone intercalated with siltstone and dark shale planar cross bedding</td>
<td>coarse conglomerates (boulder size) with calcareous matrix, intercalated by thin laminar sandstone beds</td>
<td>fossiliferous yellowish siltstone and sandstone, well preserved leaf impressions, grey siltstone, subordinate sandstone and conglomerate, ripple marks</td>
<td>greenish and yellowish calcareous sandstones and siltstones, fine grained portions contain plant debris; abundant sedimentary structures like ripple marks,</td>
</tr>
<tr>
<td>facies alluvial fan (conglomerates) to fluvial channel deposits, alluvial plain (shales)</td>
<td>facies distal alluvial plain</td>
<td>facies proximal alluvial fans</td>
<td>facies shallow lake, sandstones/conglomerates = proximal beds</td>
<td>facies shallow lake</td>
</tr>
</tbody>
</table>

Å Te 1 (Miall 1992) Å Te 2 (Miall 1992) Å Te 4 (Miall 1992) Å Te 3 (Miall 1992) —
faults (Fig. 4). The syncline is truncated and displaced by several WSW-ENE oriented faults. The central part is characterized by SW-dipping beds. In general, increased folding and shearing of the beds can be observed to the NW with decreasing distance to the Archer Fjord Fault. For example, a subordinate NW-vergent anticline on the scale of tens of meters occurs with its fold axis striking at an acute angle oblique to the Archer Fjord Fault. Folding adjacent to the southeastern Judge Daly Fault can also be observed, though these folds are open and shearing is less pronounced than in the NW.

The bedding of Palaeogene sediments in the Triangle Tertiary generally dips to the WNW to NW and locally to the E. Axes of minor folds as well as bedding/cleavage intersection lineations are oriented SW-NE to SSW-NNE and mostly plunge gently to the NNE.

Besides open folding and tilting, the deformation of the Palaeogene deposits in the basin interiors is moderate suggesting that deformation is concentrated mainly at the border fault zones. There exists, however, a striking contrast in the degree of deformation of Palaeogene strata between the northwestern and southeastern border faults.

Deformation at the southeastern border faults (Cape Back and Judge Daly faults)

Palaeogene beds in the footwall of the southeastern border faults are mostly just tilted and dip to the NW. Conglomerates in the southern parts of the Cape Back outlier nearby the fault are cut by 35°-45° NW-dipping thrust-faults, and polished surfaces and slickensides on planes sub-parallel to bedding show both reverse and normal dip-slip movements. Steep faults and shear planes and en échelon veins are

Figure 3. Geological map of the Cape Back outlier and cross sections illustrating large-scale open folding of the Palaeogene sediments and dissection of the folds by steep faults. Section C shows the fault-bounded block at Mt. Ross and the disconformity between the Palaeozoic carbonates and Palaeogene sediments. Sections C and D illustrate the interfingering of the conglomerates with sandstones.
Figure 4. Geological map of the Pavy River Tertiary outlier and lower hemisphere stereoplots of structural data.

common. Extension joints filled with calcite striking perpendicular to the bedding-parallel thrust faults opened perpendicular to the shortening direction and thus are related to thrusting. Deformation of the finer-grained Palaeogene sediments is characterized by closely spaced shear planes, faults and conjugate sets of shears. Small-scale thrust-faults probably related to Palaeogene faulting also occur in the Palaeozoic limestones. En échelon arrangement and orientation of associated shear planes and veins, together with slickensides indicate movements of the hangingwall to the SE.

Steep to overturned bedding planes of Palaeozoic units east of the Pavy River outlier are cut by the Judge Daly Fault at an acute angle. A number of splay faults enclose lens-shaped blocks of Palaeozoic limestone cut by closely spaced, sub-vertical shear planes. Most of these planes strike sub-parallel to the Judge Daly Fault and dip to the WNW or ENE. Other sets of planes with an average dip of 45-50° to the SW to WSW and of 60-70° to the NE are also well developed. The Palaeogene strata adjacent to the Judge Daly Fault, show a surprising lack of deformation. The WNW-dipping bedding planes are usually truncated by extension.
joints oriented sub-perpendicular to $s_0$. Weakly developed SE and NE dipping shear planes are scattered and form a conjugate set.

**Deformation at the northwestern border faults (Mt. Ross and Archer fjord faults)**

In contrast to rather weak deformation at the southeastern border faults shales and sandstones adjacent to the northwestern border fault are locally intensely folded, sheared and fractured. However, deformation within the Palaeogene in the footwall of the northwestern faults is highly variable along-strike.

In some sections of the Mt. Ross Fault, Palaeogene black shales interbedded with dm-thick sandstone beds directly below the fault are folded and intensely sheared with development of cleavage planes parallel to the axial planes of the small-scale folds. Some sandstone layers seem to be stretched and boudinaged during shearing and are surrounded by anastomosing shear planes. A conjugate pair of shear planes, C1 and C2, is developed in the Palaeogene beds, and a third set of planes C3 related to and associated with C1 and C2, but steeper inclined, led to the development of sigmoidal fault-bounded phacoids and horses. Slickensides indicate sinistral oblique reverse movements of the hangingwall. Palaeozoic limestones in the hangingwall and Palaeogene sediments in the footwall are truncated both by NNE-SSW and WNW-ESE sinistral shear planes and faults, interpreted as synthetic P and Riedel shears, as well as by WNW-ESE striking planes with slickenside lineations indicating dextral movements of the hangingwall, which thus could represent antithetic Riedel shears $R'$. This contrasts with the deformation of the Palaeogene only a few km to the N and S, where Palaeogene strata below the fault are nearly undeformed except for tilting of the beds and some minor shearing.

Extreme along-strike variations in the degree of deformation can be observed at the Archer Fjord Fault (Fig. 5). Along the northeastern segments on the Archer Fjord fault, deformation of Palaeogene sediments is weak and displays only tilting, local folding and formation of scattered shear planes. A broad fault zone is exposed a few km further to the SW. The fault cuts Palaeozoic calcareous siltstones leading to cataclastic brecciation of more competent layers. A coarse Palaeogene conglomerate occurs to the SE with depositional contact. Overturned and intensely sheared Palaeogene dark shales follow to the south separated by a narrow valley (Fig. 5b). They are in turn overlain by brown sandstones. A splay fault of the Archer Fjord Fault probably runs through this valley separating different Palaeogene units. The Archer Fjord Fault thus splays into subordinate faults, which preferentially use incompetent Palaeozoic siltstones as well as Palaeogene...
Figure 6. Analysis of faults and shear planes in Palaeogene and Palaeozoic rocks at the border faults as well as on the Judge Daly Promontory with sinistral synthetic and dextral antithetic Riedel shears. Note the predominance of reverse faults at the Cape Back and Judge Daly faults. Upper left: Palaeostress analysis of the fault pattern at the northwestern and southeastern border faults both showing NW-SE orientation of the compressional axis ($\sigma_1$) and vertical tensional axis ($\sigma_3$) indicating a predominant compressional rather than strike-slip regime.
shales as glide horizons. The Paleozoic siltstones strike further to the SW and the fault remains within these units. In the southwestern part of the Pavy River outlier, the degree of deformation of the Palaeogene as well as the Paleozoic units further increases. Dark cherts of the Hazen Formation are thrust on Palaeogene shales and sandstones (Fig. 5c). The contact with the Palaeogene is represented by two discrete parallel fault planes separated by a few cm dipping with 35°-40° to the WNW. The deformation of the Palaeogene units in the fault zone is characterized by closely spaced sigmoidal shear planes, small-scale faults and duplex structures on a dm-scale. Palaeogene sandstones directly below the thrust are folded into a recumbent syncline on a tens-of-metres scale passing into an anticline further below. The fold axes strike NW-SE at an acute angle to the Archer Fjord Fault (Fig. 4) indicating a component of sinistral shear during thrusting.

**DISCUSSION AND CONCLUSIONS**

**Palaeogene deformation**

Some major characteristics of Palaeogene Eurekan deformation can be summarized:

- Shearing and folding of the Palaeogene occurs close to the border faults, but only minor deformation is observed away from these faults.
- There is a striking contrast in degree of deformation between the southeastern border faults (Cape Back and Judge Daly faults) with shearing and gentle folding and closely spaced shear planes and strong folding and faulting at the northwestern border faults (Archer Fjord and Mt. Ross faults).
- Marked along-strike variations in deformation occur especially at the northwestern border faults, ranging from only tilting of the beds to strong shearing and large-scale recumbent folding.
- The border faults have steep dips ranging from 60° to verticality. The dip of a single fault is laterally not consistent.
- The degree of deformation in the footwall of the major faults depends on the dip of the fault, i.e. tight folding and shearing is related to reverse movements of the hangingwall which is mechanically only possible at sites with less steep dip, whereas folding is weaker and deformation is associated with subordinate faults and shear planes at steep to subvertical fault segments where lateral motion predominates.

- SW-NE oriented fold axes within the Palaeogene exposures strike oblique to the SSW-NNE to SW-NE trending border faults indicating a left-lateral component of shear during contraction.
- Along-strike variations of the deformation can also be attributed to local bending of the fault or to splay faulting resulting in along-strike partitioning of the overall shear in either local restraining bends forming broad shear zones predominated by strike-slip motion or areas of mainly convergent deformation.

The overall structural pattern suggests a connected system of strike-slip and (oblique) reverse faults combined with folding in the Palaeogene sediments related to movements along the faults. Oblique fold axes, steep faults, lateral along-strike variations of structural style and intensity of deformation as well as splay faulting are characteristic features of transcurrent shear zones. The geometry of the shear plane array and the fault kinematics in both the Paleozoic units and in the Palaeogene beds is similar in all of the Palaeogene exposures on the Judge Daly Promontory: Fault plane analysis of the entire data set as well as smaller sub-sets (Fig. 6) show that the most prominent and common fault trends are WSW-ENE and SW-NE to SSW-NNE and show sinistral displacement. They can be interpreted as synthetic P shears and Riedel shears which commonly strike at acute angles to the main SW-NE oriented border faults (= principal displacement zones, PDZ). Some NWW-SSE to N-S striking dextral faults which could represent antithetic R‘ Riedel shears fit well in an overall pattern of left-lateral shear. Reverse faults oriented sub-parallel or at an acute angle to the PDZ show a convergent component during fault activity so that the fault array suggests left-lateral transpression or NWW-SSE to NW-SE oblique compression. A strike-slip fault system fits well with marked strike-slip fault array on the Judge Daly Promontory (Mayr and de Vries, 1982, v. Gosen et al., in press).

However, some unfavorably oriented shear planes or faults (e.g. N-S striking reverse faults) as well as different shear senses on planes with the same orientation indicate a more complex evolution. The marked contrast in deformation, with strikingly stronger deformation in the NW and clear predominance of reverse faults in the fault plane set which in many cases are subparallel to the PDZ, are inconsistent with an overall strike-slip regime but instead point to SE-directed thrusting and associated folding during NW-SE compression. On the other hand, a model of orthogonal compression could not explain the fault plane array in both the Palaeogene exposures.
Figure 7. Eurekan deformational episodes in Ellesmere Island related to the regional plate-tectonic framework. (a) Stage 1: Paleocene to Early Eocene sinistral strike-slip displacement along the Wegener Fault and associated faults. This episode includes Palaeogene pull-apart basin formation. The large Judge Daly basin was located between the Wegener Fault in the SE and the Archer Fjord Fault (or a parallel fault) in the NW and formed due to transtension at releasing bends or fault oversteps. Spreading in the Labrador Sea and Baffin Bay leads to northward drift of Greenland and sinistral relative movements between Greenland and Ellesmere Island. (b) Stage 2: Ceased spreading in the Labrador Sea and Baffin Bay and increasing spreading in the North Atlantic result in dextral relative movements of Greenland and Svalbard (De Geer Fracture Zone) and (slightly sinistral oblique) convergence between Greenland ad Ellesmere Island (Wegener Fault). Pre-existing faults inherited from stage 1 were reactivated. The SE to SSE directed shortening results in oblique thrusting or transpression along NE-SW striking reactivated faults in the Judge Daly Promontory and to thrust-dominated deformation in southern Ellesmere Island along W-E striking faults. The shortening direction in the Judge Daly Promontory inferred from fault-plane analysis in the Palaeogene and adjacent Palaeozoic units corresponds well will the directions derived from the plate kinematics. Plate tectonic reconstructions after Srivastava, 1985; Rowley and Lottes, 1988; Roest and Srivastava, 1989; Müller and Spielhagen, 1990).
and the Paleozoic in the hangingwall which can be linked with the fault pattern on the Judge Daly Promontory. Moreover, it cannot explain the steep dip of the faults because thrusting would imply formation of and movement on gently dipping faults for mechanical reasons. Note that most of the reverse faults show an oblique slip sense of displacement (Fig. 6).

A two-stage structural evolution during the Palaeogene is suggested here to solve these inconsistencies.

**Structural evolution**

(1) **Left-lateral strike-slip tectonics**

The large-scale fault pattern on the Judge Daly Promontory reflects sinistral strike-slip deformation (Mayr and de Vries, 1982, v. Gosen et al., in press). Wrench faulting also caused the steep dip of the faults and produced the array fault lines including the Palaeogene border faults. Therefore, the Eurekan structural evolution of the Judge Daly Promontory comprises an episode of left-lateral strike-slip tectonics. This episode also includes the Palaeogene basin formation and deposition. The Paleocene Judge Daly Basin was oriented SW-NE and extended about 100 km parallel to Nares Strait (Miall, 1982). Structural and sedimentological data suggest a pull-apart basin formation at releasing bends or sidesteps of the strike-slip faults (Saalmann et al., 2005). The southeastern border fault was located in Nares Strait and would follow the inferred Wegener Fault whereas the Archer Fjord Fault could represent the northwestern border fault (Fig. 7). Wrench faulting might have evolved with time and progressive motion from predominantly transtensional to dominant transpressional (see below).

(2) **NW-SE shortening and compression**

Strike-slip tectonics was followed by contractual deformation resulting in deformation of the Palaeogene deposits and their preservation of in the form of narrow fault-bounded relics (Eurekan deformation s.s.) (Fig. 7). Palaeostress analysis of faults and shear planes indicate a NW-SE orientation of the maximum principal stress (Fig. 6) with the intermediate stress axis being horizontal so that the regional stress field can be attributed mainly to a compressional phase. NW-SE compression caused a reactivation of the major pre-existing faults on Judge Daly Promontory. However, because of their steep dip, thrust displacement was mechanically unfavourable. Depending on their more or less pronounced oblique orientation to the maximum principal stress and due to their steepness, the convergence was therefore largely accommodated by oblique sinistral reverse movements, explaining the observed strong lateral component along these faults. Hence, the orientation and geometry of pre-existing faults strongly controlled the style of deformation.

**Plate tectonics**

Eurekan deformation is related to the plate tectonic events that resulted in the opening of the North Atlantic, Labrador Sea and Baffin Bay. Rifting and sea-floor spreading in these areas and the resulting plate motions caused changes in the tectonic regimes at the plate margins during Palaeogene times. One of the major problems in resolving Eurekan tectonics, especially regarding the Nares Strait debate, is to get plate-tectonic models in agreement with on-shore geologic and structural observation.

The plate boundary between North America and Greenland is inferred to lie in Labrador Sea and Davis Strait and could be traced to Baffin Bay with a northern continuation located in Nares Strait which represents a left-lateral transform that links Baffin Bay with the Nansen-Gakkel Ridge in the Eurasian Basin (Jackson et al., 1979; Srivastava, 1985; Srivastava and Tapscott, 1986; Rowley and Lottes, 1988). Seafloor spreading in the Labrador Sea already began at chron 34 in late Cretaceous time (Roest and Srivastava, 1989) or started in the early Paleocene at chron 27 (Chalmers and Laursen, 1995) and stopped in Oligocene times at chron 13 (Roest and Srivastava, 1989) or later at about chron 7 (Rowley and Lottes, 1988).

Activation of the Wegener Fault in Nares Strait as a major transform is linked with coeval sea-floor spreading in Baffin Bay and at the Nansen-Gakkel Ridge. The nature and age of the crust in Baffin Bay is controversial. Seafloor spreading in Baffin Bay is assumed to have started either in Paleocene time (chron 24) (Jackson et al., 1979; Menzies, 1982; Srivastava, 1985; Geoffroy et al., 2001) or earlier at chron 27 (Chalmers and Laursen, 1995) or already at chron 30 or 32 (Rowley and Lottes, 1988). Spreading ceased at about Eocene time (chron 20) (Menzies, 1982; Srivastava, 1985). It is also possible that true sea-floor spreading in Baffin Bay occurred during a much shorter time interval between chron 24 and 23 (Jackson et al., 1979). Spreading at the Nansen-Gakkel Ridge started in early Eocene time at chron 24 (Vogt et al., 1979; Jackson et al., 1982), or possibly earlier (Srivastava, 1985; Srivastava and Tapscott, 1986).

Focusing on the Judge Daly Promontory, Palaeogene basin formation could be related to transtensional movements along the Wegener Fault. Sinistral relative displacement between Greenland and Ellesmere Island along the Wegener Fault caused by sea-floor spreading in Baffin Bay and Labrador Sea and only minor spreading in the North Atlantic between chron 26 and 24 (i.e. ca. 61-56 Ma) would result initially in a transtensional regime in this area which
Implications for the Nares Strait problem

Plate tectonic models estimate sinistral strike slip in Nares Strait of at least 125 km between chron 25 and 21 derived from the trajectories and exceed displacements in the range of only 0-25 km inferred from stratigraphic, structural, geological or geophysical markers. Mayr and de Vries (1982) reconstructed sinistral displacement along the Judge Daly Fault Zone of the order of ~19 km. The amount of displacement along the other fault lines could not be estimated due to the lack of reliable marker horizons (Mayr and de Vries, 1982, v. Gosen et al., in press), major faults exposed on Judge Daly Promontory (e.g. Archer Fjord Fault, Mt. Ross Fault, Cape Back Fault) may show comparable displacements, however. Thus, an accommodation of 100-120 km left-lateral shear by a network of faults is possible and compatible with the amount of sinistral displacement demanded by plate-tectonic models. Hence, strike-slip as well as later reactivation and reverse dip-slip were probably distributed through a broad fault zone with displacements along individual major faults in the range of 10-30 km. This could explain the continuity of stratigraphic and structural markers as well as of geophysical features across Nares Strait.

Minor displacements along the Wegener Fault argue against a distinct and well defined sinistral transform plate boundary. Moreover, Eurekan structures are not developed on the Greenland side (Okulitch et al., 1990; Okulitch and Trettin, 1991) but restricted to the west of Nares Strait. We suggest that the Wegener Fault represents the easternmost major lineament of a number of faults exposed in Ellesmere Island which were active already during Palaeogene basin formation. The Wegener Fault therefore does not represent the principal displacement zone which controls the overall structure of the belt. The present day fault line probably represents the frontal thrust fault of the Eurekan orogen (Damaske and Oakey, 2003). The plate boundary – if it does exist – is therefore partitioned in a bundle of faults truncating continental crust on- and off-shore (Saalmann et al., 2005). If the southern continuation of Wegener Fault is not clear. Geological and geophysical investigations in the southern parts of Nares Strait show no evidence for lateral displacements (Funck et al., 2003; Oakey and Damaske, 2003) but rather continuity of the Greenland and Canadian Precambrian basement (Harrison et al., 2003). Hence, the fault cannot be traced to the southern Nares Strait and northern Baffin Bay but probably curves or jumps off to the west in the area south of Dobbin Bay. Future on-shore geological studies in these areas are necessary to test this interpretation.

ACKNOWLEDGEMENTS

We would like to thank all members of the CASE 5 and CASE 6 groups for discussions and support in the field and afterwards. We acknowledge financial support for the project by the German Federal Ministry of Education and research as a project of the Canadian-German cooperation in science and technology (WTZ). We thank Robert A. Scott and Dennis Thurston for thorough review of the manuscript.

REFERENCES


ABSTRACT
This study of the Laptev, East Siberian, and Chukchi sea shelves concentrates mainly on the specific structure of the sedimentary cover and briefly discusses age and structure of the basement. An elaboration of the tectonic model for the entire Russian eastern Arctic shelf is accompanied by discussions of the existing views on problems which have been hampered by absence of deep offshore wells. The most relevant information was obtained from deep seismic profiling, which provides data on thicknesses and unconformities within the sedimentary cover, tectonic structures, and structure of the basement surface. But without well control the age and nature of the basement and sedimentary cover remain thus far, speculative and will have to be verified by future research activities.

INTRODUCTION
The Laptev, East Siberian, and Chukchi sea shelves (Fig. 1) and their transition to deep water basins are still in a very early stage of geological exploration. All existing bathymetric, magnetic, gravimetric and geologic data, including shallow wells in the New Siberian Islands area and deep wells in the U.S. sector of the Chukchi Sea shelf, were recently summarized in the process of compiling the State Geological Map of the Russian Federation at a scale of 1:1,000,000. The work was carried out by the Geological Map Division of the All Russian Scientific Research Institute for Geology and Mineral Resources of the Worlds Oceans (VNIIOkeangeologia) during the last 7 years and includes a special geologic map index that unifies the entire offshore area. The level of earth science exploration varies drastically from one area to another, and even in the relatively better studied parts it remains insufficient for unambiguous characterization of geological structure and history, especially in places where subbottom geology is imaged exclusively on the basis of geophysical evidence and cannot be reliably correlated with island outcrops.

DATA SETS
This paper is based mainly on the State Geological Map at a scale of 1:1,000,000, every sheet of which presents a separate geoinformational system (GIS) layer. All existing bathymetric, magnetic, gravimetric and geologic data, including shallow wells in the New Siberian Islands area and deep wells in the U.S. sector of the Chukchi Sea shelf, were used to construct maps sheets. A set of maps consists of tectonic, deep-structural, bedrock, Quaternary and bottom sediment, and geomorphologic maps. Onshore portions of the digitized database come from geological mapping of the eastern Arctic mainland and islands by the Russian Geological Survey in the middle of the last century. Geological investigations of the offshore area are rather poor. They include bottom sampling, piston cores and rare shallow seismic profiles. Although several shallow boreholes were collected around the New Siberian Islands and exploration wells were drilled on the U.S. Alaska Shelf, the major source of offshore information used in this study is from various geophysical methods of investigations by the Marine Arctic Geological Enterprise (MAGE) of Murmansk (Sekretov, 1998; 1999a; 1999b; 1999c), Laboratory of Regional Geodynamics (LARGE) from Moscow (Drachev et al., 1998), and joint research efforts by the oil company “Sevmorneftegeofizika” of Murmansk and the Geological Institute BGR of Hannover, Germany (Roeser et al., 1995; Hinz et al., 1998; Franke et al., 1999), and by the oil company “Dalmorneftegeofizika” of Yuzhno-Sakhalinsk, Russia and Halliburton (USA).

STRATIGRAPHY OF THE SEDIMENTARY COVER
Two large provinces are distinguished on the Russian eastern Arctic Shelf. These two provinces in the western Chukchi and East-Siberian seas are generally underlain by basement of either late Mesozoic or Caledonian age. The boundary between these is complex in some areas, with the former overprinting the latter, but can generally be traced from Cape Lisburne in western Alaska, northwest along the northern margin of the Herald Arch to100-150 km north of Wrangel Island, and on across the East-Siberian Sea to Vil`kistsky Island, southwest of the De-Long Archipelago. These provinces are generally characterized by unique sedimentary cover with distinct stratigraphic ages and are traced using seismic reflection data and borehole sections from northern Alaska and the Chukchi Shelf where American investigators (Grantz et al., 1975; 1982; and 1990 and Thurston and Theiss, 1987) identified two major types of sedimentary cover with good confidence. They divide sedimentary cover into 2 main sequences, the Ellesmerian and Brookian, separated by the regional Lower Cretaceous unconformity (LCU) at the base of the Barremian...
Stage. Later researchers have distinguished a third sequence in the U.S. Chukchi Sea—the Rift Sequence (Sherwood et al., 2004) which we have not mapped but may be equivalent in part to our J-K1h sequence in Figure 3. This sequence underlies the Brookian and represents the initial stage of opening of the Canada Basin. The second major sequence is the Late Devonian to Early Cretaceous Ellesmerian Sequence, which is generally found in the province with Caledonian basement and had source rocks located to the north in the area of the present day Arctic Ocean. The Early Cretaceous to Cenozoic Brookian Sequence sediments are distributed in both provinces, covering the Ellesmerian Sequence in one, and in the other, covering late Mesozoic folded basement rocks and comprising the whole volume of the sedimentary cover.

Ellesmerian Sequence strata can be distinguished on seismic records in the North Chukchi Basin where their thickness reaches 7-8 km, and where the total thickness of the sedimentary cover is no less than 20 km. Grantz et al. (1975 and Grantz et al., 1982) had reported the presence of older, possibly Franklinian or “Eoellesmerian” Sequence in the North Chukchi Basin at the base of the Ellesmerian cover but it is not clear what age they are. We believe his data show lower Ellesmerian Endicott Group strata in some grabens that may be as thick as 7 km, in the North Chukchi Basin. Figure 2 shows our interpretation of a seismic profile collected by “Dalmorneftegeofizika” Enterprise which divides the sedimentary cover of the North Chukchi depression into 7 seismic units separated by reflectors Ch-I-VII, and a correlation to the reflectors mapped by Thurston and Theiss (1987). Older, basal sedimentary cover exists north of the North Chukchi Basin, as evidenced from lower Paleozoic rocks dredged on the continental slope, on the Mendeleev Ridge, and on the steep eastern slope of the Northwind Escarpment, where a continuous stratigraphic section beginning with Cambrian rocks was reported by Grantz et al. (1998). Similar data were obtained by Russian investigators from the southern part of the Mendeleev Ridge, where they dredged moderately lithified carbonate-tuffrogenous rocks exhibiting the occurrence of kaolinitic cement in sandstones, and were

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**Figure 1.** Scheme of geographical objects. **Legend:** 1 — shelf break; 2 — boundaries of geographical objects; 3 — small islands. **Map: Objects:** 1 – Khatanga bay, 2 – Bolshoy Begichev island, 3 – Tsvetkov cape, 4 – Olenek bay, 5 – Tiksi bay, 6 – Buor-Khaya bay, 7 – Yano-Indigirskaya lowland, 8 – Bol’shoy Lyakhovsky island, 9 – Kotel’ny island, 10 – Zemlya Bunge, 11 – Faddeevsky island, 12 – New Siberia island, 13 – Vilykistsky island, 14 – Zhokhov island, 15 – Benetta island, 16 – Henrietta island, 17 – Lomonosov Ridge, 18 – Mendeleev Rise, 19 – Chukchi Basin, 20 – Wrangel island.
Figure 2. Structure of the sedimentary cover of the Southern side of the North-Chukchi depression with interpreted reflectors (Ch-I) – (Ch-VII). The age and correlation with the seismic horizons of the U.S. Chukchi Sea are shown at right: uBU (Upper Brookian Unconformity), mBU (Mid-Brookian Unconformity), LCU (Lower Cretaceous Unconformity), PU (Permian Unconformity), and EU (Ellesmerian Unconformity). Seismic profile fragment SC-90-01 of “Dalmorneftegeofizika” Enterprise.
Figure 3. Fragment of seismic profile (89001, LARGE) approximately 170 km east-southeast of New Siberian Island. The boundary between Epi Late Mesozoic (to the left of the fault) and Epi-Caledonian (to the right of the fault) regions of the East Siberian Sea shelf. B-I reflector is equivalent to the Lower Cretaceous Unconformity (LCU) and $PZ_3$-$K_1$ is equivalent to the Ellesmerian Sequence.
paleontologically dated as Upper Silurian to Lower Permian in age (Kaban’kov et al., 2001).

On the East-Siberian shelf, Ellesmerian Sequence strata were recognized on a LARGE seismic profile (Fig. 3) 170 km east-northeast of New Siberian Island (Drachev et al., 2001). Here, two reflectors “A” and “B” are distinguished beneath reflector “B-1” (LCU). The strata between these reflectors thin and are truncated by reflector B-1 to the north (toward the De-Long rise), but to the south, they increase to 7 km in thickness. Besides the increase in thickness, there is an increase in deformation of the strata, which gradually becomes more intense until it is manifested as a rather sharp transition to acoustic basement. We believe this transition of Ellesmerian Sequence to the folded state marks the boundary between late Mesozoic and Caledonian basement provinces.

On New Siberian Island, drilling revealed late Mesozoic basement (folded Jurassic terrigenous rocks) beneath Pliocene–Quaternary sediments. Magnetic anomalies over New Siberian Island are similar to the anomalies seen from the cassiterite-bearing granites of Bolshoy Lyahovsky Island. (See Figure 1 Number 8). Placer cassiterite found on New Siberian Island, and grains of molybdnite and sphalerite in the basal Pliocene layers overlying the deformed Jurassic rocks suggest the presence of stanniferous granites in the basement.

Caledonian basement is exposed on Henrietta Island (See Figure 1, Number 16) where outcrops of folded volcanic and elastic rocks, bearing sills, dykes, and sheets of basalts, andesite-basalts, and porphyritic diorite occur (Vinogradov et al., 1974). Basalts and porphyritic diorite dated by the potassium–argon method are 310–450 ma and porphyritic diorite dated using the argon–argon method are in the 400–440 ma interval. Fragments of gneissic, granitic, and quartzitic rocks and schists in gneisite of Henrietta Island are evidence that the Caledonian fold basin zones in the De-Long rise province include blocks of older consolidation. More evidence of these older rocks is the presence of flat-lying Cambrian and Ordovician strata on Benetta Island (See Figure 1, Number 15), but these lower Paleozoic rocks cannot be correlated to seismic reflection signatures typical of sedimentary cover on nearby marine profiles.

The Ellesmerian Sequence is divided by American authors (Grantz et al., 1975; 1990 and Thurston and Theiss, 1987) into two parts: the Lower Ellesmerian and the Upper Ellesmerian Sequences, separated by a Permian unconformity (PU) at the base of Upper Permian strata. In the wells drilled on the Alaska coast, the highest stratigraphic interval is the Upper Ellesmerian Sequence, and the “PU” reflector marks the acoustic basement. However, in deep depressions where the “PU” horizon separates subparallel reflectors of the Lower and Upper Ellesmerian sequences, it may be lost. This is what we believe happened on the seismic profile in Thurston and Theiss (1987) Plate 5, published before the wells in the Chukchi Sea were drilled. Within the Ellesmerian Sequence below the Lisburne Group, there is a well expressed reflector identified as an unconformity between Endicott Group (D1-C1) and Lisburne Group (C2-3). On the southern slope of De-Long rise, the presence of Lisburne Group rocks is confirmed by dredge samples that contained fragments of siliceous limestone with C2-3 fauna in Neogene volcanics of Zhokhov Island.

The Brookian Sequence (Barremian Stage to Cenozoic) is divided by American scientists into Lower Brookian (K₁br–al) and Upper Brookian (K₂–KZ) sequences. In our interpretation, based on seismic data analysis across the whole eastern Arctic Shelf (from the Laptev to Chukchi seas), we propose dividing the stratigraphic section into Cretaceous (K₁br–K₁) and Cenozoic parts. The lower sequence is much thicker and is characterized by numerous plastic and disjunctive deformations as a result of fragmented topography of the basement surface. The upper sequence is a continuous but less thick mantle type deposit. It is seismically transparent and is not disturbed by syndepositional deformations.

**Discussion**

It has been previously argued that the sedimentary cover of the Lavtev shelf ranged in age from Proterozoic to Cenozoic. However, new multichannel seismic data collected by the Marine Arctic Geological Expedition (MAGE, Murmansk, Russia) and by Regional Geodynamic Laboratory (LARGE, Moscow) reveals that the sedimentary cover on the Laptev shelf (along Oleneksky Bay and Buor-Khaya Bay; Fig. 1 numbers 4 and 6) is continuous with that of the East Siberian shelf described above (Drachev et al., 1998; Vonogradov and Drachev, 2000; Gusev et al., 2002). It is composed of Cretaceous to Cenozoic strata deposited on folded Early Cretaceous and older rocks.

On a seismic profile along the Khatanga Bay (Fig. 1 number 1) there is a sharp change in the character of folding at the Late Mesozoic fold front, which can be traced from the eastern coast of Bolshoi Begichev Island to Tsvetkov Cape (Fig. 1 numbers 2 and 3) on the southern-eastern coast of the Taimyr Peninsula (Vinogradov and Drachev, 2000). Horizontal seismic reflectors, typical for the thick sedimentary cover of the northern flank of the Siberian Platform are sharply terminated at this boundary. On the northeastern end of the seismic profile the reflection character becomes
chaotic and looks similar to the acoustic basement along all the profile. Therefore, it appears that the cover sequences of the Siberian Platform do not continue beyond Khatanga Bay in the Laptev Sea. Rather, the Laptev Shelf is part of the Mesozoides of Northeastern Russia.

There is still a difference between the west and central parts of the Laptev Sea and its eastern part. In the eastern part the main stage of folding is connected with Early Cretaceous time because coal-bearing molasse rocks of Aptian-Albian age on Kotel’ny Island (Fig. 1 number 9) overly folded Paleozoic and Mesozoic formations with a sharp unconformity. On the central and western part of the Laptev Shelf, riftogenous processes were superimposed at the final stage of the Early Cretaceous deformation resulting in

Figure 4. Main structures of the Russian East-Arctic Shelf.

Legend: 1-4 – uplifts: 1 – cover thickness less than 1-3 km and projections of the folded basement; 2 – cover thickness less than 6-7 km; 3 – cover thickness 8-9 km; 4 – cover thickness 16-17 km; 5-6 – monoclines and structural steps: 5 – cover thickness less than 1-3 km, 6 – cover thickness 4-5 km; 7-10 – grabens and depressions: 7 – cover thickness 3-5 km; 8 – cover thickness of 5-7 km; 9 – cover thickness of 9-12 km; 10 – cover thickness of 20 km; 11 – Nansen and Amundsen deep basins; 12 – Gakkel Ridge; 13 – continental structures bordering coastal plains and the shelf; 14-16 – boundaries: 14 – of the major structures, 15 – of the large structures, 16 – of the grabens and horsts on the De-Long rise and the Chuckchi-East Siberian basins.

rather weakly deformed basement rocks being buried under rift formations.

The Late Cretaceous sedimentary and stratigraphic history of the Laptev Shelf is characterized by intense denudation of the area northeast of the Lena River mouth. This is supported by the presence of erosional products in Paleocene strata from granite batholiths emplaced at the end of Early Cretaceous to the beginning of Late Cretaceous in northeast Russia. Further evidence is deep erosional truncation and weathering crust reported in the Tiksi Bay area, where Paleocene sediments, including quartz siltstones, cover the greenish Verkhoyan suite. Considering the area of uplift (Figure 4--I; and number 6), and assuming a depth of emplacement of granitic intrusives of 3 km or less, a volume of eroded rocks over Northeast Russia during Late Cretaceous time, may be up to 6.5 million km$^3$. On the adjacent shelf, continental slope and deep Eurasian Basin, nearly 7.5 million km$^3$ of sediments were deposited.

A quiescent stage set in between the Mesozoic and Cenozoic manifested by Paleocene peneplain facies in the areas of Tiksi Bay, Yano-Indigirskaya lowland, and New Siberian Islands. On seismic records this boundary is expressed by a high contrast reflection representing a regional unconformity between Upper Cretaceous strata. These strata exhibit numerous reflectors that show syndepositional deformation, filling rift grabens, and they are overlain by continuous seismically “transparent” Paleocene strata.

In the North Chukchi Basin, Upper Cretaceous strata with widely spread clinoforms attain the importance an independent sequence (Upper Brookian Sequence), separated from Barremian–Albian rocks by a sharp angular unconformity (Thurston and Theiss, 1987, Plate 7).

**STRUCTURAL PROVINCES**

The structure of the sedimentary cover of the eastern Arctic shelf of Russia is generally an assemblage of large basins and rises that separate zones of persistent depressions such as the Laptev basin (Fig. 4, I), New Siberian System of horsts and grabens (Fig. 4, II), Chukchi-East Siberian basin (Fig. 4 IV), De-Long rise (Fig. 4, III), and a series of perioceanic depressions along the Shelf margin (Fig. 4, V).

The Laptev Basin (Fig. 4, I) occupies the central and western parts of the Laptev Sea. It is about 400 km wide at its northern end near the continental slope, and is less than 100 km wide to the south-southeast in the Buor-Khaya Bay area. On the south and west, the Laptev Basin is bounded by mountainous, folded Mesozoides, and on the east by the Lazarev fault that separates the basin from the New Siberian System of horsts and grabens. The internal structure of the Laptev basin is rather complex due to numerous faults, uplifts and troughs. Sedimentary strata reach thicknesses of 10-12 km in troughs and thin to 5-6 km over horsts.

The New Siberian System of horsts and grabens extends from the continent to the shelf margin between the Laptev and Chukchi-East Siberian basins. It is 600 km wide in the south and about 400 km in the north. Overall, it is an uplifted province with reduced and discontinuous sedimentary cover, except in Novosibirsky and Anisinsky grabens, where sedimentary cover is thickest (Fig. 4, numbers 1 and 2). The structural expression of the western and the eastern slopes of the New Siberian System are dissimilar with the western slope being highly dissected by horsts and grabens and the eastern slope rather gentle with sparse grabens. The axial zone of the New Siberian System, can be regarded as a submeridional horst that is a direct continuation of the Lomonosov Ridge--basement is exposed along it length. The western slope of the Lomonosov Ridge is also dissected by numerous horst and grabens, and the eastern slope is relatively gentle with rare grabens. This resemblance between the New Siberian System and Lomonosov Ridge suggests that they are part of a transregional positive tectonic feature that trends from the continent, through the shelf, to the Arctic Ocean basin. The axial zone of this transregional feature is traced further on land in the Yano-Indigirskaya lowland as a series of basement protrusions, including several granite massifs of Cretaceous age collectively named the Chokhcheuro-Chekurdakhsky row. (Fig. 4, number 25)

The Chukchi-East-Siberian Basin is the largest structural province of the eastern Arctic shelf (Fig. 4, IV). It extends in latitudinal direction over 1300 km, widening from 450 km in the west to 900 km in the east (in the U.S. Chukchi Sea). From the west, the basin is bounded by the New Siberian System of horsts and grabens, from the north by De-Long Rise, from the south by mountainous Mesozoides of North-East Russia. The coastal lowlands are part of the southern flank of the basin. To the east in Alaska, the basin is bounded on the south by the Herald Arch, Lisburne Hills, and Brooks Range and in the north -- by the Barrow Arch.

The basin can be divided into northern and southern parts based on basement age. The northern province is underlain by Caledonian basement, the southern province by late Mesozoic basement. The provinces are separated by large high–angle faults. In the northern province on northeast Russia shelf, two deep depressions--Zhokhovsky (Fig. 4 number 3) and North-Chukchi (Fig. 4 numbers 5)--are separated approximately along 174 degree west longitude by the
Jannetsky transverse uplift (Fig. 4 number 3a).

The Zhokhovsky depression extends for 600 km from the east, where it is 200 km wide, to the west where it gradually narrows and disappears in the boundary zone between the New Siberian System of horsts and grabens and De-Long Rise. Upper Paleozoic–Cenozoic strata in the axial zone of the depression reach 10-12 km in thickness.

The North-Chukchi Basin within the Russian shelf is also traced for 600 km and its width varies from 250 km on the east to 160 km on the northwest (Fig. 4 number 5). This basin is notable for its great sedimentary thickness. On seismic records, acoustic basement is reliably recognized at the depth of 18 km (Fig. 3). It has an asymmetrical structure with its southern flank dipping steeper than its northern flank. The northern flank axis of the basin is crossed by the transverse Andrianovsky uplift along the 170° meridian (Fig. 4 number 5a). The uplift has up to 3000 m of relief at the level of the Barremian-Albian strata (LCU), but has practically no expression on the top of the Upper Cretaceous strata (mBU). Twenty to twenty five kilometers north, the axial zone of the North Chukchi basin shifts to younger strata.

The North-Chukchi basin is dissected by sublatitudinal and younger submeridional faults. Sublatitudinal faults often bound half grabens on the southern flank of the basin. Submeridional faults offset Upper Cretaceous rocks and bound grabens and horsts, which sea floor expression. In the north, the basin is bounded by the arch-like North Chukchi rise (Fig.4 number 4), which extends west-northwest for 500 km and varies from 50 to 75 km in width. To the northwest, the North Chukchi rise conjugates with the southeastern flank of De-Long rise (Fig. 4, III), and on east-southeast probably with the Barrow Arch. Stratigraphic thickness over the North Chukchi rise is estimated at 6-7 km.

The southern part of the Chukchi-East-Siberian basin (Fig. 4, IV) differs from its northern part by the presence of predominantly submeridional structural trends inherited from the late Mesozoic basement. Features with sublatitudinal strike, typical of the northern part, are partly preserved only on the Chukchi Shelf, such as the Herald Arch (Fig. 4 number 14) with thin sedimentary cover and basement projections (Wrangel Island; Fig.1 number 20), and the South Chukchi depression (Fig. 4 number 15) with up to 4-6 km of overlying Cretaceous-Cenozoic strata.

Over most of the East-Siberian shelf the structural trend is predominantly submeridional with symmetrical features. The axial zone of the structural assemblage is characterized by the Melvillian graben (Fig. 4 number 8), where Aptian-Cenozoic strata reach 10 km in thickness. This graben structure is actually a zone of extension, flanked by uplifts of the Chukchi and East-Chersky (Fig. 4 numbers 7 and 9), and then flanked again by subsidence features of the South-Denbarsky and Ambarchiksky grabens (Fig. 4 numbers 6 and 10). The basement fault zones, responsible for origin of these structures extend to the north, dissecting Zhokhov depression and De-Long rise. It is very likely that submerged parts of the East-Siberian shelf have riftogenous nature, similar to the Laptev basin.

The De-Long rise has a block-like round-to-triangular form (Fig. 4, III), elongated in a west-northwest direction for 800 km. It is 400 km wide in the west and narrows to the east to 150 km. For the most part it is covered by thin (less than 1 km) Cretaceous-Cenozoic mantle type deposits with several projections of Caledonian and probably older basement. On the slopes of De-Long rise the cover is 3-4 km thick, underlain by Mesozoic and Paleozoic strata. The rise is dissected by faults, and bounded grabens and horsts.

CONCLUSIONS

We draw four main conclusions on the age and structure of the sedimentary cover of the eastern Arctic Shelf of Russia.

1. The structure of the sedimentary cover was formed during two stages that resulted in two structural trends and related tectonic features. The time boundary between these stages lies in Barremian-Aptian. In the older stage sublatitudinal trends predominated, and in younger stage--submeridional.

2. Older sublatitudinal zonation was caused by structural features of the basement, whereas younger submeridional zonation appeared as a result of a unique ocean and shelf riftogenous processes. Structural elements of the Arctic Ocean, such as the Eurasian Basin, Lomonosov Ridge, Mendeleev Rise, and Chukchi Basin generally express continuations of the shelf structure in their sedimentary cover.

3. Another important relationship between ocean and shelf is the position of a thick lens of the Lower Brookian Sequence in the North-Chukchi basin just opposite the Chukchi basin in the Arctic Ocean, between Mendeleev Rise and Chukchi Plateau.

4. Active riftogenous processes started simultaneously over all of the eastern Arctic Shelf in the Early Cretaceous and finished by the beginning of the Cenozoic and active tectogenesis migrated in time toward the Arctic Ocean. In the Cenozoic, on the shelf-ocean boundary, perioceanic depressions were formed, whereas on the shelf in a quiescent environment, a thin and continuous sedimentary mantle was deposited.
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THREE-DIMENSIONAL STRUCTURAL MODEL OF THE SOUTHEAST COASTAL SHELF OF THE EAST SIBERIAN SEA AND THE ROLE OF PROCESSES OF BASIFICATION IN THEIR FORMATION (A NEW GRAVITY INTERPRETATION)

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ABSTRACT
A study of the structure and rock type on the coastal shelf of the East Siberian sea, bounded by the coordinates 70° and 70° 40’ N Lat., and 165° 15’ and 174° E. Long., was made based on the interpretation of a gravity survey on the ice using multiple observations on a 2x2 km grid. Quantitative interpretation of these results was done on the basis of gravity modeling techniques developed at the Laboratory of Regional Geophysics of the North East Interdisciplinary Scientific Research Institute of the Far East Branch of the Russian Academy of Sciences, Magadan (Vashchillov, 1984; Vashchillov, 1994; etc). These methods were used to develop a 3-dimensional block model of anomaly sources.

INTRODUCTION
The area of investigation is located offshore of the Chukotka Fold Belt (Fig. 1). The largest tectonic structures in the region are the Anyuy and Chaun folded zones, which are separated by the Late Mesozoic Rauchansky depression. For density and geological-petrological interpretation of the gravity data, we used the results of geological and petrology-density studies of the rock types that are exposed along the Arctic coast of the studied offshore region.

GELOGIC SETTING, METHODS, AND TECHNIQUES
The rocks of Yanranajsky, Kujviyemsky, Severny, Pyrkanayansky and other granitic massifs of the Chaun folded zone have an average density of about 2620 kg/m³ and 2690 kg/m³ at depths of 1 and 6 km respectively. These granitic massifs were used like natural density “bench-marks.”

Three-dimensional density and geological models of the offshore area were determined for horizontal datum levels at depths of 1, 6, 10, 20, 30, 40 and 60 km (gravitational tomography) and vertical sections – one E-W and two N-S profiles (Fig. 2). The E-W section extends along the 70° 20’ N Latitude line. Vertical density distributions in the lithosphere of this offshore sector are composed on base of 14 interrelating regional vertical density sections of lithosphere on the mainland and horizontal density sections at depth of 10, 20, 30, 40 and 60 km in Northeast Russia (Vashchillov, 1994). These sections were used to construct horizontal sections of the lithosphere of the shelf at the same depths. Interpretation of contrasting lateral changes in density on the vertical block’s limitation, allowed the absolute values of density to be carried from one density block to another. There are many comparisons which show the advantage of the new gravity interpretation technique, even before deep seismic
Figure 2. The vertical section XV of the lithosphere on the southeastern East-Siberian sea (setting of the profile is shown on Fig. 3): 1- granite; 2- granodiorite, adamellite; 3-diorite; 4-mafic; 5-mafic-ultramafic; 6-ultramafic; 7- eclogitic subcrust substance and ultramafic; 8- eclogitic subcrust substance; 9-sedimentary-terrigenous, partly metamorphosed rocks; 10- metamorphic rocks; 11- vertical limits of density heterogeneities (blocks), arrows show the direction of density decreasing, the figure is a lateral contrast of density (kg/m$^3$·10$^{-3}$); 12-rock density of the given depth (kg/m$^3$·10$^{-3}$); 13- indexes of lamination surface of lithosphere (see the text); 14- upper and lower limits of the density heterogeneities-blocks; 15- quasi-horizontal surfaces of division; 16 - the curve of the gravity $\Delta g$ (mGal).
surveys are conducted (Vashchillov, 1984; 1995). Deep seismic research (DSR) has complexities in determining velocity and it has a two-dimensional character. A geological and petrological interpretation of deep seismology is more difficult than the same interpretation of gravity. The new methodology of gravity interpretation is thousands of times cheaper than DSR and yet solves all the same tasks and even much more. The interpretative accuracy of DSR isn’t better (and is frequently worse) than of this gravity interpretation.

RESULTS
Geologic and petrologic interpretations were made for the horizontal and vertical sections. Geological interpretation and the assignment of rock age were carried out for the upper parts of the Earth’s crust (Fig.3). Horizontal density variations at a depth of 1 km were transformed into a geological map showing lithology and rock ages. At a depth of 1 km (Fig. 3), rocks to the west of the North-Ayonsky fault have a density of 2540-2570 kg/m³. This density is characteristic of the Lower Cretaceous rocks of Neocomian age. Neocomian sediments at a depth of 1 km can be traced to the east from the North-Ayonsky fault. They occur among the sedimentary rocks of Triassic, Jurassic, Paleozoic and Lower Cretaceous intrusives. These last rocks are crossed by the horizontal section at depth of 1 km. The maximum thickness of the Neocomian sedimentary rocks is about 2.5 km according both to the geological data and the results of gravity anomaly interpretation. As a whole, rock densities range from 2510 to 2720 kg/m³ at a depth of 1 km. The intrusive rocks (granite, granodiorite and diorite of Early Cretaceous age) are shown on the horizontal section X-V.

The distribution of faults and blocks of different density was compiled. The numbers give minimum possible depths of the location, calculated on the basis of quantitative interpretation of gravitation data. All these were combined with the map of gravity anomalies in the Bouguer reduction (Fig. 4).

The N-S fault, approximately parallel to 168°35’ E Long. (Fig. 4), which we have named the North-Ayonsky fault, divides all the offshore area into two parts. The region of relatively smooth gravity field, not complicated by local gravity anomalies and with features suggesting a stable massif or platform, is located to the west from the fault. The eastern, larger part is characterized by disturbance of the gravity field and more anomalies, faults, and blocks. We distinguish the North-Ayonsky fault as the western limit of a N-S superlineament, that goes through all of Chukotka, Koryakiya, and along the underwater Shirshov Ridge of the Bering sea to the Emperor Seamount Range in the NW Pacific Ocean (Fig. 5). The depth of penetration of North-Ayonsky fault is

Figure 3. Density horizontal section and the geological scheme of the southeastern marine area of the East-Siberian sea at the depth of 1 km. 1--granite; 2--granodiorite; 3--diorite; 4--mafic; 5--mafic-ultramafic; 6--ultramafic; 7--eclogitic subcrust substance and ultramafic; 8--eclogitic subcrust substance; 9--sedimentary terrigenous, partly metamorphosed rocks; 10--metamorphic rocks; 11--vertical limits of the density heterogeneities, berg-striation show the direction of density decrease; 12--initial values of density (kg/m³·10⁻³) on the geological bench-mark—granitic massifs; 13--calculated density of rocks at the given depth (kg/m³·10⁻³).
not less than 65 km.

The depth to the base location of most faults and blocks is from 65-70 km to 8-10 km. Besides the North-Ayonsky fault, there are other large faults in the offshore area. One is the coastal Shelagsky fault, delineating the coast here and then extending onshore near Kiber Cape. This fault, branching off in the East-Siberian Sea to the north-east, like the other ones is located by the zones of increased horizontal gravity gradients and have penetration to depths of 40-38 km—they pierce through the Earth’s crust. The system of N-S faults between 171° E and 171°30’ E long., extend from the coast and are bounded by the above mentioned north-eastern striking faults to the north. The N-S fault system extends to the north from the Aachim Cape and has a depth of penetration of at least 65-80 km.

The great thickness of the Earth’s crust seen here (38-40 km) is characteristic for crustal platform and continental sections, but not for oceanic crust.

The E-W cross-section X-V across the region along 70°20’ N Lat., (Fig.2) shows that the Mohorovičić boundary (M), located at the depth of 35-42 km, is distinctly seen everywhere. The changes of rock density near the M boundary take place within the interval of 3160-3220 kg/m³. Near the top

Figure 4. Block tectonics of the southeastern water area of the East Siberian Sea after gravity interpretation: 1- conventional isoanomalies of the gravity field (mGal); 2- faults, distinguished according to the zones of increased gradients ∆g; berg-striation show the direction of the density decreasing; 3- line and number of the lithosphere vertical section; 4- minimum possible depth of the fault location; 5- minimum possible depth of location of the density heterogeneities (blocks). SH – Shelagsky fault; NA – North-Ayonsky fault. 

Figure 5. Asia – Pacific NS lineament (dotted line)
of mafic layer B, the density range changes to 2850-2900 kg/m³; density interfaces within mafic layer B₁ are 2880-2950 kg/m³; and at the top of the mafic-ultramafic layer M₁ they are 2960-3050 kg/m³. Surfaces inside the mafic-ultramafic layer M₂ have the densities of 3010-3130 kg/m³; surface P₁ in the upper mantle has a range of density change of 3210-3270; and surface P₂ has a range of 3280-3340 kg/m³.

Boundary P₂ is shown on the vertical section X-V (Fig. 2). It is traced distinctly in the western half of the profile. The top of the mafic layer B, appears to change or “step” up to the east. In the region east of the North-Ayonsky fault, at some places, it comes near the surface to depth as shallow as 5-4.5 km. In other places it is deeper, down to depths of 10 km and more, but not deeper than 15-16 km. The boundary of the low crust (the top of mafic-ultramafic layer) M₁ is distinguished in the central parts of the section but only sporadically identified elsewhere.

Figure 6 shows good coincidence between the theoretical gravity field Δg, that is calculated for the model of lithosphere, and observed fields Δg. The accuracy of the calculation is ±1 mGal. Other such comparisons of theoretical and observed gravity maps are nonexistent. Theoretical gravity maps, having such accuracy, are unique.

CONCLUSIONS

There are two distinct regions, located to the east and the west of the North-Ayonsky fault. The region to the east of the fault is an area with reduced or completely reworked "granitic" layer with thickness sometimes less than 5-6 km. The crust thickness is stable - 36-40 km. To the west of the North-Ayonsky fault, the Earth’s crust is of the normal kind with a developed "granitic" layer 20 km thick or more. The crustal thickness is the same as in the eastern part of the section, where the thickness of the "granitic" layer is reduced by processes of basification. It is possible that we have identified a fragment of the platform that is on the west from the North-Ayonsky fault.

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GLOBAL TECTONIC ACTIONS EMANATING FROM ARCTIC OPENING IN THE
CIRCUMSTANCES OF A TWO-LAYER MANTLE AND A THICK-PLATE PARADIGM INVOLVING
DEEP CRATONIC TECTOSPHERES: THE EUREKAN (EOCENE) COMPRESSIVE MOTION OF
GREENLAND AND OTHER EXAMPLES

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ABSTRACT

The intra-Eocene Eurekan (s.l.) compression, extending from Ellesmere Island to western Svalbard, has vergence that varies along strike and appears to be the product of a broadly northward force acquired by Greenland. But how? To approach this puzzle we re-examine the pattern of mantle motion implied by the movement of plates.

Three kinds of evidence indicate that subducting oceanic plates actually incorporate hot mantle at depth, one of these being the tomographic reappearance of this heat as slab reheating during active subduction. The latter apparently results in the partial melting of the subducted oceanic crust, producing lumps of stishovitic residue that fall into the lower mantle. Tomographic patterns of apparent slab penetration into the lower mantle accord with this reinterpretation, implying that we have two-layer mantle convection but with very slow upward seepage/diffusion of lower mantle material though the 660 km boundary to offset the volume input as lumps.

To explain the presence of this heat in oceanic plates, a deep-axial-mantle-'crack' model of their genesis at MORs is outlined, whereby the LVZ is formed as a mechanically integral part of the much thicker (80 km? initially) plate. Solid-state mineral phase-change transformation causes inward bulging and contact of the crack walls, generating powerful push-apart force and this 'sucks' mantle into the crack along-strike and the process repeats. An ultraslow-spreading variant of this model accords well with Gakkel Ridge features.

Remarkably, thicker plates even confer advantages in respect of the subduction downbend process. Rather than elastic flexure an escalator-like through-plate step-faulting mechanism is outlined that also supports outer rises. This provides a basal subduction tectonic erosion (STE) mechanism that offers progress in understanding how flat subduction, now widely recognized, arises and may be the precursor to the imbricate thrust systems in collision and UHP terranes.

A exciting further benefit of thicker plates is that it offers to validate the anticlastic curvature mechanism at plate downbend as the cause of back-arc opening in the presence of ridge push forces, thus resolving a paradox of very long standing.

For cratons, seismological, petrological and geochemical evidence is assembled which support that they have tectospheric keels that reach to well beyond 500 km. This, in the context of a two-layer mantle, raises the problem that when a craton is split, where can the new MOR 'suck' the mantle from to put beneath the growing ocean plate? For the opening Atlantic the eastward motions of the Caribbean and Scotia plates offer an answer.

For the Arctic Ocean, a northward drag upon Greenland's keel, causing the Eurekan compression, could be one of the results. The northward motion of India, due to deep mantle flow drawn through the West Siberian gap between cratonic keels, could well be another. Thirdly, flow through the Bering gap may have initiated the Aleutian arc Thus the tectonics of the Arctic, both its MOR and the Eurekan compression, offer a new dimension to the debate on the form of mantle convection, previously limited to seismology and geochemistry.

INTRODUCTION

Lacking information about the ocean floors, Wegener saw the moving of a continent as involving only the continental crust. With the advent of sea-floor spreading and plate tectonics in the 1960s it became clear that plates must include a limited thickness of mantle, but with functions limited to cooling shrinkage on the flanks of MORs, all the mantle below that being assigned, at least initially, the jobs of moving the plates about by convection and the (ad hoc) provision of sources of intraplate heating and magmas.

Since that time, development of the plate tectonics paradigm has continued to pursue this thin-plate mode in which even the somewhat thicker lithosphere recognized beneath continents was not thick enough to provide significant constraints upon mantle flow patterns. The theme to be outlined here is that a change to a very much thicker-plate form of the paradigm is highly remunerative observationally on a wide range of scales.

One such benefit, to which attention was originally drawn by Osmaston (1977), but which will be dealt with more fully elsewhere, is that if plates are thick...
enough splitting them in a particular way provides a mechanism for intraplate magmagenesis, thus offering an escape from a dependence upon independent mantle flow patterns below.

In this contribution I outline a series of arguments (A) that, contrary to widespread perception, the present mantle convective pattern is actually a two-layer one, having changed from whole-mantle overturn during the very early Proterozoic, and (B) that the mechanically integral thicknesses of tectonic plates in general are very much greater than currently is recognized, extending beneath cratons, in particular, to near the base of the upper mantle at 660 km depth. It will be reasoned that this combination of circumstances has had, and/or is having, a major effect upon upper mantle flow patterns, resulting in a number of very major and obvious plate tectonic effects globally, of which the essentially Eocene compression that affected a crustal band from SE Ellesmere Island to west Svalbard has been especially puzzling. Although originally named for the event in the Canadian Arctic islands (Thorsteinsson and Tozer, 1970) the name Eurekan is here, for brevity, widened to include the Svalbard compression.

The Eurekan compression/thrusting was mainly SE-vergent in SE Ellesmere Island (Piepjohn et al., 2000), north-vergent in north Greenland (Lyberis and Manby, 2001) and strongly ENE-vergent in western Svalbard (Harland and Horsfield, 1974). Due to the convergence of geographic north directions at these high latitudes these directions are near-consistent with a single compressive motion (Fig. 1). In reviewing the event, both Tessensohn and Piepjohn (2000) and Lepvrier (2000) comment that although plate separation in the northern N Atlantic and Eurasian Basin began (Gulf of Labrador, via the Wegener Fault in Nares Strait) at about chron 33 time (83 Ma) (Srivastava and Roest, 1995) the compression did not begin until after separation had also got started on the other side of Greenland, at Chron 24 time (early Eocene), thus seeming to have required that the Greenland plate was independently 'adrift'.

Figure 1. Distribution of the Eurekan compression. The big arrow indicates the effective independent motion of Greenland inferable from that compression. Its amount is currently undetermined. The interpretation of its cause, developed in the text, is that the Arctic-directed flow of upper mantle material in response to the progressing separation (slow though it is), by the Arctic (Gakkel) MOR, of the deep-keeled cratons around the Arctic, pushed northward upon Greenland’s keel. (Modified from Tessensohn and Piepjohn, 2000.)

The problem lies in that any crustal shortening inevitably requires the downward ‘disposal’ of a corresponding amount of its lithospheric mantle and a considerable force to do that, or at least to initiate it. The geology along the northern side of the belt, partly continental, partly (at that time) very young ocean plate, would have hindered rather than promoted any attempt at subductive disposal, and the conflicting vergences hardly support ‘normal’ subduction either. Our job here is see how an appropriate force might have been generated.

To argue that plates are actually very thick might seem at first sight be a step in the wrong direction for this problem. The essence of the argument to be developed in this paper, however, is that the demand, generated by the craton-separating activity of the Arctic (Gakkel) MOR, for mantle to put beneath the Arctic floor (and all the way down to the 660 km discontinuity), ‘sucked’ Greenland’s deep tectospheric keel northward. Onset of the motion, and of the Eurekan compression, at the beginning of the Eocene

Osmaston
marked the completion of Greenland’s detachment from the plates on each side of it. The cessation of the Eurekan compression at the end of the Eocene then records that the north-east Atlantic ‘gap’ at depth was now wide enough for the mantle flow to the Arctic to by-pass Greenland’s keel.

The paper proceeds in three main stages:- (1) a fresh assessment of the subduction process indicating the presence of two-layer mantle convection; (2) characterization of the genesis and constitution of the proposed thicker plates at both ends of the age-scale - MORs and cratons, noting their intermediate behaviour at subduction downbend as discussed in (1); (3) brief discussion of the chemical and global tectonic consequences of this combination, of which the Eurekan compression is one.

The paper introduces new models or perspectives for a number of Earth processes, stressing their action in qualitative terms and judging their value as links in the conceptual framework. To attempt their quantification here would far overrun the limitations upon the paper’s length and, moreover, would in many cases be insecure because the new models lay stress upon some physical parameters to which, not surprisingly, little attention has been needed in the context of existing models.

Overall the paper seeks to reinforce and refine, in an Arctic-based context, but with wider applicability, our perceptions of the linkage envisaged by Hager and McConnell (1979) when they wrote (p. 1031) ‘It is likely that the motions of the lithospheric plates strongly influence the accompanying large-scale flow in the earth’s mantle’.

A FRESH LOOK AT SUBDUCTION

Many-faceted study (>25 yrs) of the subduction process has shown me that, contrary to the standard thin-plate perception, the oceanic plate arriving for subduction must actually have thermal buoyancy, especially when it is <70 Ma old. There are three lines of evidence for this:-

(a) the mechanical development of extensive 'flat-slab' interface profiles by progressive detachment of upper-plate material (STE), sometimes including that from the middle and lower crust;
(b) occurrences of wide belts of silicic/granitoid post-subduction magmatism (PSM) when young-plate subduction ceased;
(c) reappearance of that heat in the form of slab reheating during active subduction, seen in many tomographic transects of subduction zones, but often dismissed as slab drop-off.

(a) Basal subduction tectonic erosion (STE).

The basis for this process is the recognition (Fig. 2), based on Melosh (1978) and Melosh and Raefsky (1980), that the support of outer rises does not require that subducting plate downbend is by elastic flexure but can be achieved, consistently with seismological evidence, by an escalator-like through-plate step-faulting process (Osmaston, 1992a; 1994) (Fig. 3). Since the effective angle attained by outer trench slopes is only a few degrees, most of the downbend slope develops by increments of step-fault throw beneath the forearc. Each such increment offsets the interface locally and locks it, the sliver of upper-plate material trapped in the step-angle being sheared off when subduction resumes. A feature of this process is that the positive locking of the interface provides the means, much better than simple stick-slip on an already slip-prone interface, for the accumulation of the strain energy released in earthquakes of magnitude 9.0 or more (Osmaston, 2005a). Repetition of the process results in the downbend position, carved into the upper plate (Fig. 3), advancing at shallow dip beneath the margin, an overall sequence that has been called basal subduction tectonic erosion (STE) (Osmaston, 1994). This alternating process is recognizable as seismic coupling, well known to seismologists (Christensen and Ruff, 1983; Lay et al., 1989; Osmaston, 1999). Furthermore, the forces thus coupled intermittently to the upper plate are widely seen as foreland-directed thrusting that has progressed across the belt (Jordan et al., 1983 and many later papers by others), as the downbend did, enabling it to be documented that in both the Andean examples the advance has been about 300 km in the past 10 Ma (Osmaston, 1995a). Flat subduction is actually a more widespread phenomenon than generally realized and is accompanied by thrust faulting far into the interior (Gutscher et al., 2000), raising a question about the widely presumed mobile state of mantle wedges. Although STE advances the downbend at a shallow angle and starts at a shallow depth, the margin thus undercut may imbricate and foreshorten the system so that a further cycle of STE then ensues, at a

Figure 3. Sketches of the proposed mechanism of subducting plate downbend and of downbend advance by basal subduction tectonic erosion (STE) of the upper plate. At any given moment, where beneath the upper plate, all the step-angles will be full of material acquired from the hanging wall as the result of successive increments in step-fault throw, each succeeded by shearing off the hanging wall material that has entered the step-angle (A). The cylindrical faults are of the same radius, each having been initiated in the same way; this provides compression at depth and an outer rise. Details of the action are in (B). The line of shear-off is fixed by all the undisturbed part of the interface, not by the relative ‘toughness’ of the rocks involved. The process generates two different kinds of earthquake - step-fault and interface slip - a recognized feature of seismic coupling. In the step-faulting case, it appears that the seismic energy comes mostly from the upper, normally-inclined portion of the fault, masking that from the rheologically weaker reverse portion. If the subducting plate has insufficient buoyancy for good mechanical contact with the upper plate, this STE action will be weak or absent.
deeper level (Osmaston, 1992a); this appears to be why, in both the Andean 'flat-slab' segments, the interface descends quite quickly to 50-80 km depth before resuming a low angle (Isacks and Barazangi, 1977; Cahill and Isacks, 1992).

This shallower stage of the undercutting of margins by STE offers, upon imbrication of the margin, to provide slices of continental basement, much thinner than the whole crust, that have lost their lower part and are widely seen in collision belts (e.g. the Alps) and UHP terranes (Ernst and Liou, 1999; Osmaston, 1997, 2004). The STE-specific type of mélange that is generated (Fig. 3B) is then widely exposed globally (author’s unpublished observations).

STE clearly demands the presence of buoyancy in the subducting plate; slab pull is not an appropriate agent for achieving this huge 'bite' out of the mantle wedge, thus lengthening the interface profile, and buoyancy is also essential to maintain the necessary close mechanical engagement for STE to work. Studied cases of seismic coupling (Lay et al., 1989) and of extensive STE undercutting exhibit a degree of correspondence and appear to be, or to have been, where the subducting plate was/is <70 Ma old, strongly suggesting that the buoyancy is thermal. This means, in effect, not that the 'slab' is somehow not cold at all, but that the plate must actually be thicker and incorporate hot material below the 'slab'. This is accomplished by the new MOR model discussed later. The Andean segmentation suggests, however, that plate-age-related thermal buoyancy is not the only controlling factor on flat-interface development and that the buoyancy of subducted aseismic ridges may be the critical additional trigger, a long-discussed correlation recently stressed by Gutscher et al. (2000).

(b) PSM (post-subduction magmatism).

Post-subduction silicic/granitoid magmatism has been widely recognized, but under the limiting name of post-collision magmatism. The essence of PSM (Osmaston, 1992b) is that the cessation of subduction, whether as the result of continental collision or of some other change in plate dispositions, provides time for this within-plate heat, if present, to make itself evident, by conduction, as progressive partial melting of the subducted oceanic crust along the interface. A diagnostic expectation of this process is that, since the melting is reheat-time dependent, its products will appear first above the deep end of the interface (limited by the depth beyond which the melts go down instead of up, because of their higher compressibility) and work their way trenchwards.

A clear and well-documented case of this migration is the Siluro-Devonian granitoids and arc-type volcanics of Scotland, marking the completion of Iapetus closure (Watson, 1984). The dates of onset migrate southeastward from 428 Ma close to the Great Glen to 397 Ma in southern Scotland. The moment of complete closure is anchored at about 424 Ma by the arrival into the Lake District (northern England) of a sudden influx of NW-derived proximal turbidites from the Southern Uplands (SU) in the late Wenlock (~424 Ma on the Gradstein et al, 2004 timescale) (McKerrow and van Staal, 2000). Widespread sedimentary changes south of the suture, beginning at the basal turriculatus biozone (436 Ma) suggest that the SU accretionary complex encountered the southeastern shelf edge of the ocean at this moment, and underwent progressive compression before riding up onto it somewhat later (427 Ma?). Slowing of plate closure during this interval seems likely to have been what triggered the 428 Ma onset of PSM.1

Attention to a similar migration of silicic-rhyolitic magmatism was originally drawn by Coney and Reynolds (1977; Coney, 1978) in regard to the post-Laramide westward 'sweep-back' sequence in the southwestern USA. In this case the relevant cessation of subduction appears to have been, not collision, but that the very large amount of inferable downbend advance by STE, to extend tectonism all the way to Colorado, had made following the interface profile mechanically unsustainable by the subducting plate. This resulted in the establishment of a new, 'short-cut', steep interface beneath Sierra Nevada, leaving the former extremely young subducted plate in place beneath much of the belt, with its interface oceanic crust to source the PSM.

Recognizing that both STE and PSM are likely wherever the rapid subduction of young and hot plates was involved has led to progress in understanding the nature of Archaean greenstone belts and their 'continentalization' by the widespread intrusion of near-coeval TTG granitoids (Osmaston, 2001a). It seems that greenstone belts are former oceanic-crusted forearcs, of which all the lower part was removed by rapid STE (making them unrecognizable as ocean crust), until subduction was halted by the arrival of a microcraton, of which many are otherwise missing. The young hot plate this left in place below, together with

1 An interesting plate tectonic aside, in relation to this inferred SU-to-shelf encounter at 436 Ma (turriculatus biozone) is that it coincides remarkably precisely with the similar encounter in North Greenland with the then Baltica plate margin, marked by the abrupt onset of easterly- and SE-derived turbidite deposition (Hurst et al., 1983; F.Surlyk, pers. comm. 1995).
its interface oceanic crust possibly >25 km thick, thus provides both the source material for the TTG and the heat for the partial melting, each of which has been a problem hitherto.

(c) Slab reheating during active subduction.

For the purpose of inferring a two-layer mantle the final and most important step in our argument that subducting oceanic plates incorporate deeper material that is a latent source of heat is to see whether that heat appears even during active subduction. The now very numerous tomographic transect studies available (e.g. van der Hilst et al., 1998; Fukao et al., 2001, together with >150 kindly provided by E.R. Engdahl, pers. comm. 2001, 2005) contain many examples in which the high-velocity signature of the slab fades out at between 200 km and 350 km, only to be replaced by a second high-velocity trace that begins near the top of the Transition Zone (TZ) and continues on through the 660 km level into the lower mantle. It is the latter that has been getting all the attention from seismologists and others but it cannot be due to temperature; having reheated the slab it cannot become cold again. The deeper signature must be mineralogical. On observational grounds (T-waves from S-waves travelling up the slab) in a number of cases, the descending plate is continuous across the seismicity and tomographic gap, so cannot represent slab drop-off (Okal and Talandier, 1997; Okal, 2001). This is matter to which we return later.

Recognized in more detail, the tomographic transects of subduction zones show the following points. Where the plate is young (e.g. <25 Ma) the true slab signature fades by about 220 km depth and there is no seismicity beyond that, consistent with the heating being entirely accomplished by conduction. Where the subducting plate is much older there is a fading of the slab signature at 350 km if subduction is slow enough (e.g Izu-Bonin - plate age 130 Ma) but not where subduction is faster (Tonga, NE Japan). But in all these older-plate cases there is much deep seismicity (and especially in the ultrafast Tonga subduction zone), the energy dissipation from which must supplement the conductive heating enough to promote the crustal melting from which the lumps derive. Seismological evidence has been adduced (Wiens, 2001; Tibi et al., 2003) for an association of melting with deep seismicity. Some of the 'showers' descend almost vertically through the 660 km level (western Pacific) but others, notably under the Americas, have an eastward-dipping trace; this seems to record westward movement of the American source-points, relative to the lower mantle, as the Atlantic opened. Study of these traces/showers offers a new source of evidence regarding relative motions of the plate system relative to the lower mantle; a field in which the hot-spot enthusiasts have hitherto held sway.

TIME OF CONVECTIVE CHANGE-OVER TO A TWO-LAYER MANTLE

The high heating rate in the early Earth would surely have resulted in whole-mantle convection. There is much evidence (Osmaston, 2000b; 2002) that the early Earth was constructed with a wet mantle which, in the water-weakening mode, would have enabled convective disposal of heat easily not only to keep up with heat generation but handsomely to have overrun it by the end of the Archaean (McKenzie and Richter, 1981). So the existence, now of 20 years' standing, of a 2.45 Ga - 2.22 Ga gap in high-quality zircon ages for greenstones and for orogenic granitoids (Condie, 1997; Windley, 1995; B.F. Windley, pers. comm. 2003) (not to be confused with mafic-ultramafic complexes and A-type granitoids) is serious evidence that mantle...
CONSTRUCTION AND CONSTITUTION OF THICK TECTONIC PLATES

A new model for MORs

We established in preceding sections that the behaviour of subducting plates requires that much hot material be incorporated integrally below the lithospheric slab, presumably right from the start, at MORs themselves. A further requirement is that the extensive foreland-directed thrusting associated with STE action seems to require a substantially greater ridge push than is provided by the existing model. In addition, it has been clear that slab pull is not an appropriate force for maintaining the subduction of young, buoyant plates, performing extensive STE.

The key to such a model is that the oceanic low velocity zone (LVZ) does not possess the asthenospheric high-creep properties that geophysicists have customarily assigned to it on account of its interstitial melt content. In fact the affinity, for H₂O, of the <3% melt (this being wholly insufficient for mechanical disaggregation) siphons off the water-weakening of the mineral structure, potentially making it more creep-resistant than the lithosphere above (Karato, 1986; Hirth and Kohlstedt, 1996), though this may not actually be the case if the lower temperature of the latter is allowed for.

The principles of the new model (Fig. 4) were outlined by Osmaston (1995b; 2000b) but subsequent study has shown the wide adaptability of its behaviour at different spreading rates - a matter of great importance in the present context in view of the ultraslow spreading rates encountered along the Gakkel Ridge (Cochran et al., 2003; Dick et al., 2003a,b). Its central features, at medium and fast spreading rates, are as follows. It has a narrow, deep, sub-axis mantle crack into which mantle is drawn by plate separation and undergoes increasing partial melting as it rises. This melt fractionates by crystallizing mainly olivine onto the crack walls and this accretion, together with restite, offsets the separation rate, maintaining its width at a nominal figure of 20 cm, though this will obviously vary with depth. For this wall accretion to occur the temperature gradient up the walls must be superadiabatic and this is a natural result of the LVZ not being mobile. Due to the highly anisotropic thermal conductivity of olivine (Chai et al., 1996) and the contrastingly low thermal conductivity of silicate melt (Snyder et al., 1994) those crystals oriented with their a-axis (the seismically fast one) perpendicular to the walls will grow the fastest by latent heat extraction, thus building-in by crystallization the seismic anisotropy of oceanic plates, especially well-marked in the Pacific. The accretion onto the two walls will differ according to the local ability to lose the heat laterally, a process that gives the crack a self-straightening property (Osmaston, 1995b), thus explaining the straightness of MOR segments and their orthogonal segmentation.

For our present purpose the most important feature of the model is the way in which it operates and generates ridge push. This depends upon the large (some tens of times greater than thermal expansivity) volume increase per joule input produced by the high temperature solid-state phase-change transformations of mantle material from garnet- to spinel-peridotite and from spinel- to plagioclase-peridotite (Osmaston, 1973; Wood and Yuen, 1983). For the latter phase change the factor appears to be as high as 130-fold (B.J. Wood, pers. comm. 2002). Depending on the spreading rate, one or both of these will be present at an appropriate level in the crack walls. An eruption up the crack will heat the walls, but at the phase-change level the walls will bulge inward additionally, closing the crack (if it is narrow enough) and pushing the plates apart with the available recrystallization force. Certainly high, the precise evaluation of this force is currently problematic. At most spreading rates the applicable phase change is the garnet-to-spinel peridotite, so the push-apart level (at 50 km²/s?) results in a substantial rift valley. But at the fast spreading rates seen on the EPR the ridge can be expected to have a relatively restricted zone of plagioclase-peridotite near the crest, so the push-apart level is very shallow and the rift valley minimal or absent. In operation, it is envisaged that this push-apart of the plates, following an eruption at one place, will rise open the crack along strike, causing the process to repeat there, and so on.

At the other end of the scale, very slow spreading means that the mantle induced into the crack undergoes insufficient melting to achieve melt segregation for formation of typical oceanic crust, so the expected result is a peridotitic intrusion laced with melt veins and probably much wider than the crack width.
Figure 4. Principles of the deep and narrow mantle-crack model for mid-ocean ridges (MORs) (Osmaston, 1995b; 2000b). Crustal construction details omitted. Seismic anisotropy is built-in by columnar olivine crystallization on the crack walls. For spreading action see text. The incidence, or not, of rift faulting depends on the depth of phase-change push-apart. At slow spreading rate, illustrated here, the relevant phase change is the garnet-to-spinel peridotite but at fast ones the much shallower spinel-to-plagioclase peridotite change comes into play and downfaulting is suppressed. At intermediate rate, these forms may alternate. The important variant associated with ultraslow spreading is discussed in the text. The form of the lithosphere-LVZ boundary is wholly pictorial; it needs recomputing in light of the LVZ properties discussed in the text. Its interstitial melt lowers the thermal conductivity and traps heat until subduction raises the pressure and refreezes it. The depth of the effective base of the plate will depend on the applicable displacement stress levels rather than upon clear seismological criteria. For ophiolites, the counterpart model, exploiting the thick-plate paradigm in the case of those emplaced hot, was outlined by Osmaston (2001c).

operative under high-melt conditions. By the same token the wall-accretion crack-straightening mechanism will not work either, nor will orthogonal orientation of the axis operate. These effects have both been reported, but not ubiquitously, from the Gakkel Ridge, the Mohns Ridge and from the slow-spreading SWIR (Dick et al., 2003a,b, 2004; Pedersen and Hellevang, 2004; Cannat et al., 2004). The virtual lack of segmentation on the Gakkel Ridge may partly be because separation is already perpendicular to the original (and now inherited) line of continental splitting, but this does not apply to its absence on parts of the very oblique-spreading SWIR. Finally, cumulate olivine cannot crystallize on the now-poorly-defined crack walls, so seismic anisotropy should be absent. Despite these differences in action, heat from the intrusion will still bring about extra dilatation in the walls at the operative phase-change level, but in this case, instead of closing the crack first, that dilatation will push directly on the quasi-solid intrusion, thereby quite possibly generating more push-apart force than when juxtaposing just-crystallized walls after closing the crack first, as in the faster variants.

This volume dilatation within the ‘crack’ walls will not be confined to horizontal expression but will seek relief in the other two directions also. This will be inhibited (mutually) in the along-strike direction but possible in the vertical direction and this offers a basis for explaining the intermittent uplift responsible for the ruggedly ridged flank topography that is characteristic of all except fast spreading ridges. The problem of how median valley floor attains the elevated level of the flanks was noted, but not successfully resolved, by Osmaston (1971).

The push-apart prising action implies the generation of a 'suction' force around the bottom of the crack to draw mantle material into it, and having a magnitude that corresponds to the push-apart force generated by phase-change. This, as we shall see, is a very important result.
Figure 5. Characterization of the different levels in the cratonic tectosphere as discussed in the text. It is presumed that splitting is achieved by initiating a mechanically induced diapir at the base, which then completes the splitting diapirically so that the lateral boundary of the old tectosphere is near-vertical. If splitting progresses, young, much thinner plate will be created in between by progressive wall accretion to those walls, as in Figure 4. In this way cratonic deep keels retain their identity (except where young STE may have carved it away) and a coupling to forces from deep quasi-horizontal flows of the upper mantle, as inferred in this paper. The diagram is specifically directed at Archaean constructed tectosphere but is thought to be applicable in modified form (e.g. thinner depleted zone) for some way into Proterozoic time.

Construction and characterization of cratonic deep tectospheric keels (Figure 5)

The construction of Archaean continental crust by the PSM intrusion of TTG into greenstone forearcs was outlined earlier. This, as mentioned, will have left in place below the belt the oceanic plate and crust from which the TTG derive. In generating that crust the mantle part of that plate will have been depleted at an MOR. This depleted material is identified in two ways; it is the 180-200 km thick zone from which most kimberlite nodules are derived; it is also seen seismologically as a high velocity 'lithosphere'. This 180-200km limit of depletion supports that komatiite genesis was from a wet mantle (Asahara and Ohtani, 2001); for a dry mantle the source would need to have been at nearly twice this depth to achieve the required high-melting (Armstrong et al., 2005).

Although still regarded as tentative by many seismologists, the unusual behaviour of the 410 km and 520 km seismological discontinuities beneath Archaean cratons compared to that under oceans has been taken to imply that their tectospheres are at least that thick (Gu et al, 1998; Agee, 1998). Other seismological investigations, in places where young magmathermal effects (e.g. kimberlite events) are probably absent (e.g. western Australia, Simons et al., 1999; Siberian craton, Priestley et al., 2003), find an apparent LVZ at 200 km and beyond, leading them to infer that the cratonic keel extends no deeper than 200 km, with a substantial drop in wave velocity below that. How can these results be reconciled?

The deeper mantle xenoliths exhibit an increasing degree of C,H,O,S, metasomatism by fluids (Wyllie, 1987; Burgess and Harte, 1999; B. Harte, pers comm 2000) clearly coming from below. Seismologically, below the high velocity 'lid', the inverse correlation of bulk soundwave to shearwave velocity closely resembles that of the Pacific LVZ (e.g. G. Masters, pers. comm. 2002). In the context of the perspective reached above that the latter is not mobile, an identical conclusion seems applicable here too. The presence of these fluids strongly suggests that this underlying mantle is Archaean or early Proterozoic still-somewhat-wet mantle that was brought into place, perhaps with the depleted oceanic 'lid' above. That mantle would bear a level of depletion appropriate to its age but certainly would be richer than the present sub-oceanic upper mantle, and therefore be denser because of higher garnet content. It was argued by Jordan (1979; 1988) that the mantle in this position could not be garnet-rich because that would make it 'drop off', but it is inferred here that the fluid content and the steeper thermal gradient (see below) must offset the denser mineralogy.

A further property conferred by interstitial fluid in this zone is of interest. It is a matter of physics that, because of a change in the mode of heat conduction,
liquids generally have a significantly lower conductivity than the parent solid, so it was argued (Osmaston, 1977) that interstitial fluid in LVZs generally should exhibit low thermal conductivity. More recently the very much lower conductivity of silicate melt has been confirmed experimentally (Snyder et al., 1994). It appears that this property is part of the reason that, as mentioned earlier, even a 130 Ma oceanic plate still retains enough heat at depth to show up during subduction (see also the note about this in Fig. 4). Very possibly this is why, as has long been recognized, the mantle heat-flow into the base of the plate is of similar age. By the same reasoning, the mantle geotherm, rather flat to ~180km, must steepen beyond that, and is confirmed by temperature data on xenoliths (Nixon et al., 1981; Wyllie, 1987).

### Preliminary conclusion on plate thicknesses

The near-axis thickness of the oceanic plate generated at MORs by the new model (Fig. 4) depends on further assessment of a number of factors, including petrogenetic considerations, that are outside the scope of this paper. It seems likely to exceed 60 km, perhaps substantially, and is likely to increase with age, but much more slowly than in the existing models.

The subduction downbend mechanism outlined in Section 2 and Figures 2 and 3 escapes from the constraint, imposed by the elastic flexure concept, that outer rises demand thin flexible plates. Its through-plate step-faulting replacement is not particularly constrained by the plate thickness at the time of subduction. An interesting effect, however, is that at older-plate downbends, graben, rather than steps, are increasingly prevalent on trench outer slopes. The graben widths vary down effectively to zero width, whereas the step ‘treads’ seldom go below 5 km width. This suggests that as the vertical extent of the fault increases (thicker plate), imperfection of their cylindricity increasingly causes the faults to yawn upon displacement.

It is noteworthy in this context that microseismicity occurs to a depth of >70 km in the subducting plate below the flat interface of the NE Japan (Tohoku) forearc (Hasegawa et al., 1991; Zhao, 2001), nearly three times the elastic thickness (28 km) calculated on the elastic flexure assumption by Caldwell et al. (1976) from the outer rise nearby, where the plate is of similar age.

A final item in the thin-plate view of oceanic plates has been the presence of moats around volcanoes. Recognition of mantle phase-change properties, already discussed, appears in principle to offer an alternative interpretation compatible with plates being thick (Fig. 6) but without constraint upon the actual thickness, apart from any petrogenetic ones from the volcanic product.

The plate/tectosphere thickness of cratons (Fig. 5) may lie somewhere between 520 km and the base of the upper mantle at 660 km. That thickness has NOT been achieved by progressive cooling over time but may have remained little changed since it was constructed below its corresponding crust. The properties seem to depend greatly on the characteristics of the Archaean mantle. After its disconnection from the lower mantle, the mobile part of the upper mantle would, in the Proterozoic, have progressively lost these by the depletion and drying that has finally led to its present state. So it is not clear at present for how long through the Proterozoic this deep-keeled mode of construction may have prevailed.

Since the observation of intrinsic properties, as discussed above, is not able to provide firm answers as to thickness of cratonic tectospheres we will approach the matter (below) from the perspective of the actual mechanical behaviour in the plate tectonics realm.

### BACK-ARC OPENING AND STRONG RIDGE PUSH: RESOLUTION OF THE PARADOX?

Since the early proposal of Karig (1971) it has become widely believed that the opening of back-arc basins during active subduction shows that subduction is driven by slab pull, of which roll-back has been seen as a possible consequence. Recognition that some subduction zones show firm evidence of a compressive environment in the upper plate (Uyeda and Kanamori, 1979; Uyeda, 1983; Jordan et al., 1983) had led Uyeda and Kanamori (1979) to refer to these as ‘Chilean type’ as distinct from ‘Mariana type’. Nevertheless, Hamilton (1988) ignored this paradox, asserting firmly (and influentially) that subduction produces an extensional state in the upper plate.

Our emphasis, in preceding sections of this paper, of the wide importance of, and evidence for, compression in the upper plate, together with the difficulty of achieving extensive STE and downbend advance by means of slab pull, merely serves to underline the severity of the paradox. Fortunately, however, this is yet another case in which the thick-plate perspective yields a benefit. The phenomenon known as anticlastic curvature, illustrated in Figure 7, is well-known to engineers (e.g. Case et al., 1999) and familiar to anyone bending a thick sheet of metal or plastic without clamping it, and is due to the along-axis extrusion of material compressed at the inside of the bend. The thicker the plate, the stronger the effect.
Figure 6. Sketch of the way in which, within a thick-plate paradigm, the very high volume change per joule associated with thermal displacement of the garnet-peridotite to spinel-peridotite phase change (see text), and vice versa, may be the cause of moats around oceanic volcanoes and not the result of flexure under load. In the waning stage, reversal of the phase change, in response to reduced heating, begins nearest to the volcanic axis.

Most back-arc basins that formed while subduction was active - Mariana, Ryukyu, Kuril, Banda, Cretan Sea – demonstrate that the ends of the arc have remained pinned, being excluded from the basin-forming motion. This discrimination is not to be expected of roll-back. The stronger effect with a thicker plate also applies to where it is older, e.g. for the Mariana arc, where it also makes the downbend steeper in the middle (e.g. Fukao et al., 2001), as expected from Figure 7.

Early on, Uyeda and Kanamori (1979) drew attention to the now well-established very low level of shallow and intermediate depth seismicity experienced by the Marianas as marking a minimal degree of seismic coupling. This suggests that the apparent restriction of both seismic coupling and STE action to plates younger than about 70 Ma may be due not only to higher thermal buoyancy but also to a lack of significant anticlastic curvature effects.

Japan Sea and Lau Basin are Exceptions; ‘sideways’ displacement of the arc by convergence vector partitioning, attributable to prior segmentation by STE (Osmaston, 1992a) seems applicable here (right-lateral shear in the Japan Sea case (Jolivet et al., 1999)) but is outside the scope of this paper. As discussed below, the Aleutian Arc has only pre-arc floor behind it.

Figure 7. Resolution of the subduction paradox. Anticlastic flexure makes curved arcs; roll-back does not. The proposed anticlastic curvature mechanism (see text) whereby anticlastic flexure of the downbending plate in the presence of strong ridge push (needed to drive the downbending action) can cause the bowing of island arcs and the opening of back-arc basins behind them, all without the help of slab pull.

CHEMICAL AND MECHANICAL EFFECTS OF DEEP-KEELED CRATONS IN THE CONTEXT OF A TWO-LAYER MANTLE

Chemical effects.
The descent, into the lower mantle, of stishovitic lumps of subducted and transformed oceanic crust (to
build D") requires that, with 2-layer convection since 2.22 Ga, a compensating volume of lower mantle material migrates upward through the 660 km discontinuity into the TZ. If as much as half the crustal volume subducted annually (Crisp, 1984) succeeds in entering the lower mantle as dense lumps it is readily calculated that the global upward migration ('seepage') rate would currently be about 2.5 mm/century (25 km/Ga) a rate possibly achievable by diffusion rather than by puncturing the boundary. It was contended by Anderson and Bass (1986) that the velocity/depth profile through the TZ, arrived at in the global seismic velocity model PREM, could not be satisfied isochemically. Maybe this reflects the presence of this seepage zone material.

If craton tectospheres are thick enough that splitting them draws up between them late-Archaean-degree rich (and volatile-rich) ex-lower mantle material from this 'seepage' zone within the TZ, we should see this in the chemistry of the resulting magmas. There is good evidence of this in two situations. The Dupal isotopic signature (Dupré and Allègre, 1983; Hart, 1984) implies a contribution from a source that has not suffered depletion since 2.5-3.0 Ga. Notably, this signature overprints both OIB and MORB in much of the Indian Ocean (Hart, 1984; Osmaston, 2000a) which, as we know, has mostly been made by the separation of cratons. The other situation is in the case of the so-called Iceland plume. It is now recognised petrologically that the high North Atlantic magmatism is characterized by higher water content, centered upon Iceland (Nichols et al, 2002), rather than by the higher mantle temperature envisaged by geophysicists, and recent observations have shown that this wetness extends for a short distance along the Gakkel Ridge (Michael et al., 2003a). Such wetness is the other anticipated characteristic of lower mantle material that has entered the TZ seepage zone. The separation here was of the Greenland craton from the Archaean Scottish Western Isles and from the Baltica craton further north.

These results tell us that cratonic keels do indeed go deep enough to stir up the TZ seepage zone material, or to draw upon it as they separate, but that is not definitive of the depth into the TZ because the seepage layer has an unknown thickness, both because we do not know the integrated input to it since 2.2 Ga and because it has clearly been depleted by being drawn upon and stirred throughout that interval, not least by the 660 km ponding of subducted former slabs. So it is unlikely even to have a globally uniform thickness.

Mechanical-tectonic effects.

With a two-layer mantle, the 660 km discontinuity has evidently behaved as a barrier to large-scale upward movement of mantle, except at the very slow diffusional rates that have built the seepage zone in the TZ. If cratons have keels that extend right to 'the 660', splitting one to create an ocean between must raise the immediate difficulty of where to get the mantle from to put beneath that widening ocean. Likewise, the convergence of two cratons must expel mantle 'sideways' from between them; thus the westward ‘indenter’ action of the Moesian cratonic microplate (Bulgaria-Romania) may result from Arabia-Baltica/Russia convergence (Osmaston, 2005b). The examples discussed below may not demand that cratonic keels actually reach the 660 but if the arguments given by Osmaston (2003) for a physical engagement between the East Antarctica keel and the top of the lower mantle are sustained then this possibility for other cratons will have to be entertained.

Atlantic opening.

In the opening of the South Atlantic the mantle supply problem does appear to have been encountered but solved by the drawing of Pacific mantle through the Caribbean (see Alvarez, 2001) and Scotia gaps, both of whose floors have, on abundant evidence, moved or spread eastward, and are probably continuing to do so. A possible sign that the South American plate is indeed almost 'scraping the floor' as it moves westwards relative to the 660 was the deep Bolivian earthquake of 1994 for which the analysis showed multiple horizontal rupture at close to 636 km depth (Beck et al., 1995; Kirby et al., 1995). For the North Atlantic a similar problem arose, so the inferred sub-Caribbean mantle flow must have done, and still be doing, service there too, of which the Puerto Rico Trench at its northeastern margin may afford evidence.

But what actually motivated these flows? The standard MOR model, with divergent mantle flow at the axis, requires the assistance of slab pull to induce the upwelling. The magnitude of that theoretical slab pull force has depended in major part upon the density of the slab beyond 300 km depth, and especially of the part thought to enter the lower mantle (Minear and Toksöz, 1970). Our new perception of what is actually happening at subduction zones has to be that the presumed high density of slabs is being lost by reheating and that the sinking of detached lumps entering the lower mantle can exert no, or very little, pull (Osmaston, 2003). For these reasons it is no longer appropriate to argue that the South Atlantic is being 'pulled open' by the action of sub-Andean slab pull or, even more pointedly, the North Atlantic by the (now largely absent) subduction under western North America, and the extensive flat subduction under it during the Sevier-Laramide sequence 95-44 Ma.
Our new MOR model, however, as we have shown, should generate much more ridge push than the standard divergent mantle flow model. More important in the Caribbean-Scotia context is that the model also keeps itself, and the growing space beneath it, supplied with mantle by means of the 'suction' force, of comparable magnitude, that it generates at depth below the axis. This importance is because the foreland-directed thrusting exhibited by the Andean subduction suggests that the EPR ridge push may approximately neutralize that of the SMAR in the opposite direction.

Gakkel suction and Eurekan compression.

We will now consider the northward extension of the North Atlantic past Greenland and into the Arctic. As mentioned in the Introduction, it apparently began as a split at about 83 Ma (chron 33) in the Gulf of Labrador (Srivastava and Roest, 1995), being followed only later by a split up the eastern side of Greenland (chron 24, early Eocene). The splitting of the Lomonosov Ridge from the southern (shelf-edge) margin of the Eurasia basin prior to Chron 24 is not seen in the spreading-type anomalies but is marked by a magnetically muted zone 50-100 km wide next to the Lomonosov Ridge and by a not-so-quiet zone along the southern margin (Vogt et al., 1979; Glebovsky et al., 2000; Brozena et al., 2003). It is likely that this is due to the anomaly-muting effect of heavy syn-separation sedimentation, long known from the Gulf of California (Larson et al., 1972; Klitgord et al., 1974) and more recently from the area west of the Antarctic Peninsula (Larter and Barker, 1991). This may also be the cause of the similar situation in the Gulf of Bothnia.

In northern Greenland both Paech (2001) and Lyberis and Manby (2001) found signs of a pre-Eurekan deformation with some south-vergent features. It is suggested that this actually records strong thermal upwarping (to be expected from phase changes within a thick-tertospheric craton) of the northern edge, associated with this early initiation of the Gakkel Ridge, so is not of the same kind as the Eurekan deformation. The pre-64 Ma intraplate alkali basalt magmatism at Kap Washington (Estrada et al., 2001) seems to support this picture.

The temporal dependence of the Eurekan compression upon Greenland becoming an independent plate (Tessensohn and Piepjohn, 2001) strongly suggests that this plate was the one having force applied to it. I therefore infer, as foreshadowed in the Introduction, that the responsible force was due to a northward deep upper mantle flow, in the northern Atlantic region, to satisfy the demand for mantle by the suction of the Gakkel Ridge.

If the Gakkel Ridge suction argument is to be applicable we must assume, therefore, that this suction was in existence before the Eurekan compression, and has continued since. To conclude the paper this is briefly discussed next.

Arctic Ocean demand for mantle, the Indian collision and the Aleutians.

The Arctic Ocean is circumscribed by cratonic blocks with presumably deep keels. Apart from the present Greenland-Baltica gap, just discussed, two notable gaps are the West Siberian basin and the Bering Strait. The extent to which the Verkhoyansk foldbelt overlies another potential gap is is unclear, but if the effective fulcrum of Arctic opening is currently near the Lena delta, the resulting compression at depth below the belt may mitigate against its use as a mantle flow path. Opening of the present ocean basins is thought to have begun as early as Late Jurassic (Vogt et al., 1979). The West Siberian gap has been a permanency throughout this period, and longer. Southward of this gap, Kazakhstan exhibits an intensely disrupted tectonic structure, suggesting that any keels of the scattered older elements do not form a coherent barrier to the supply of mantle northward. I suggest, therefore, that India's northwards compression is identical in nature to that inferred above for the action of Arctic MOR suction upon Greenland. In this case, however, there is, around southern India, a globally uniquely deep (>80 m), and hitherto unexplained, depression in the unfiltered geoid (e.g. Nerem et al., 1994) which might be the dynamical result of that suction. The extraordinarily widespread disruption across central Asia associated with the Indian collision seems consistent with the squeeze force being applied right across the Asian plate. Speculating still further, it might have been this suction that detached India from Gondwanaland at about the time that the Arctic began to open, and brought about its northwards 'flight' and eventual collision. Superficially rather far-fetched, this would necessarily depend upon the suction flow being channelled by an effective corridor of cratonic keels, laying onus upon the layout of cratonic dispositions at the appropriate times.

The Bering Strait gap, by contrast, appears to be quite young in origin. Various reconstructions have been proposed but will not be discussed here. The Aleutian Arc, thought to have been initiated in the Palaeocene (there is 'no pre-Eocene record' (Vallier et al., 1994)), is not the result of back-arc extension (Cooper et al., 1976) but its perfect arc is original and appears to have been initiated by the convergence of sub-Pacific mantle flow upon the Bering gap. If correct, this timing offers some constraint upon Arctic
reconstructions and more particularly upon the development of a Bering Strait gap between tectospheric keels.

A final point is that the Gakkel Ridge exhibits marked along-strike changes in the fertility of its source mantle (Michael et al., 2003b; Hellebrand et al., 2003). These differences may reflect the upper mantle streams that supply it from beneath three different oceans.

CONCLUDING DISCUSSION

It has been shown that the Eurekan compression was the product of forces applied to the deep tectospheric keel of Greenland by the deep upper mantle flow needed to supply the early MOR (Gakkel) spreading, and the upper mantle volume beneath it, forming the Eurasia Basin. This is one of several plate motions showing the action of this mechanism. One appears to be the eastwards motions of the Scotia and Caribbean plates, drawing Pacific upper mantle to put under the widening Atlantic. Another, again due to the demands of an opening Arctic basin largely rimmed by deep cratonic keels, except at the major West Siberian gap, may be the northwards detachment of India from Gondwanaland and its 'flight' to the collision with Asia, within which it has done so much damage. Another, but relatively minor, consequence may have been the initiation of the Aleutian Arc by the convergence of Pacific mantle flow upon the Bering gap.

For this kind of explanation to be available three conditions must be satisfied and it has been reasoned here that they are. These are:-

(A) the 660 km discontinuity must be a barrier to mantle flow (but not to slow chemical diffusion), requiring a new perspective on the subduction process that owes much to appreciating the character and evident heat content of the oceanic plates that feed it;

(B) Archaean-early Proterozoic cratons have tectospheric keels that extend to, or nearly to, the base of the upper mantle, having largely been constructed like that, not developed by progressively deeper cooling;

(C) the MOR plate-constructing process is that of a new MOR model that (i) incorporates the oceanic LVZ as an integral part of the plate, and (ii) produces a sub-plate suction to provide it with the mantle it requires, not only for incorporation into the oceanic plate but for the entire upper mantle below the widening ocean.

A direct consequence of (A) is that slab pull can no longer be the major player that drives plate tectonics. Evidence (Atwater et al., 1993; Searle et al., 1993) of important changes of plate motion at the times of the beginning and end of the long Cretaceous Superchron, however, led Osmaston (2003) to reason that polar electromagnetic coupling clockwise torque at the CMB, transmitted through the high-viscosity lower mantle, has been picked up by the deep tectospheric keel of East Antarctica and that this has been the main plate tectonic agent for at least the past 150 Ma. In this context, the actions include the initiation of the South Atlantic and the anticlockwise rotation of Africa. It may also be noted that the initiation of the opening of the Gulf of Labrador at Chron 34-33 (83.5 Ma) is among several global events now known to be temporally tied to the end of the Superchron.

Conversely, with no craton currently occupying the Arctic, to pick up a corresponding torque, this could explain why plate tectonics is relatively inactive in the northern hemisphere. If accepted, this line of reasoning provides additional support for (B).

The thick-plate perspective developed here is clearly in conflict with that based on interpretations of apparent plate flexure. Such geophysical interpretations seem never to have taken account of the substantial differential vertical motions that mantle phase changes, now seen to occur within the thickness of the plate, could produce as the result of quite small redistributions of temperature. An obvious example (Fig. 6) is that in its dying stages, the mantle duct supplying an intraplate volcano will transmit less heat to the surrounding mantle, enabling a reversal of phase-change to occur there, with resulting subsidence that is not the product of loading by the edifice. Another case, as mentioned at the outset (Fig. 2), is that outer rises at trenches are no longer a measure of plate flexural rigidity. Thirdly, and uniquely, the thick-plate perspective renders applicable the anticlastic curvature mechanism (Fig. 7) for the curving of arcs and opening of back-arc basins in the presence of strong ridge push, thus resolving a paradox of long standing (Uyeda and Kanamori, 1979).

Finally, the recognition here that the presence of low seismic velocities directly below the high-velocity lithosphere, both in the oceans and under cratons, does NOT signify mobility of that material, requires an important change of terminology. The mobility term 'asthenosphere' which currently is applied to this material, by default, is no longer appropriate unless its use is justified in that particular case.

Each of the foregoing new perspectives is hardly acceptable as a stand-alone change to the present form of the plate tectonics paradigm. It is only when they are put together that the real benefits emerge and lead to the remarkable conclusion that upper mantle flows associated with MOR activity in the Arctic have been a major agent in global tectonics, at least throughout the Cenozoic. This result offers a new dimension to the debate on the form of mantle convection, previously
limited to seismology and geochemistry.

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REGIONAL PALEOTECTONIC INTERPRETATION OF SEISMIC DATA FROM THE DEEP-WATER CENTRAL ARCTIC

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ABSTRACT

We examine regional features of sedimentary cover within the “Makarov Basin–Lomonosov Ridge–Amundsen Basin” area where seismic reflection investigations were carried out by Russia from “North Pole” drifting ice-stations. Analysis of the acquired information allowed us to determine a number of regularities in the seismic reflection configuration of the sedimentary cover, as well as, some peculiarities of the seismic facies pattern. Using available data on the periphery of the Amerasian Basin, a paleotectonic interpretation of the main features of the sedimentary cover is offered in advance of data received from the upcoming ocean drilling on the Lomonosov Ridge.

INTRODUCTION

The Arctic Ocean is divided into two basins, the Eurasian and Amerasian. The Eurasian Basin opened by Cenozoic spreading along the Gakkel Ridge. In contrast, detailed information on the nature and the age of the Amerasian Basin and its subbasins (Canada and Makarov basins) is not available. The opinions about their structure are mostly based on speculation. According to the conventional model (Grantz et al., 1998), the oldest Arctic deep-sea basin (Canada Basin) was formed in the Cretaceous by seafloor spreading. During or after the opening of the Canada Basin, the Makarov Basin formed.

All main tectonic events that have occurred in the Arctic Ocean during the development of the Makarov Basin should be embodied in its sedimentary cover as unconformities or non-depositional hiatuses.

The purpose of this paper is to present the seismic reflection data within the “Makarov Basin–Lomonosov Ridge–Amundsen Basin” area, although we include only part of the Amundsen Basin. This Central Arctic region was investigated by the Russian long-lived ice-stations “North Pole” (NP) during 1973-1988 (NP21-NP28), which collected more than 10,000 km of seismic reflection data. We discuss a paleotectonic interpretation of the main features of the seismic reflection configuration of the sedimentary cover.

DATA ACQUISITION

During the years 1973-1988 data acquisition techniques from NP drifting ice-stations varied little. The spacing of the seismic reflection soundings was irregular because of the variable drift speed of the ice floe. Average spacing was usually about 800 m. An end-on or central 24-channel spread of dynamic geophones with length of 1150 m and analog magnetic recording were used. For the reflection shots, 3-5 electric detonators were fired at a depth of 3-5 m below the water surface.

To process seismic reflection data, the acquired analog information was recently digitized with 2 ms sample rate. Digital data were processed by means of the ProMAX 2D Software. For an enhancement of the signal-to-noise ratio, all seismic data were stacked after Shot Point (SP) number sorting (but not Common Depth Point number sorting) and after formal Normal Moveout correction of SP gathers. Owing to narrow seismic aperture in the deep water, velocity analyses were not performed. A number of processing procedures such as an accurate interactive spectral analysis and Time and Space-Variant Frequency Filtering before stacking, as well as, Trace Math Transforms after stacking, were applied to enhance the data. The processed data characterizes well the seismic reflection configuration of the sedimentary cover along the tracks.

RESULTS

As seen on the data collected from NP28 (Fig. 1), a strong reflector is traced in sedimentary cover of Makarov Basin (within each bathymetric level of the basin). It is the main seismic marker in the sedimentary cover of the Makarov Basin. Using data from NP23, NP24, NP28 and published data acquired by R/V Polarstern in 1991 and 1998 (Jokat et al., 1995; Jokat, 1999), a collection of sections across the “Lomonosov Ridge–Makarov Basin” area were compiled (Fig. 2). A strong similarity in seismic reflection patterns indicates that this seismic marker extends over both the Lomonosov Ridge and the Makarov Basin. A discordance between the marker reflector and overlying or underlying reflectors can be seen on different fragments of the sections (Figs. 1, 2). While this discordance is apparently characterized by onlap or toplap type signatures in the Makarov Basin (Figs. 1, 2), the erosional truncation type seems to be unique to the upstanding blocks of the Lomonosov Ridge (Figs. 2, 3). Furthermore, it is possible to see the strong similarity in seismic reflection character of seismic
facies under the seismic marker both in the Makarov Basin and on subsided blocks of Lomonosov Ridge (Fig. 2). At the same time, these seismic facies are not displayed identically on upstanding blocks of Lomonosov Ridge (Fig. 2, 3).

What this seismic marker represents is the key question.

**DISCUSSION**

Taking into consideration visible attributes of discordance and broad occurrence of the main seismic marker, we interpret it to be the regional unconformity in the investigated area. The regional unconformity appears to divide sediments into two stages that may have been deposited under different depositional environments.

When could this regional unconformity have formed? The age of the unconformity is usually defined from the age of the oldest overlying stratum. It can be seen clearly on NP23 data (Fig. 3) that the regional unconformity is covered by older strata in the Makarov Basin than on upstanding blocks of the Lomonosov Ridge. In addition, the seismic data show that a significant part of sediments under the regional unconformity were eroded from upstanding blocks of the Lomonosov Ridge (Fig. 3). In other words, the non-depositional hiatus corresponding to the regional unconformity increases significantly from Makarov Basin to the upstanding blocks of Lomonosov Ridge. Therefore, we should formulate the question about the age of the regional unconformity a little differently, namely—what tectonic event could induce a regional non-depositional hiatus in the Central Arctic region?

According to available geological data from the periphery of the Amerasian Basin, the stratigraphic interval between Late Oligocene and Early Miocene is characterized by regional horizons of chemical weathering (horizons of speckled clays) (Table 1). The Late Oligocene-Early Miocene interval of chemical weathering is up to 8 m thick in the Laptev Basin, up to 20 m thick in the East Siberian Basin, and up to 40 m thick in the Chukchi Basin (Kim et al., 1991). In the Beaufort-McKenzie Basin, the prominent Late Oligocene seismic marker TK is tied to the top of the Kugmallit sequence (Enachescu, 1990), which also has a chemically weathered interval up to 40 m thick (Kim et al., 1991) (Fig. 4). The seismic sections of the Beaufort-McKenzie Basin (Canadian Geological Atlas, 1995) show that seismic marker TK can be interpreted either as a boundary between structural stages (Fig. 5,

![Figure 1. Seismic section on drift line NP28 across Makarov Basin showing three bathymetric levels and the relation and character of the regional unconformity (RU).](image_url)
Figure 2. Tracing of the main seismic marker within the “Lomonosov Ridge – Makarov Basin” area.
Figure 3. Seismic section on drift line NP23 along Lomonosov Ridge and edge of Makarov Basin showing the regional unconformity (RU) and its variation in seismic character.

Table 1. Geological data on the periphery of Amerasian Basin (Kim et al., 1991)

<table>
<thead>
<tr>
<th>Stratigraphic interval: Late Oligocene - Early Miocene</th>
</tr>
</thead>
<tbody>
<tr>
<td>Place of geological survey</td>
</tr>
<tr>
<td>----------------------------</td>
</tr>
<tr>
<td><strong>Laptev Basin</strong></td>
</tr>
<tr>
<td>Bolshevik Is., Svyatoy Nos Cape, Lena Delta</td>
</tr>
<tr>
<td><strong>East-Siberian Basin</strong></td>
</tr>
<tr>
<td>Wrangel Is., Chaunsk Bay, Foothills of Kular Ridge</td>
</tr>
<tr>
<td><strong>Chukchi Basin</strong></td>
</tr>
<tr>
<td>Shmidt’s Cape, Daurkin Peninsula, Walen Lagoon</td>
</tr>
</tbody>
</table>

Figure 4. General geology of Beaufort-Mackenzie Basin. Amauligak area (Enachescu, 1990, Fig. 3).
Figure 5. Seismic sections of Beaufort-Mackenzie Basin showing the Late Oligocene marker horizon “TK” and its variation in seismic character (from “Canadian geological atlas of the Beaufort-Mackenzie area”, 1995, Fig. 27, 28).
Top), or as a planed (eroded?) surface (Fig. 5, Bottom). In addition, a distinct erosional surface between Oligocene and Miocene sediments has been observed on the Canadian Arctic Archipelago, for example on Banks Island (Kim et al., 1991).

Thus, the Late Oligocene is associated with a thick horizon of chemical weathering that broadly extends along the periphery of the Amerasian Basin. Hence, we can presumably connect a regional non-depositional hiatus in the deep-water Central Arctic region with the global Late Oligocene minimum. This eustatic event corresponds to the largest known global fall of sea level in the history of the Earth (Vail et al., 1977).

There is an alternative paleotectonic interpretation of the regional unconformity in the Central Arctic region. W. Jokat, in his recent work (2003), suggested that the regional unconformity was caused by a major tectonic event in the Arctic Ocean that significantly changed the sedimentary environment. In his opinion, the most obvious candidate is the beginning of the Eurasia Basin opening at the Paleocene-Eocene transition (Jokat, 2003). But in that case we should not see a regional unconformity in the sedimentary cover of Amundsen Basin.

It can be seen clearly on NP21 data (Fig. 6) that the same regional unconformity (or very similar in seismic reflection character) is traced from Lomonosov Ridge to the Amundsen Basin. Furthermore, we interpret a thick sedimentary sequence under the regional unconformity in the Amundsen Basin (Fig. 6). Hence, the regional unconformity began to develop much later than the beginning of the Eurasian Basin opening. That is why we regard connecting a regional non-depositional hiatus to the global Late Oligocene minimum as the most probable paleotectonic interpretation.

In the Makarov Basin area, we interpret identical seismic facies under the regional unconformity both in the Makarov Basin and on subsided blocks of Lomonosov Ridge (Fig. 2). According to Sangree et al. (1997) the seismic reflection character of sandy-argillaceous facies is mainly dependent on water depth of deposition (shelf, slope/rise, deep water). In our opinion, the type of seismic facies underlying the regional unconformity corresponds to shallow shelf facies based on the seismic facies units classification (Sangree et al., 1997). Therefore, these facies could have accumulated under identical shallow shelf

![Figure 6. Seismic section on drift line NP21 from Amundsen Basin to Lomonosov Ridge.](image-url)
conditions both in the Makarov Basin and on subsided blocks of Lomonosov Ridge.

Based on the above evidence, we believe that the Makarov Basin existed as a trough (possibly under shallow marine conditions) long before the Late Oligocene regional non-depositional hiatus. After that (i.e. in Miocene), the Makarov Basin began to develop as a pelagic sedimentary basin (upper sedimentary stage). Miocene subsidence of the Makarov Basin was probably caused by influence of intraplate tectonics.

CONCLUSION

More than 10,000 km of seismic reflection data collected from Russian “North Pole” ice-stations image the seismic stratigraphic configuration of the sedimentary cover in the deep-water Central Arctic region. Analysis of the seismic configuration of the sedimentary cover makes it possible to interpret a regional non-depositional hiatus, which divides sediments into two stages deposited under different depositional environments. Detailed consideration of the variety of data from the periphery of Amerasian Basin allows us to connect the regional non-depositional hiatus with the global Late Oligocene eustatic minimum. Because of the apparent correlation of seismic facies under the regional unconformity both in the Makarov Basin and on subsided blocks of Lomonosov Ridge, we propose that the Makarov Basin first began to develop as a pelagic basin after the regional non-depositional hiatus, i.e. in Miocene.

We hope that the problem of paleotectonic interpretation of the regional non-depositional hiatus in the Central Arctic region can be finally resolved in the near future by analysis of data from the Integrated Ocean Drilling Program cruise to Lomonosov Ridge.

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A SEISMIC MODEL OF THE EARTH'S CRUST ALONG THE "EAST-SIBERIAN CONTINENTAL MARGIN – PODVODNIKOV BASIN – ARLIS RISE" GEOTRAVERSE (ARCTIC OCEAN)

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ABSTRACT
The deep seismic sounding (DSS) and reflection experiment was carried out along the Delong Islands – Arlis Rise Geotransect by the Polar Marine Geological Research Expedition (PMGRE) in 1989&1991. Three layers (c. 5.0 km/sec – layer of consolidated bedrock, c. 6.1 km/sec – upper crystalline crust and c. 6.7–7.0 km/sec – lower crust) was followed along the all Geotransect by the DSS data. The overall thickness of the crust undergoes appreciable fluctuations within the limits of the Geotransect, with a maximum thickness of 41 km beneath the East-Siberian Continental Margin, decreasing to 20 km beneath the Podvodnikov Basin and increase to 27 km beneath the Arlis Rise. The East-Siberian continental shelf is a typical example of a continental crust. The 6.1 km/sec layer has been interpretation as "granitic". From the constructed section it is visible, that revealed on a shelf layers are traced in the both Podvodnikov Basin and Arlis Rise also. However these layers are appreciably reduced.

INTRODUCTION
A refraction and reflection methods was carried out along the lengthy (1487 km) Delong Islands – Makarov Basin Geotransect in 1989-1991 by the Polar Marine Geological Research Expedition (PMGRE). It has crossed large morphostructures: the East-Siberian Continental Margin, the Podvodnikov Basin, the Arlis Rise and the Makarov Basin (Fig. 1). The Geotransect (so-called the “Transarctic 1989-1991”) included the deep seismic sounding (DSS), reflection method (with the exception of 1989), measurements of seismic velocities of sedimentary cover, spaced gravity and airborne magnetics within a 100 km wide zone along the transect. The seismic section displayed here is most part (1000 km) of the Geotransect from the East-Siberian Continental Margin to the Arlis Rise. The results of the experiment were published in Russian (Zamansky et al, 1999, Gramberg, 1993) and part of the “Transarctic-1989” by Sorokin et al (1999). The new interpretation of the DSS data is executed with use modern interpreting software package and it will lighten to compared from the Lomonosov Ridge – the “Transarctic-1992” transect (Ivanova et al., 2002) and from the Mendeleev Ridge – the “Arctic-2000” transect (Zamansky et al, 2002) (Fig. 1) supports this interpretation methods.

A summary of the bathymetric and subsurface of the “Transarctic 1989-1991” area is difficult since it is extensive and heterogeneous region of the Arctic Ocean. The East-Siberian continental margin is

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The widest shelf in the world. It has smooth variation of depth from 50 m to 250 m on the shelf, in the Geotransect’s area. The continental rise has two parts: sharply fall to 2000 m and smooth descending to the Podvodnikov Basin. The bathyal plain adjacent to the East-Siberian margin is the Podvodnikov Basin with water depths of about 2800 m. The Arlis Rise poorly develops to the south Podvodnikov Basin (up to 2600 m) and smooth descends to the north Podvodnikov Basin with water depths up to 3200 m.

Magnetic field of the “Transarctic 1989-1991” area has negative smooth numbers from -120 nT to -80 nT above the East-Siberian Continental Margin. This area of the magnetic field has become complicated positive irregular short-period anomalies up to 220 nT. The Geotransect crosses transitional zone of the magnetic field in deeper ocean. This zone divided the super-differentiated mosaic field of the Amerasian Province from the poorly anomalous field of the Lomonosov Ridge (Verba and Petrova, 1986; Leonov, 2000). The magnetic field of transitional zone is described by sublinear west-north-west anomalies, which are registered as periodic anomalies from –320 nT to 550 nT along the Geotransect line.

Gravity field of the East-Siberian Continental Margin has sharp variations from 0 mGal to 55 mGal and up to 75 mGal over the rise of the Margin (Glebovsky et al, 2002). Similar pronounced gravity anomalies are traced along rises of a continental margin in the Arctic Ocean. Resembling pattern of gravity field is present over the Atlantic Margin (edited by Gramberg, 1985). Gravity anomalies of 0/-12 mGal are observed over the Podvodnikov Basin along the Geotransect line and they are smoothly increasing (up to 25 mGal) in the direction of the Arlis Rise.

The reflection profile of ice station NP-28 (Sorokin et al, 1999) and both the reflection and the DSS profile of “Arctic-2000” (Zamansky, 2002) crosses the “Transarctic 1989-1991” (Fig. 1).

**METHODS, TECHNIQUES**

The DSS and reflection experiment presented here was carried out in April of 1989 and 1991. The 1991 profile was recorded from the East-Siberian continental margin to the Podvodnikov Basin along SW–NE line from 153°11’ E / 76°23’ N to 166°01’ E / 81°31’ N, (about 630 km). The 1989 profile forms a continuation of the preceding profile from the Podvodnikov Basin to the Arlis Rise along S–N line from 164°32’ E / 80°53’ N to 170°21’ E / 85°02’ N, (about 470 km). The experiments were carried out by an identical method (Fig. 2). The seismic recorders and the shot points were placed along the line, in three segments. Supplementary segment was carried out for compensation of same breakdown of the recording instruments. Seismic arrivals were recorded by TAYGA-2 analog tape recorders placed on the drift ice with assistance of a MI-8 helicopter. Every TAYGA-2 station had a 6-channel array of low-frequency geophones (resonance frequency 5 Hz). The array’s length was between 200 and 500 m. Seismic recorders was installed near all shots to register the time marks and the shot instant. The location of the sources and the receivers were determined by GPS units (MX-4400 and MX-1502) with an accuracy up to 250 m. The sources were trinitrotoluene (TNT) charges detonated under water a depth from 30 to 100 m. Simultaneously with the DSS researches the base reflection sounding were carried out for determination of velocities parameters in sedimentary cover.

However “Transarctic-1989” and “Transarctic-1991” experiments had some differences. In 1989, the ten recording instruments of segment were installed with interval from 10 km to 14 km. 18 shot points were carried out; the interval between the shot points was 30–70 km; a maximum distance of the shot-recorder was 170 km. It is necessary to note that the PMGRE carried out the first PMGRE’s DSS experiment in the deep-water Arctic Ocean in 1989. It explains low quality of the DSS materials in comparison with the
Figure 3. Geophysical fields and seismic model of the earth's crust on the “Transarctic-1989” and “Transarctic-1991” geotraverses. On diagrams of geophysical fields: GF - the diagram of anomaly gravimetric fields in free air reduction with points of observation; Ta - the diagram of an anomaly magnetic field. On model of the earth's crust: fat lines show seismic boundaries; thin lines are velocity isoline through 0.1 km/s and their numerical values; indexes are revealed seismic sequences of the sedimentary cover and layers of the consolidated crust; letters are indexes of seismic boundaries; triangles designate points of explosion. Sequences of the sedimentary environment be allocated on the crustal model: 1, IIa, IIIb (V=1.7 - 2.8 km/s) are friable and weak lithified deposits; III (V=3.3 - 3.8 km/s), IV (V=5.0 - 5.5 km/s) are lithified deposits of a various degree of consolidation. Crystal layers: V (V= 6.0 - 6.5 km/s) - the upper crust, VI (V= 6.7 - 7.4 km/s) - the lower crust, VII (V*= 7.9 - 8.2 km/s) - the upper mantle.

materials of the DSS experiments in the next years. In 1991, interval between the sixteen recording instruments of segment was about 7 km. Intervals between the shot points were 30–60 km. A maximum distance of the shot-recorder was 210 km.

In “Transarctic-1991” the reflection sounding was carried out in points of DSS receivers.

RESULTS

Model of the Earth's crust along these two profiles has been constructed (Fig. 3) with used interactive package SeisWide (Dr. Deping Chian). Four type of section parts could be distinguished along this profile: 1) the East-Siberian continental margin, 2) the rise of the margin, 3) Podvodnikov Basin, 4) the Arlis Rise.

Sedimentary cover (velocities, thickness, stratification) along the Geotransect was constructed with used data of reflection experiments in the “Transarctic - 1991” and NP-28.

The East-Siberian continental margin has thin veneer of unconsolidated (Vp < 2.0 km/sec) and partially consolidated (Vp = 2.2–3.6 km/sec) sediments of a total thickness from 0.2 to 1.3 km. The underlying layer of consolidated bedrock (Vp = 5.1–5.5 km/sec) has thickness of 0.5–5.0 km. The next layer represented by upper crystalline crust with Vp = 6.1–6.5 km/sec has thickness from 12 to 22 km. It would be desirable to note that the top boundary of this layer is very broken. The underlying lower crust has Vp = 6.7–7.0 km/sec and thickness of 14–19 km. The total thickness of crust
in the East-Siberian continental margin varies from 35 to 41 km.  

The rise of the margin and Podvodnikov Basin has thicker sediment cover. Unconsolidated and partially consolidated sediments ($V_p = 1.7$–$2.8\ \text{km/sec}$) have a thickness of c. 2.0 km. Upper boundary of underlying more consolidated sediment ($V_p = 3.5$–$3.8\ \text{km/sec}$) was traced in first arrival of DSS data. The layer has a thickness of 2.0–7.0 km. The underlying layer of consolidated bedrock has $V_p = 5.0$–$5.2\ \text{km/sec}$ and thickness of c. 4 km. Upper crust with $V_p = 6.1$–$6.3\ \text{km/sec}$ has a thickness up to 6 km, but it is absent (so-called basalt window) under precontinental deep (about picket 400 km, see Fig. 3). The underlying lower crust on this part of profile has $V_p = 6.9$–$7.4\ \text{km/sec}$ and thickness of 6–11 km. Moho depth varies from 18 to 25 km.  

The Arlis Rise has a thin veneer of unconsolidated ($V_p < 2.0\ \text{km/sec}$) and partially consolidated ($V_p = 2.2$–$3.7\ \text{km/sec}$) sediments of a total thickness up to 2.5 km. The underlying layer of consolidated bedrock has $V_p = 5.0$–$5.2\ \text{km/sec}$; the thickness is c. 4 km. Upper crystalline crust with $V_p = 6.0$–$6.2\ \text{km/sec}$ has thickness of 2–6 km. The underlying lower crust on the Arlis Rise has $V_p = 6.8$–$7.2\ \text{km/sec}$ and reaches a maximum thickness of c. 17 km (with maximum of a Moho depth at c. 30 km).

**CONCLUSIONS**

The constructed model of the “Transarctic 1989&1991” have good correlation by both velocities and boundaries depth with the crossing profiles “Arctic–2000” and NP-28.  

The East-Siberian continental shelf is typically developed below the shelf, and rocks with velocity of 6.1 km/sec (e.g. granites) crop out on the Henrietta Island (Vinogradov et al., 1975). From the constructed section it is visible, that the layers revealed on a shelf are traced in the both Podvodnikov Basin and Arlis Rise also. However these layers are appreciably reduced.

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SEISMIC DATA ACQUISITION IN THE NANSEN BASIN, ARCTIC OCEAN

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ABSTRACT
During the NPD-POLAR-2001 expedition to the Arctic Ocean in the fall of 2001 a multi-channel seismic (MCS) and sonobuoy acquisition program was carried out in the sea ice on the Hinlopen margin north of Svalbard and in the Nansen Basin. For this purpose, specially adapted acquisition and processing techniques were developed, resulting in ~1100 km of 2-D MCS profiles and 50 wide-angle velocity profiles. The data may be interpreted in terms of four main sedimentary units with distinct seismic velocities and reflection character.

INTRODUCTION
The Nansen Basin (Fig. 1) has formed by seafloor spreading along the Gakkel Ridge, the Arctic end-member of the Mid-Atlantic Ridge system, since the late Paleocene (~55 Ma) (Karasik, 1968). Due to perennial sea ice, only a few MCS and wide-angle seismic profiles sample the ocean basin north of 81°N (Duckworth and Baggeroer, 1985; Jokat et al., 1995; Kristoffersen and Husebye, 1985). Following Norway’s ratification of the UN Convention on the Law of the Sea in 1996, the Norwegian Petroleum Directorate (NPD) has initiated a research program to map sediment thickness and the foot-of-slope for the purpose of delineating legal outer shelf boundaries beyond 200 nautical miles. As a part of this program, the NPD and the Universities of Bergen and Oslo conducted a seismic survey aboard the Swedish icebreaker MV Oden in the autumn of 2001. Here, we present some preliminary results.

DATA ACQUISITION AND PROCESSING
Major breakthroughs in seismic exploration of the Nansen Basin have been achieved during the two-ship expeditions ARCTIC-91 (Fütterer, 1992) and AMORE (Thiede and the Shipboard Scientific Party, 2002), where the PFS Polarstern collected seismic data in a trail cut open by another icebreaker ahead. However, such operations are costly and involve inefficient sharing of ship-time between two scientific parties. On the other hand the MV Oden, with its 30 m wide bow, is an adequate platform for simultaneous icebreaking and seismic surveying, provided that the seismic equipment is robust and flexible (Kristoffersen, 1997). Because the trail of open water rapidly closes, the airguns and hydrophone cable have to enter the water in the turbulent zone just behind the ship, where variable speed, bouncing ice and high noise is a challenge to the acquisition. The airguns (2×4 l or 1×20 l) and hydrophone cable (300 m active section, 12 channels) were suspended from a light depressor device which was able to move around ice fragments and effectively kept the airguns and the hydrophone cable separated in the water.

The raw seismic data contain geometrical and noise effects typical for ice-acquired data. Because the vessel had to navigate around pressure ridges and speed varied because of ice resistance, the shotpoint distances are not constant and the traces had to be stacked in 25 m bins along the ship track. Furthermore, trace-to-trace noise variation and delayed ghost effects from the lack of hydrophone cable depth control had to be filtered out. Further processing steps involved geometrical spreading correction, predictive deconvolution and normal move-out correction, stacking, and filtering. The processing significantly reduced ringing of seismic reflections. The sonobuoy profiles were not processed after the geometrical binning.

VELOCITY MODELLING
Of the 60 sonobuoys deployed, 50 had sufficient quality and range (up to 28 km) for reduction. Most profiles contained both refracted arrivals from the top of basement and intra-sedimentary reflections. These arrivals were solved for initial velocities using standard slope-intercept and $T^2-X^2$ methods assuming 1-D conditions (plane-layers). Then, velocities were combined with interpreted reflector geometries into 2-D velocity models which were modified by ray tracing and traveltime inversion (Zelt and Smith, 1992). By perturbing the final sediment velocities while recording the traveltime and model $\chi^2$ misfits, we calculated uncertainty in sediment thickness of about ±7%. The 2-D models also show that the original 1-D reduction over-estimates sediment thickness in areas of high basement relief.
Fig. 1. Existing seismic database and new lines of the NPD-POLAR-2001 survey in the Nansen Basin and adjacent shelf. IBCAO bathymetry (Jakobsson et al., 2000), contoured every 0.5 km. AB, Amundsen Basin; AWI, Alfred Wegener Institute for Polar and Marine Research, Bremerhaven, Germany; COT, continent–ocean transition (Engen, 2005); FJL, Franz Josef Land; LR, Lomonosov Ridge; MAGE, Marine Arctic Geological Expedition, Murmansk, Russia; MJR, Morris Jesup Rise; NB, Nansen Basin; NPD, Norwegian Petroleum Directorate; NPI, Norwegian Polar Institute; UIB, University of Bergen, Norway; YP, Yermak Plateau.
Figure 2. Seismic data example and line drawing of the easternmost profile, line NPD-POLAR-15. Four sedimentary units (NB-1–NB-4) overlie oceanic basement. Annotated columns show velocities, in km s$^{-1}$, sampled from the 2-D velocity model in the sonobuoy deployment positions. CDP, common depth point.
RESULTS
The processing yielded ~1100 km of good-quality MCS data imaging the basement surface and the main sedimentary sequences. The sediment thickness increases both from the Gakkel Ridge to the margins and from west to east, where the sharp basement relief is completely covered (Fig. 2). There is good agreement between the velocity models and the seismic stratigraphy that the sediments can be subdivided, from bottom to top, into four regional units (Fig. 3). The boundaries between units probably correspond to major paleoceanographic events such as the onset of late Cenozoic glacial deposition and the opening of the Fram Strait gateway between the Arctic and North Atlantic oceans (Engen, 2005). Faults cut the sediments from basement to shallow levels, indicating recent tectonism (Fig. 2).

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IS GROUNDING OF AN ICE SHELF POSSIBLE IN THE CENTRAL ARCTIC OCEAN? A MODELING EXPERIMENT

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ABSTRACT
A numerical ice sheet model was used in a first test towards evaluating the hypothesis that, during a period of large-scale glaciation, an ice shelf emanating from the Barents/Kara Seas grounded across parts of the Lomonosov Ridge to a depth of around 1000 m below present sea level (Jakobsson, 1999; Polyak et al., 2001). Despite that we not include complex ice shelf physics or grounding line mechanics in our model and treat the process of marine melting in a simple manner, our experiments are the necessary first steps toward providing a comprehensive reconstruction of the former ice-sheet/ice-shelf system in the Arctic Ocean. A series of model runs was performed where ice shelf mass balance and ice shelf strain per unit time (strain rate) were adjusted. The mass balance and shelf ice strain rate are the key model parameters that govern the flux of ice into the Arctic Ocean. Grounding on the Lomonosov Ridge was not modeled when the ice shelf strain rate was 0.005 year$^{-1}$ (i.e. a free flowing ice shelf). Even with low rates (<10 cm/year) of basal melting, the ice shelf thickness was always less than 100 m over the central part of the ridge. Our experiment suggests that grounding on the Lomonosov Ridge by a free-flowing ice shelf is not possible. When the strain rate in the shelf ice was reduced to zero, however, the shelf thickness increased substantially. Such conditions are likely only to have occurred during periods of large-scale glaciation if substantial stagnant and thickened sea ice was present in the ocean, buttressing the ice shelf flowing from the Barents Sea. A comprehensive study using a coupled ice-sheet/shelf/sea-ice model would build on these preliminary results and have the potential to further constrain the history of circum-Arctic Ocean ice sheets.

INTRODUCTION
The extent of Pleistocene glaciations in the high Arctic and Arctic Ocean is a widely discussed and unresolved problem. The notion of a former huge Arctic Ocean floating ice shelf was first developed by Mercer (1970), who pointed out the similarity in physiographic setting between the present-day Arctic and West Antarctica. However, for the late Weichselian the hypothesis of a huge pan Arctic ice sheet has been disproved through recent field investigations within the European Science Foundation program QUEEN (QUaternary Environments of the Eurasian North) (Svendsen et al., 1999; 2004). Such a glacial scenario remains, nevertheless, at least possible for pre-Weichselian marine glaciations, and marine geophysical mapping during recent years has indeed revealed glaciogenic features at substantial water depths in the Arctic Ocean.

Survey of the Lomonosov Ridge crest in the central Arctic Ocean was carried out by deploying a chirp sonar subbottom profiler from the Swedish icebreaker Oden (Jakobsson, 1999). The high-resolution subbottom stratigraphic images reveal large-scale erosion down to a depth of the ridge crest of 1000 m below present sea level (Figs. 1 and 2). Ice grounding was proposed as one possible explanation for this erosion but the data were not conclusive. The hypothesis for ice-grounding on the Lomonosov Ridge was, however, later supported by side-scan sonar and additional chirp sonar data collected in 1999 during the SCICEX expedition onboard USS Hawbill (Polyak et al., 2001). These data revealed glacial fluting at the deepest eroded areas down to about 1000 m water depth and subparallel scours from 950 m depth to the shallowest parts of the ridge crest. The directions of the mapped glaciogenic bed-forms and the re-deposition of eroded material on the Amerasian side of the ridge indicate ice flow from the Eurasian Basin across the ridge (Polyak et al., 2001) (Fig. 2). This interpretation is supported by sea-floor sediment cores, which show that the source of the material derives from the Barents-Kara Sea region (Spielhagen et al., 1997; 2004). In addition, sediment core studies reveal that this erosion took place during Marine Isotope Stage (MIS) 6 (Jakobsson et al., 2001).
Glacial geological evidence strongly suggests that the Late Saalian (MIS 6) ice sheet margin reached the shelf break of the Barents-Kara Sea (Mangerud et al., 1998; Knies et al., 2001). This glacial scenario leads to two hypotheses about the causes of ice erosional features on the Lomonosov Ridge. One is the grounding of a floating ice shelf and the other is the scouring from large deep icebergs. The former could either imply a huge continuous floating ice shelf covering the entire Arctic Ocean, analogous to the hypothesis of Mercer (1970), or an ice shelf emanating from the Barents/Kara Seas. The latter explanation requires the Late Saalian Barents-Kara Ice Sheet to produce huge deep icebergs that drifted within the permanent Arctic sea ice toward the Lomonosov Ridge crest, perhaps clustered together in an armada of icebergs as proposed by Kristoffersen et al. (2004). The St Anna Trough is the most prominent glacial trough on the Barents-Kara Sea margin and a likely source for the large iceberg during Pleistocene glaciations. Support

Figure 1. Map showing the bathymetry of the Arctic Ocean, the maximum extent of the Saale ice sheet (Svendsen et al., 2004), and the main direction of ice scour features on the Lomonosov Ridge (Polyak et al., 2001).
for the iceberg hypothesis may be the discovery of deep iceberg ploughmarks on the Yermak Plateau that are mapped to more than 850 m below the present sea level (Vogt et al., 1994).

One conclusion of the QUEEN project was that the Saalian ice sheet was far larger than those of the Weichselian (Svendsen et al., 2004) (Fig. 1). The grounded margin of this ice mass would have reached the continental shelf break of the Barents/Kara Seas, within bathymetric troughs (i.e. Voronin, St Anna, Franz Victoria, and Bjørnøya Troughs), at depths of several hundred meters below sea level.

Although the idea of a thick Arctic Ocean ice shelf has been discussed by several authors (e.g. Hughes et al., 1977; Grosswald, 1980; Grosswald and Hughes, 1999), the conditions that permit ice grounding of the Lomonosov Ridge have not been quantified. In this paper we model the growth of an ice shelf within the Arctic Ocean, emanating from the Eurasian ice sheet during a phase of large-scale glaciation. In doing so we
provide an assessment of the gross conditions needed to ground ice in the central Arctic Ocean by an ice shelf fed from the Barents-Kara Seas. This experiment constitutes a first step in a more comprehensive evaluation of the former ice-sheet/ice-shelf system in the Arctic Ocean. We take a simple approach where the complex ice shelf dynamics and grounding line mechanics not are included. Such an approach allows us to undertake a sensitivity analysis of the problem, in a way not possible with a complex coupled model. In this way we are able to explore a full range of possible situations, the result of which will guide future modeling investigations over the next few years.

**ICE SHEET MODEL**

Numerical ice sheet modeling is a well-established tool for constraining the large-scale flow and form of ice sheets with well-understood empirical laws. We use a numerical model that has been used previously to model the Eurasian ice sheet during the last glaciation (Siegert and Dowdeswell, 1995; Siegert et al., 1999). Brief details of the model are qualitatively outlined below and we refer to Siegert and Dowdeswell (1995) and Siegert et al. (1999) for a more detailed description of the numerical model.

The growth of glacier ice is modeled over a regular grid of bathymetry and topography. This base data set was prepared from the International Bathymetric Chart of the Arctic Ocean (IBCAO) (Jakobsson et al., 2000, version 1.0) and the Global Seafloor Topography from satellite altimetry and ship soundings (Smith and Sandwell, 1997, version 8.2). The latter data set was used in the areas included in the model below 64°N, which is south of the IBCAO coverage. Both these data sets have derived the land topography of the Eurasian continent from GTOP030 (U.S. Geological Survey, 1997). A grid with $20 \times 20$ km cell spacing was prepared on a Lambert equal area projection (Projection center 90°N, central meridian 44°E) by sub-sampling the original re-projected data points from the input databases using a nearest neighbor algorithm.

The numerical ice-sheet model is based upon the continuity equation for ice (Mahaffy, 1976), where the time-dependent change in the ice thickness of a model grid cell is associated with the specific net ice mass budget of that particular cell. The net flux of ice from a given model grid cell is function of ice thickness and ice velocity in that cell and its neighbouring cells. The ice mass balance is controlling the ice thickness. Mass balance of an ice sheet is defined as the difference between the mass gained by snow accumulation and that lost by ice ablation and calving of icebergs. In order to make an ice shelf grow out into the Arctic Ocean we removed the model iceberg calving function and instead allowed the ice to flow out off the shelf edge and into the ocean.

The ice shelf model is simple. The velocity of floating ice is determined by:

$$u_f = u_{f0} + \Delta x \cdot \varepsilon_s,$$

where $u_f$ is the velocity of floating ice, $u_{f0}$ is the velocity of ice coming into the cell of floating ice, $\Delta x$ is the grid cell spacing (20 km) and $\varepsilon_s$ is the strain rate of the ice shelf, which is varied between 0.005 and 0.000 year$^{-1}$ (Table 1). The strain rate is the rate at which deformation of the ice shelf occurs. The ice shelf model can be understood by considering that a freely flowing ice shelf is likely to have a strain rate around 0.005 year$^{-1}$ (Payne et al., 1989). Thus, we model both a freely flowing ice shelf, and one in which no acceleration can take place (when the strain rate is zero). The rationale for the latter experiment is that

**Table 1. Numerical ice sheet modeling strategy and boundary conditions.**

<table>
<thead>
<tr>
<th>Model experiment</th>
<th>Rate of marine melting (cm/year)</th>
<th>Ice surface accumulation over shelf</th>
<th>Strain Rate (year$^{-1}$)</th>
<th>Grounded ice in the Arctic Ocean?</th>
</tr>
</thead>
<tbody>
<tr>
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<td>N</td>
<td>0.005</td>
<td>N</td>
</tr>
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<td>30</td>
<td>N</td>
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<td>N</td>
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<td>N</td>
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<td>N</td>
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<tr>
<td>4</td>
<td>50</td>
<td>Y</td>
<td>0.005</td>
<td>N</td>
</tr>
<tr>
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<td>80</td>
<td>Y</td>
<td>0.005</td>
<td>N</td>
</tr>
<tr>
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<td>50</td>
<td>N</td>
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<td>Yermak (10 kyr), Gakkel (17.4 kyr)</td>
</tr>
<tr>
<td>7</td>
<td>50</td>
<td>Y</td>
<td>0.0</td>
<td>Yermak (6 kyr), Gakkel (11 kyr), Morris Jessup Rise (12.2 kyr), Lomonosov W (12.8 kyr), Lomonosov E (15.2 kyr)</td>
</tr>
<tr>
<td>8</td>
<td>80</td>
<td>Y</td>
<td>0.0</td>
<td>Yermak (10 kyr), Gakkel (18.8 kyr)</td>
</tr>
<tr>
<td>9</td>
<td>100</td>
<td>Y</td>
<td>0.0</td>
<td>Yermak Plateau only (14.2 kyr)</td>
</tr>
</tbody>
</table>
increased flow rates are curtailed by a counter force, which could come from thick sea-ice acting against the terminus of the ice shelf.

It should be noted that the model contains no grounding line physics. It simply switches between grounded ice and ice shelf flow between cells of grounded and floating ice. Although this model is simple it is adequate for the purpose of this investigation, to assess the gross-conditions required for a thick ice shelf to form within the Arctic Ocean.

Iceberg calving is modeled using a depth-related function, which can describe the rate of icebergs calved from grounded margins (Pelto and Warren, 1991). This treatment is applied only to the western margin of the ice sheet. In all experiments iceberg calving was not allowed within the Arctic Ocean.

The included ice shelf model has a very simple marine melting function, which is an over simplification of the ocean/ice conditions expected in a complex oceanographic environment such as the Arctic Ocean. However, full quantification of the problem involves the use of coupled ocean/ice sheet models and this is far beyond the scope of our first test, but will be in further experiments, as mentioned above. The degree of ice shelf basal melting is difficult to assess for glacial periods in the Arctic Ocean due to the poorly known oceanographic conditions. In this paper we force the basal melting of the ice shelf floating into the Arctic Ocean at a constant value. This value is varied within our sensitivity tests (Table 1). The surface accumulation of ice was included following the ice-sheet/AGCM model inter-comparison work of Siegert and Marsiat (2001). This accumulation distribution is that required to form a Late Weichselian style ice sheet across the Eurasian Arctic.

We experiment with two modes of ice shelf surface accumulation: one in which no accumulation takes place (so that the effects of ice flow and basal melting can be assessed) and one in which accumulation is retained in the model, following Siegert and Marsiat (2001).

**EXPERIMENTAL DESIGN**

In order to establish the likelihood of an Arctic Ocean ice shelf supplied exclusively from the Eurasian ice sheet, numerical model experiments were undertaken with a variety of oceanographic and climate boundary conditions. The model was run for a maximum of 20,000 years (starting at zero with no ice) assuming constant environmental forcing. Table 1 shows the parameters used for each of the model runs discussed in Section 6.

**RESULTS**

Of the many experiments undertaken we have chosen to discuss nine (Table 1). These examples display the full range of results that were produced in our investigation. For each of the nine experiments the ice sheet volume through time is provided in Figure 3.

*Experiment #1. 10 cm/year melting, zero ice shelf accumulation.* (Figs. 4 a, b, c, d): 1800, 2800, 3800 and 4200 years

The first thing to note about this experiment is that the velocity field starts to become ‘unstable’ (that is, the Arctic Ocean basin fills with ice, which results in the model miscalculating the ice shelf velocity) after about 4200 years. Prior to this, however, the model produced some interesting results (Fig. 4). An ice shelf readily formed within the Arctic Ocean as ice growth took place in the Barents Sea. Major outlet drainage pathways within the St. Anna and Franz Victoria Troughs were the main suppliers of ice to this ice shelf.
Figure 4a. 1800 year time snapshot from experiment #1, 10 cm/year marine melting, zero ice shelf accumulation. The modeled ice is shown with a color table representing ice velocity in meters/year. The white contours are ice thickness in meters. As in figure 2, a white semitransparent plane has been inserted at 1000 m below present sea level to emphasize the areas of the Lomonosov Ridge that are shallower than 1000 m. LR=Lomonosov Ridge, GK=Gakkel Ridge.

Figure 4b. 2800 year time snapshot from experiment #1, 10 cm/year marine melting, zero ice shelf accumulation.
Between 3800 and 4200 years the ice shelf covers the region of the Lomonosov Ridge where ice was once grounded, but the thickness at this time was very small (Figs. 4d and e). At 4200 years, the ice shelf in front of the Franz Victoria Trough over the Lomonosov Ridge was less than 15 m thick. This experiment showed that it was possible to grow a thick ice shelf rapidly within the Arctic Ocean provided that the marine melting rate was low. However, the thickness required for grounding was never reached. The general form of the ice shelf at 4200 years is as one might expect from an Antarctic point of view; the grounding zone is thick (around 700 m) and the ice shelf thins to around 100 m some 500 km from the grounding zone. At no time did ice ground across any part of the Lomonosov Ridge. The results from this experiment were similar to those when marine melting is increased to 20 cm/year.

Experiment #2, 30 cm/year marine melting, zero ice shelf accumulation (Fig. 5): 9000 years

By increasing the rate of marine melting to 30 cm/year the ice shelf within the Arctic Ocean fails to fill the whole basin. Instead, lobes of floating ice are established in front of the St. Anna and Franz Victoria ice streams. The St. Anna floating ice shelf covers the region where ice was grounded, but the thickness of the ice shelf in this position is never greater than 80 m. The velocity field in this experiment becomes unstable at about 10,000 years.

Experiment #3, 50 cm/year marine melting, zero ice shelf accumulation (Fig. 6 a, b): 6000 & 9000 years

By increasing the rate of marine melting to 50 cm/year, the model gave similar results to experiment 2. The ice shelf grew from the St. Anna Trough across the region of grounded ice, but the ice shelf thickness in this run was never greater than 50 m.

Experiment #4, 50 cm/year marine melting, ice accumulation included (Fig. 7): 6000 years

When ice accumulation was included in the model with 50 cm/year of marine melting, the ice shelf became unstable after 6000 years. This is because the accumulation (which is greatest toward the west) offset the amount of ice loss due to melting and encouraged ice shelf growth to occur across the entire Arctic Ocean. This experiment is a benchmark for our tests, because it suggests that marine-melting conditions where ice loss rates are less than 50 cm/year would cause complete ice shelf coverage of the Arctic Ocean under the precipitation conditions that occur across the Barents Sea. Notably, prior to the model becoming unstable at 6000 years, the ice shelf in front of the Franz Victoria Trough had a thickness of more than 100 m over the Lomonosov Ridge close to the continental margin of northern Greenland. The thickness fails to get much greater than this value. Only when the flow field is unstable does the ice shelf...
thicken to more than 500 m over the Lomonosov Ridge.

The ice shelves immediately in front of the Franz Victoria and St. Anna Troughs have a thickness of over 700 m. In this experiment, as in others, the thickness reduces to no more than 80 m by the time the ice gets to the Lomonosov Ridge.

Experiment #5, 80 cm/year marine melting, ice accumulation included (Fig. 8): 10,000 years

In this experiment small stable ice shelves existed across the mouths of the St. Anna and Franz Victoria Troughs. However, these ice shelves failed to cover the Lomonosov Ridge. Under these environmental conditions it is unlikely that a substantial ice shelf could have existed within the Arctic Ocean.

Experiment #6, 50 cm/year marine melting, zero ice shelf accumulation, strain rate=0 (Fig. 9 a, b): 10,000 & 20,000 years

In experiments 1-5 grounding of ice across the Lomonosov Ridge was never achieved. Despite low basal melt rates, the ice shelf spread and thinned from the Barents/Kara margin such that it was less than 100 m thick across the ridge. One way to stop the ice shelf thinning is to reduce the strain rate. As there appears to be no way to ground an ice shelf with a strain rate of 0.005 year⁻¹, the strain rate was reduced in a further set of tests.

By reducing the strain rate of ice to zero, the velocity of the ice shelf was modified significantly. Whereas in experiment #3 ice less than 50 m thick flowed out across the Lomonosov Ridge from the Kara Sea, now the ice is thick enough to ground across the continental margin.

In the model, ice that is grounded well below sea level may be treated as if it were in an ice stream (as in the Bjørnøya Trough ice stream, north of Norway). In this experiment, such anomalous grounded ice velocities can be seen across the northern continental slopes (e.g. the Yermak Plateau) (Fig. 9a, b). This rather unusual velocity profile is unlikely to be realistic (longitudinal stresses, neglected in the model, would in reality act to smooth the velocity in these regions). However, as the grounded velocities in such places are surrounded by slow-flowing floating ice, the ice sheet profile is not adversely affected (in other words the model is not unstable). Thus, we are able to detect the places where grounding occurs by observing those locations where unusually high ice velocities are calculated.

In this experiment, ice was grounded across the Arctic Ocean in two places. First, after 10,000 model years, the Yermak Plateau was covered by grounded ice. Second, after 17,400 years, grounded ice covered the shallowest region of the Gakkel Ridge where a pinnacle in the ridge known as the Langseth Ridge reaches above 1000 m water depth according to the

Figure 5. 9000 year time snapshot from experiment #2, 30 cm/year marine melting, zero ice shelf accumulation. The modeled ice is shown with a color table representing ice velocity in meters/year. The white contours are ice thickness in meters. As in figure 2, a white semitransparent plane has been inserted at 1000 m below present sea level to emphasize the areas of the Lomonosov Ridge that are shallower than 1000 m. LR=Lomonosov Ridge, GK=Gakkel Ridge.
Figure 6a. 6000 year time snapshot from experiment #3, 50 cm/year marine melting, zero ice shelf accumulation. The modeled ice is shown with a color table representing ice velocity in meters/year. The white contours are ice thickness in meters. As in figure 2, a white semitransparent plane has been inserted at 1000 m below present sea level to emphasize the areas of the Lomonosov Ridge that are shallower than 1000 m. LR=Lomonosov Ridge, GK=Gakkel Ridge.

Figure 6b. 9000 year time snapshot from experiment #3, 50 cm/year marine melting, zero ice shelf accumulation.
Figure 7. 6000 year time snapshot from experiment #4, 50 cm/year marine melting, ice accumulation included. The modeled ice is shown with a color table representing ice velocity in meters/year. The white contours are ice thickness in meters. As in figure 2, a white semitransparent plane has been inserted at 1000 m below present sea level to emphasize the areas of the Lomonosov Ridge that are shallower than 1000 m. LR=Lomonosov Ridge, GK=Gakkel Ridge.

Figure 8. 10,000 year time snapshot from experiment #5, 80 cm/year marine melting, ice accumulation included. The modeled ice is shown with a color table representing ice velocity in meters/year. The white contours are ice thickness in meters. As in figure 2, a white semitransparent plane has been inserted at 1000 m below present sea level to emphasize the areas of the Lomonosov Ridge that are shallower than 1000 m. LR=Lomonosov Ridge, GK=Gakkel Ridge.
Figure 9a. 10,000 year time snapshot from experiment #6, 50 cm/year marine melting, zero ice accumulation, strain rate=0. The modeled ice is shown with a color table representing ice velocity in meters/year. The white contours are ice thickness in meters. As in figure 2, a white semitransparent plane has been inserted at 1000 m below present sea level to emphasize the areas of the Lomonosov Ridge that are shallower than 1000 m. LR=Lomonosov Ridge, GK=Gakkel Ridge.

Figure 9b. 20,000 year time snapshot from experiment #6, 50 cm/year marine melting, zero ice accumulation, strain rate=0.
Figure 10a. 15,200 year time snapshot from experiment #7, 50 cm/year marine melting, ice accumulation included, strain rate=0. The modeled ice is shown with a color table representing ice velocity in meters/year. The white contours are ice thickness in meters. As in figure 2, a white semitransparent plane has been inserted at 1000 m below present sea level to emphasize the areas of the Lomonosov Ridge that are shallower than 1000 m. LR=Lomonosov Ridge, GK=Gakkel Ridge.

Figure 10b. 15,600 year time snapshot from experiment #7, 50 cm/year marine melting, ice accumulation included, strain rate=0.
IBCAO map. No other regions of the Arctic Ocean floor were affected by grounded ice.

**Experiment #7, 50 cm/year marine melting, ice accumulation included, strain rate=0** (Fig. 10 a, b): 15,200 & 15,600 years

When ice accumulation is accounted for, the ice sheet grounds across much of the Arctic Ocean. Such a situation is highly unrealistic, but it makes the point that significant ice thickness can be produced within the Arctic Ocean by a stagnant ice shelf whose surface balance exceeds its basal melting.

After 6000 years of model time, the Yermak Plateau was covered with grounded ice; 4000 years earlier than in experiment #6. Later, at 12,200 model years, the Gakkel Ridge experienced grounded ice. By 15,200 years, the Lomonosov Ridge was covered by grounded ice in the region where there is evidence for ice grounding.

**Experiment #8, 80 cm/year marine melting, ice accumulation included, strain rate=0**

By increasing the rate of subglacial melting to 80 cm/year, the ice shelf extent was reduced significantly from that in experiment #7 (to something closer to experiment #6). In this case only the Yermak Plateau and the Gakkel Ridge, again at the Langseth Ridge area, were affected by grounded ice.

**Experiment #9, 100 cm/year marine melting, ice accumulation included, strain rate=0**

Increasing the rate of melting to 100 cm produced an ice shelf that was restricted to the Barents-Kara margin. Only the Yermak Plateau was covered by grounded ice.

**DISCUSSION**

The results of our model are not conclusive evidence for former ice shelf grounding. We do not model ice shelf physics very well, nor do we include grounding line mechanics. Further, we treat the process of marine melting within the Arctic Ocean by applying the same rates of melting at all places. This oversimplification is required in the absence of highly complex coupled ice-sheet/shelf/sea modeling. Our study focuses solely on defining the gross mass balance requirements of an ice shelf within the ocean fed by ice from the Barents and Kara Seas. The results of this experiment must be viewed with the above limitations in mind.

Modeling experiments 1-5 fail to produce floating shelf ice lobes emanating from the Barents and Kara Seas that are thick enough to ground on the Lomonosov Ridge. The modeled ice sheet behaves like what is seen in Antarctica today where the shelf ice thins fairly rapidly from beyond the grounding line where it starts to float. Measurements of ice thickness along calving lines on the Antarctic shelf generally yield a thickness on the order of 200-300 m, although a thickness of more than 500 m has been observed at the margin of the Filchner Ice Shelf (Drewry, 1983) and recent plowmarks are reported from Antarctic waters from depths as deep as 600 m (Orheim and Elverhøi, 1981). Tabular icebergs may increase their draft by 50% if capsized (Lewis and Bennett, 1984) and this may explain some of the deepest plowmarks found on the Antarctic continental shelf as well as in the Arctic Ocean. Capsized deep icebergs are, however, less likely to have caused the spatially extensive and directionally consistent flutes and large size scours mapped over large areas of the Lomonosov Ridge and the Chukchi Borderland (Polyak et al., 2001). In experiments with ice shelf basal melting included, the model produces a shelf which resembles a series of ‘lobes’ originating from the fast flowing grounded outlets. In reality the growth of such lobes may be restricted by lateral flow of ice, not accounted for well in our ice shelf model. Without this lateral spreading, however, we give the ice shelf the maximum possible chance of extending out to the Arctic Ocean.

We reach a thickness of about 700 m immediately in front of the Franz Victoria and St. Anna Troughs in experiment #4 when ice accumulation was included in the model with 50 cm of marine basal melting. If we assume a sea level as much as 150 m below present during a large glaciation this scenario has the potential of producing very deep icebergs, although a keel depth of about 850 m is required to ground on the deepest areas of the Lomonosov Ridge crest where glacigenic bedforms have been mapped (Jakobsson, 1999; Polyak et al., 2001). By adopting the idea that capsized tabular icebergs could have produced some of the deepest scours mapped in the Arctic Ocean the pre capsized tabular icebergs would have to have drafts less than 570 m in order to reach 850 m while capsized. Poljak et al. (1997) reported that ice filled the St Anna Trough down to 630 m below present sea level during the last glaciation and this may be a potential source for some of the deep ice scours found, although not the deepest on the Lomonosov Ridge, which in addition to being deeper predates the last glaciation (Jakobsson et al., 2001).

The key parameter that prevents the ice lobes from the Franz Victoria and St. Anna Troughs from grounding on the Lomonosov Ridge is the ice shelf
strain rate. In experiments # 1-5, the strain rate was set at 0.005 year\(^{-1}\), which is a generalization for free-flowing ice in water (Payne et al., 1989). Therefore, these experiments are intended to describe how a free-flowing ice shelf would behave within the Arctic Ocean. The strain rate, multiplied by the distance over which the strain operates, amounts to an increase in ice velocity as the ice flows under its own weight. This increase in velocity causes the ice shelf to thin with distance from the grounding zone. By lowering the strain rate to zero a completely different scenario emerges. In experiment #7, which accounts for accumulation and a basal melting of 50 cm/year, the ice sheet grounds after 15,200 model years on the western Lomonosov Ridge and after 15,600 years in the areas of the ridge where grounding has been mapped (Figs. 10a, b). This leads to the question regarding the type of conditions that could lead to an extraordinary low ice strain rate. We can speculate that such conditions could have prevailed during periods of large-scale glaciation with a substantially thickened and stagnant sea ice present in the ocean. This could have a buttressing effect on the ice flowing from the Barents Sea and, thus, preventing the ice from rapidly thinning while flowing out into the Arctic Ocean.

Today the deep Arctic Ocean sea ice cover is generally not thicker than 5 m (for discussion about sea ice thickness distribution in the Arctic Ocean see Wadhams, 1997). While discussing the hypothesis of a floating ice shelf in Arctic Ocean Broecker (1975) pointed out that sea ice reaches its equilibrium thickness when winter growth just matches summer ablation and, therefore, different climatic conditions would produce different steady-state thicknesses. It was proposed by Crary (1960), and later discussed by Mercer (1970), that if the heat from the inflowing warm Atlantic water were reduced to below one third of its present value almost unlimited thickening of the sea ice would occur. Mercer (1970) suggested that this could be an important mechanism for growing an ice shelf during glacial conditions. Thus, the existence of such an ice shelf is compatible with the model’s requirement for grounding that the sub ice shelf melt rates are less than 50 cm/year.

The experiments testify to the importance of the ice shelf’s net mass balance of the location of grounded ice in the Arctic Ocean (under zero ice shelf strain rates). In the model, grounding is only possible over the Lomonosov Ridge when surface ice accumulation is included and basal melt rates are low (50 cm/year). As the melt rates are increased, so the maximum distance from grounded ice in the Arctic Ocean to the Barents margin is reduced. Thus, when the melt rate is 100 cm/year, grounding is only possible over the Yermak Plateau (experiment #9). The model treats basal melting in the oceans as being the same rate in all places. It would probably be more realistic to assume that rates of basal melting would be spatially variable. For instance, one could argue that melt rates closer to the Fram Strait would be greater than those further into the Arctic Ocean (due to influx of warm Atlantic Water). Such a situation would not affect the grounding of ice across the Yermak Plateau, as it is covered by grounded ice in all the experiments #6-9. However, the large-scale grounding of the Gakkel Ridge may be curtailed under a geographically-modified melt regime. Furthermore, as the area of the Lomonosov Ridge that is known to have been affected by ice grounding is far from the Fram Strait, we would expect lower rates of melting here compared with regions farther to the west.

**CONCLUSIONS**

A numerical ice sheet model was used to test the hypothesis that an ice shelf once grounded across the Lomonosov Ridge to a depth of around 1000 m below present sea level. We undertook nine tests with a variety of mass balance (basal melting and surface accumulation) conditions and ice shelf strain rates. In all experiments iceberg calving is not allowed (as this process would dismantle the ice shelf). The following conclusions could be drawn from the model experiments:

- When the ice shelf strain rate was 0.005 year\(^{-1}\), under relatively low rates of basal melting (less than 80 cm/year), an ice shelf flowed from the Barents/Kara Seas across the Lomonosov Ridge. However, the ice shelf was never thicker than 100 m across the ridge. Reducing the rate of subglacial melting to zero caused the ice shelf to thicken, but only by a few tens of meters. A free-flowing ice shelf from the Barents Sea is therefore unlikely to have been responsible for the grounding observed across the ridge.

- When the ice shelf strain rate was reduced to zero floating ice velocities could not increase from the grounding zone. This allowed the ice shelf to become much thicker than in previous experiments and, in some experiments where the subglacial melting was less than 80 cm/year, grounding of ice across the Lomonosov Ridge was modeled.

- An ice shelf flowing with a low strain rate could not be considered to be ‘free flowing’. Instead, the ice shelf must be supported or ‘buttressed’ by a counter force. For the case of the Arctic Ocean, such buttressing could come
from the action of ultra thick sea ice held fast within the Arctic Ocean basin.

- Thus, grounding of ice across the Lomonosov Ridge, emanating from the Barents Sea, is possible but only if the ice shelf velocities and basal melt rates are kept low.

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SEDIMENTARY THICKNESS ESTIMATIONS FROM MAGNETIC DATA IN THE NANSEN BASIN

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ABSTRACT

In spite of several geophysical expeditions during the past decade, the seismic documentation of sediments in the Nansen Basin is still sparse. At the same time, airborne surveys conducted by the US Naval Research Laboratory in 1974-75 and 1998-99 make up a grid of magnetic profiles spaced by 18 to 10 km in the western half of the basin. It is known that the depth to oceanic basement may be calculated from magnetic data by a multiple-source Werner Deconvolution method (MSW). Recent bathymetry compilations provide relatively good control of the seafloor. Sediment thickness represents the difference between depths of basement and seafloor.

Results of analysis regarding the possibility to use MSW calculations as a support to seismic lines in the mapping of sediment thickness in the study area are presented. The study area, at 80–86°N, 11–35°E, includes the Nansen Basin, Gakkel Ridge flank, Yermak Plateau and north Svalbard margin.

Principles of analysis and sorting of MSW solutions are discussed in detail. Uncertainty of the method is estimated. It is established that in general, both the magnetic basement surface and the sediment thickness values estimated from magnetic data fit well with seismic data from icebreakers Oden and Polarstern (2001). Possible reasons for differences between magnetic and acoustic basement are discussed. The regional map of the sediment thickness in the study area shows that in general the thickness of sediments decreases from the Barents Sea shelf toward the Gakkel Ridge. Least thickness or absence of sediments is observed in the axial zone of the Ridge. The largest accumulation of sediments (4-5 km) is discovered on the continental shelf and slope, north and south of Yermak Plateau, and in the southern part of Litke Trough. The thickness of sedimentary cover on Yermak Plateau varies from 1 to 2 km.

INTRODUCTION

Information on the sediment thickness is important for both the regional study/modeling of geological processes related to development of sedimentary basins and the assessment of oil and gas reserves. Moreover, sediment thickness of the ocean floor has become a major concern to many coastal states in relation to the revision of outer limits of the continental shelf in accordance with Article 76 of the UN Convention on the Law of the Sea.

Existent maps of sediment thickness in the Arctic Ocean (e.g. Jackson and Oakey, 1988; Gramberg and Puscharovsky, 1989; Gramberg et al., 1999) are mainly based on widely spaced and irregular seismic lines. Thus wide areas of the maps are based on interpolation of seismic data. The most effective method of sediment thickness mapping is obviously based on the integrated analysis of seismic, gravity, magnetic and bathymetry information.

For both scientific and Article 76 purposes, Norwegian and Russian institutions (Norwegian Petroleum Directorate; Department of Geosciences, University of Oslo; VNIIOkeangeologia and Polar Marine Geological Research Expedition) entered into a collaborative project to study the feasibility of using potential field data integrated with seismic and bathymetry data to determine the ocean floor and sediment thickness in the Arctic Ocean. The area 83°50’-86°N, 13-34°E in the western part of the Nansen Basin (small box, Fig.1-B) was chosen as the test area. This part of the basin has the best coverage of seismic data in addition to high-quality digital aeromagnetic and aerogravity information. Later the area of investigation was extended to the south (large box, Fig.1-B). This paper summarizes results of the mapping of oceanic basement relief and estimated sedimentary thickness from aeromagnetic data integrated with bathymetry.

Mass depth to magnetic source estimations is the common technology to study the geomagnetic structure of oceanic crust. Euler Deconvolution (Nabighian, 1974) and Werner Deconvolution (Werner, 1953) are the most widely used methods of estimation. The first method is usually considered as a fast one for rapid estimation of depths to magnetic sources (upper edges) from gridded magnetic data. The second method is more laborious and needs magnetic anomaly profile data. At the same time it allows estimation of parameters of magnetic sources more precisely. That is why it was used in this project. Seismic data were used
both to calibrate and check results of independent estimations from magnetic anomalies.

**DATA PROCESSING AND INTERPRETATION**

Several geophysical data sets from the Nansen Basin were used under the project (Fig. 1-A): seismic data collected in 2001 from the icebreakers *Oden* and *Polarstern* (Gjengedal, 2004; Jokat and Micksch, 2004), free air gravity and magnetic profiles collected by the Naval Research Laboratory (NRL). Bathymetry information was presented by the latest version of the IBCAO grid (Jakobsson et al., 2000), by paper copy of the Russian map “Bottom Relief of the Arctic Ocean” at scale 1:5,000,000 (1999), and by bathymetry

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**Figure 1.** Data coverage and area of study. A) shows position of seismic lines from *Oden*-2001 (1) and line 20010100 from *Polarstern*-2001 (2), aeromagnetic tracklines of NRL 1974-75 (3) and NRL 1998-99 (4). B) shows positions of the initial test area (small box) and the extended study area.
observations along profiles acquired in 2001 from the icebreaker *Oden* and the research vessel *Polarstern*.

The preliminary processing and analysis of the data sets included:

- Seismic interpretation and construction of depth sections.
- Levelling, adjustment and compilation of NRL 1998-99 aerogravity data sets (Brozena et al., 2003), satellite data (Laxon and McAdoo, 1998) and Norwegian marine gravity observations to construct an updated grid and map of gravity anomalies.
- Digitizing and gridding of isobaths from the Russian bathymetry map, comparison of the resulting grid and the IBCAO grids with shiptrack bathymetry (Thiede, 2002) and Russian profiles to upgrade existent information and/or to select the most reliable grid values among multiple sources.
- Levelling and adjustment of aeromagnetic data sets. Adjustment included navigational correction (shift) of old (NRL 1974-75) magnetic anomaly profiles with poor navigation using recent data sets (NRL 1998-99) with excellent navigation as reference information. The average magnitude of shift was about 1-2 km and the maximum value ran up to 5 km. The final coherent magnetic database was used both for mapping and for mass depth to magnetic source estimations.

Depths to oceanic basement were calculated by a multiple-source version of the PDEPTH program (Cordell et al., 1992) based on the Werner Deconvolution method (Hansen and Simmonds, 1993). It estimates depths to magnetic sources approximated by simple models: magnetic contact between two differently magnetized bodies or a magnetized thin dike. The contact model was chosen as the basic one after test calculations. Initially the magnetic anomaly profiles in the Nansen Basin were filtered by a six-point running average filter with 3 km window length to remove high frequency noise. Input parameters for the calculations, including number of points in the running window, radius of cluster (a group of solutions around the “real” solution, generated by the program), number of solutions in cluster etc., were defined on the basis of test estimations within the swath of the geotransect “De Long Islands-North Pole” (Lickhachev, 1999). This swath is covered by 5 km spaced aeromagnetic profiles. The basement relief here is known from seismic data (Sorokin et al., 1999). Parameters that gave the best fit between the real and estimated depths were chosen as optimal.

Mass depth to magnetic sources estimations were carried out and these formed the initial database for further analysis. Next, detailed analysis and sorting of all estimations were performed to define those estimations belonging to magnetic basement. The editing procedure included several steps:

- Single solutions were regarded least reliable and were removed from the initial database. Solutions in clusters were averaged (Fig. 2). Only averaged solutions (AS) were accepted for further analysis and

![Figure 2. Analysis and sorting of depth estimations along aeromagnetic trackline of NRL 1999. Magnetic anomalies and their numbers (1) are shown at the top. All computed (2) and averaged (3) estimations in clusters (4) are shown in the middle section. In the lower section only averaged solutions (3) are shown, those that have been removed from the database (too shallow and too deep) (5), and seafloor relief (6) extracted from the IBCAO grid.](image-url)
Figure 3. Points of final depth estimates from magnetic data presumably connected with basement and depths to basement in km below sea level.

- The most shallow AS that fell above the ocean floor known from bathymetry were removed from the database and recorded in a separate file. Analysis of these solutions shows that a part of them is connected with high frequency components of the magnetic field (noise), but another part (especially in the axial zone of the Gakkel Ridge) may be connected with real highs of the sea bottom relief smoothed in the bathymetry grid.
- AS from magnetic anomalies with unusual shapes (i.e. elongated along profile or known to differ significantly from the reference anomaly shape assumed in the Werner Deconvolution method) were removed from database.
- AS related to magnetic profiles that are not orthogonal to strike of magnetic anomalies (within limits 0-30°) were corrected (reduced by \( \cos \lambda \), where \( \lambda \) is the angle between directions of profile and the perpendicular to the magnetic anomaly axis). Solutions were removed from the database if \( \lambda \) exceeded 30°. The cutoff limit of \( \lambda \) was determined from model calculations.
- All remaining AS are presumably connected with oceanic basement were plotted and compared with seismic information in the test area (Fig. 1-B).

It is necessary to note that seismic information was used not only for comparison with the results of depth to magnetic basement estimations, but also for the calibration of results. Thus the process of mass calculations and editing was iterative. Accordingly, the input parameters for depth calculations chosen at first on the basis of Russian data (Sorokin et al., 1999) were gradually adjusted to the best fit of magnetic estimations with acoustic basement in the test area (Fig. 1-B). After the completion of this process the majority of these estimations agreed well with the seismic data collected from the icebreaker Oden (2001). The average error was within a few hundred meters. However, some estimations conform to magnetic sources at depths noticeably below acoustic basement. It was supposed that such deep sources could be situated within or close to faults. As a result it became necessary to map fracture zones and other structures in the study area.

The structural map was developed mainly on the basis of potential field and bathymetry maps constructed under the project. Seismic data and recent publications (Thiede et al., 2002; Brozena et al., 2003; Engen et al., 2003; Glebovsky et al., 2003) were used as additional information. Main fracture zones were interpreted from displacements of magnetic lineations, and from characteristic features of potential fields and their horizontal gradients compared with bathymetry data.

Deep AS corresponding to linear lows in the bottom relief and located near faults were removed from the database. Further analysis made it clear that these solutions may be identified also by statistical analysis along each profile. The deep AS is generally more than two standard deviations away from the average AS of the same magnetic profile.

Since the results of estimations in the test area were encouraging, the mass depth to magnetic sources
estimations and further sorting were carried out in the wider region at 80–86°N, 11–35°E (Fig. 1). All finally selected AS presumably related to the top of basement (see Fig. 3) were transformed to grid with cell size 10×10 km using the Kriging method (Cressie, 1990). This method allows grids to be calculated from widely spaced and irregularly distributed point observations. Later, this grid was used to estimate both the sediment distribution in the study area and the uncertainty of the MSW method.

Analysis of the bathymetry information shows that there are essential differences between different data sets especially in the axial zone of the Gakkel Ridge. At the same time it revealed that the IBCAO grid seems to be the most reliable bathymetry model available at present. The final sediment thickness grid and map (Fig. 4) were constructed by subtracting the IBCAO grid (Jakobsson et al., 2000) from the above-mentioned grid of estimated depth to basement.

UNCERTAINTY OF METHOD

The uncertainty of the method was estimated by comparison of results of depth to magnetic basement calculations with depth sections based on seismic data collected in 2001 from the icebreaker Oden and the research vessel Polarstern (Jokat and Micksch, 2004). While data from Oden was used to calibrate depth estimations, data from Polarstern was only used to estimate uncertainty of the method. Results of the comparison are presented in Figure 5. Both final point depths to magnetic basement projected on seismic profiles within swaths of different width (0-3, 3-6 and 6-10 km) and the difference between seismic basement and magnetic basement extracted from final grid are shown.

The correlation factor (R) and the standard deviation (σ) between seismic and magnetic basement depths were calculated for statistical estimations of uncertainty of the method (Fig.6). Both point estimations (A) and those extracted from the depth to magnetic basement grid within a 10 km swath (B) were analysed. An integrated analysis of the results presented in Figure 5 and Figure 6 allows for several conclusions:

- A high coefficient of correlation between magnetic and acoustic basement depths as well as a small standard deviation testify to relatively high accuracy of estimations from magnetic data. The average error is less than 20% of depth to basement and usually about 10%.
Errors of point estimations are presumably related to two main effects: non-uniqueness inherent in the method of depth to magnetic sources estimation, and natural differences between magnetic and acoustic basements.

In spite of the relatively high resolution of the method, the final grids constructed under the project determine only regional fluctuations of the basement surface and the distribution of sediments. This may be explained by the fact that the density of magnetic profiles in the study area is too low to map basement elevations and lows visible on seismic profiles. Aeromagnetic trackline spacing is about 10-18 km. The majority of basement lows and highs are not more than 10-15 km wide.

A total amount of 732 final point estimations remained in the data base after analysis and sorting within the 218,810 km² study area. Only 66 of these occurred within 10 km wide swaths around seismic profiles and were used to estimate the uncertainty of the method.

Presumably, the effects of non-uniqueness of the method might be reduced if the spacing between magnetic profiles is reduced to about 5 km.

SUMMARY AND CONCLUSIONS

1. Depth to magnetic source calculations by the MSW Deconvolution method may support seismic investigations for the estimation of the regional sediment distribution in the Arctic Ocean. The crucial prerequisites to get positive results are the following:
   - High quality of observations and correct direction of magnetic anomaly profiles. Any artefacts of magnetic anomaly shape result in incorrect fitting of input parameters for MSW calculations and finally in depth estimation errors. Direction of profiles must not differ noticeably from the orthogonal to the strike of magnetic anomalies in order to avoid sizeable corrections.
   - The tectonic and structural maps constructed on the basis of all available potential field and bathymetry data have to be considered as additional information in...
the formalized procedure of depth to basement estimation sorting/selection.

2. It is established that, in spite of relatively high resolution of the MSW method, the final maps of the basement surface and sediment distribution maps represent only regional variations in these two features.

3. In spite of its regional character the final sediment thickness map (Fig. 4) looks more detailed than earlier published maps (Jackson and Oakey, 1988; Gramberg and Puscharovsky, 1989; Gramberg et al., 1999) and shows the principal features of sediments distribution in the study area. In accordance with the physiographic classification of Jakobsson et al., 2003, this area includes a few first-order provinces: Gakkel Ridge; Barents Abyssal Plain, Continental Rise and Slope, and Continental Shelf; Yermak Plateau and Continental Slope.

4. In general the thickness of sediments within the whole study area decreases from the Barents Continental Shelf toward the Gakkel Ridge. The least thickness of sediments (less than 0.5 km) is observed in the axial zone of the Gakkel Ridge. The rift valley has practically no sediments. Large accumulations of sediments (3-4 km) are identified within both the Barents Continental Slope and the Yermak Continental Slope north and south of the Plateau. Even larger accumulations of sediments (4-5 km) are found within two small areas confined to the Barents Shelf and Continental Rise and located in the very southwest part of the study area in the Litke Through (Bottom relief of the Arctic Ocean (map), 1999). The thickness of sedimentary cover both within central and eastern parts of the Barents Continental Rise and Abyssal Plain varies from 2 to 3 km and gradually decreases toward the Gakkel Ridge. The thickness of sediments within the Yermak Plateau varies from 1 to 2 km.

The sediment thickness map (Fig. 4) also indicates that the predominantly buried relief of the basement follows the two main tectonic trends of the area, i.e. an ENE-SWS trend (the trend of the Gakkel Ridge, Eastern Yermak Plateau and the Barents Continental Slope) and a NNW-SSE trend (parallel to the fracture zones of the Gakkel Ridge). Applying the MSW method over larger parts of the basin eastwards may confirm whether this is a systematic feature of the sediment thickness distribution of the Nansen Basin.

It is proposed to use the new sediment thickness grid and map under international project “Map of the Arctic Sediment Thickness” (MAST).

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Figure 6. Statistical estimations of the uncertainty of the method. Correlation between depth to acoustic basement from seismic data and two types of magnetic depth estimates are shown. In A, depth estimates are extracted from the database (Fig. 3) and projected onto seismic profiles from different distances. In B, the same points are extracted from the depth to basement grid and projected onto the lines. R and σ are the computed correlation factor and standard deviation (km), respectively.
information required for uncertainty estimation of the approach developed, and for constructive comments to our study. We sincerely thank Robert Scott and Dennis Thurston for improvements on the manuscript and Olav Eldholm for fruitful discussions and collaboration.

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RESULTS OF DENSITY MODELING OF THE MAJOR STRUCTURAL ELEMENTS OF THE ARCTIC OCEAN

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ABSTRACT

Four density models based on Russian deep seismic sounding, seismic refraction and reflection data were constructed in the deep Arctic Basin. Density structures of the Lomonosov, Mendeleev Ridges and the Makarov and Podvodnikov basins were defined using available seismic data as reference information. In the absence of seismic data, configurations of the main deep surfaces (basement and Moho) were revealed from gravity anomalies. In particular it was done along some parts of profile that stretches from the shelf of Frantz Josef Land to the Lomonosov Ridge and crosses the Gakkel Ridge. The gravity and magnetic modeling software GM-SYS was applied. Results of depth to magnetic source estimations from available aeromagnetic data were used as additional information.

INTRODUCTION

It is generally agreed that the Eurasia Basin including the Gakkel Ridge and adjacent basins was developed by seafloor spreading and that the Lomonosov Ridge is a continental sliver rifted off the Eurasia continental margin. At the same time, the deep structure and evolution of the Amerasia Basin up till now has remained the subject of continuous debate. In particular there is no common point of view on the age and geological nature of the crust of the Alpha and Mendeleev Ridge as well as of the Makarov and Podvodnikov basins.

The main goal of the work was to construct integrated geophysical models of the deep structure of the above-mentioned geological provinces on the basis of 2-D gravity modeling of available seismic data and results of depth to magnetic source estimations.

We propose solutions to the following problems:

1) To specify the structure of crust within individual parts of deep seismic sections with uncertain correlation of seismic boundaries;
2) To define density parameters of layers and blocks in the models of crust;
3) To determine both the basement and the Moho relief in the areas where there is no seismic registration;
4) To identify problematic areas where the results of seismic and potential field data interpretation disagree with each other and where additional analysis of all available information is required.

DATA

Initial seismic information used for modeling is presented by four seismic sections that were constructed by our colleagues from both Polar Marine Geological Research Expedition (PMGRE) and VNIIOkeangeologia. The location of seismic and density models is given in Figure 1.

Seismic section 1 is based on information that was...
collected along central line of the geotransect “De Long Islands – Makarov Basin” that crosses the East-Siberian shelf, the continent-ocean transition zone, and the Podvodnikov and Makarov basins. Its total length is 1480 km. It consists of three parts investigated during three seasons from 1989 to 1991. Seismic investigations included deep seismic sounding (DSS) and single-channel seismic reflection. Detailed description of seismic observation procedures, data processing and final seismic sections along profile 1 are presented in a few Russian publications (Gramberg et al., 1993; Sorokin et al., 1999; Zamansky et al., 1999). Gravity observations were carried out from stations on drifting ice along the seismic line. Magnetic data in the area is represented by a 200 km swath of detailed aeromagnetic surveys. The seismic section constructed by Zamansky et al. (1999) using the IRIS software was chosen as the most reliable for gravity modeling (Fig. 2a).

Seismic section 2 is based on information that was collected in 1992 within geotransect “Amundsen Basin – Podvodnikov Basin” that starts in the Eurasia Basin and crosses the Lomonosov Ridge (Fig. 3a). Its total length is 280 km. The central profile 2 is provided by DSS and single-channel seismic reflection investigations. The seismic velocity model was constructed using the SeisWaid program (Ivanova et al., 2002). There were no direct gravity observations along the seismic line, and so the gravity anomalies for density modeling were extracted from a grid with cell size 10×10 km (Glebovsky et al., 2002). There are not many real points of gravity observations nearby the

**Figure 2.** Seismic (Zamansky et al., 1999) (a), and density (b) models along seismic section 1. Legend: 1. Block number; 2. Velocity, km/s; 3. Moho; 4. Sediments: I - (1.7 - 2.0 km/s), II - (2.0 - 2.6 km/s), III - (2.3 - 3.25 km/s), IV (3.4 - 4.2 km/s), V (4.5 - 5.6 km/s); 5. Crust: VI - (5.89 - 6.52 km/s), VII - (6.5 - 7.62 km/s); 6. Mantle: VIII - (7.81 - 8.2 km/s); 7. Density (g/cm³); 8. Depths to magnetic sources; 9. Seismic boundaries.
Seismic profile and so gravity anomalies are smoothed under interpolation and gridding. Magnetic data in the area is represented by a 200 km swath of detailed aeromagnetic surveys as in geotransect 1.

Seismic section 3 “Podvodnikov Basin – Mendeleev Basin” is based on information that was collected in 2000 along a DSS and single-channel seismic reflection profile that crosses the Mendeleev Ridge. The length of the profile is 460 km (Fig.3b). Seismic observations were accompanied by on-ice

Figure 3. Seismic (Ivanova et al., 2002; Zamansky et al., 2002) (a), and density (b) models along seismic sections 2 and 3. Legend 1. Block number; 2. Velocity, km/s; 3. Moho; 4. Sediments: I-IV - (1.7 - 5.6 km/s); 5. Crust: V-VII - (6.0 - 7.5 km/s); 6. Mantle: VIII - (7.9 - 8.2 km/s); 7. Density (g/cm³); 8. Depths to magnetic sources; 9. Seismic boundaries; 10. Density boundaries.
gravity measurements. The seismic velocity model along line 3 was constructed using the SeisWaid program (Zamansky et al., 2002). Magnetic data were not used for modeling because of poor quality and data coverage.

Seismic section 4 “Frantz Josef Land – Lomonosov Ridge” that crosses the Gakkel Ridge and the adjacent Nansen and Amundsen basins (Fig. 4a) is compiled from several data sets (Daragan-Suschov et al., 2002): (1) A 200 km long seismic refraction profile that starts from the shelf of the Frantz Josef Land, crosses continental slope and reaches out to the Nansen Basin; (2) single seismic reflection soundings in the Nansen Basin; and (3) seismic reflection observations from drifting ice-stations “North Pole-23” and “North Pole-24”.

Gravity anomalies along profile 4 were extracted from a grid with cell size 10×10 km (Glebovsky et al., 2002). In the deep part of the Eurasia Basin the gravity anomalies are smooth while seismic data are detailed. Therefore, the relief of both seafloor and basement was smoothed by a 20-km filter to improve correlation with gravity data.

Magnetic data in the area is taken from both Russian (PMGRE - 1993, 1998) and US Navy (NRL - 1975, 1999) aeromagnetic surveys.

**MODELING PROCEDURE**

Gravity models were constructed by applying the GM-SYS software. The basis for the 2-D modeling is the algorithm of Rasmussen and Pedersen (1979). Densities for crustal layers were calculated according to density-velocity relations (Nafe and Drake, 1967; Rusakov, 1991).

Magnetic profile data were used as additional information for 2-D modeling of basement relief. Depths to magnetic sources were calculated by a multiple-source version of the PDEPTH program (Cordell et al., 1992) based on the Werner Deconvolution method (Hansen and Simmonds, 1993). Depth estimates from magnetic profiles within 50 km of the seismic profile zone were projected on the seismic section. Detailed description of the method is presented in (Likhachev et al, 2004; Glebovsky et al, this volume).

Gravity modeling along seismic sections 1-3 (Figs. 2, 3) was aimed at specifying physical characteristics of the layers and blocks in the model. Seismic observations that were carried out along each of these sections revealed boundaries within the sedimentary cover and the configuration of the acoustic basement surface and Moho. Results of forward density modeling show considerable difference between computed and observed gravity anomalies. The difference was minimized by inserting model blocks with altered densities in the mantle and consolidated crust, and by minor correction of Moho and basement boundaries.

Seismic section 4 (Fig. 4) reveals the structure of the sedimentary cover and fragments of basement relief. Density modeling here was aimed at defining Moho and basement relief (were there is no seismic data), and the density structure of the lithosphere. At the first stage of modeling the gravitational effect of the upper crust was estimated using available seismic data. Next, residual gravity anomalies were calculated as the difference between observed anomalies and those from the upper crust. Residual anomalies were the basic information to model the Moho surface, using a technique developed by Fedorov et al. (2001). The technique is based on the generalized method of regularization that marginally determines physical-statistical regularities of crust bending. The Moho surface was similarly determined along the full length of profile 4 except of the axial zone of the Gakkel Ridge. High amplitude gravity anomalies and the thermal softening of the lithosphere in the axial zone of the Ridge was compensated by 3.1 g/cm³ assigned density of the mantle block. Basement relief in the Nansen Basin where there is not enough seismic data was modeled using depths to magnetic source estimations.

Thus, 2-D gravity modeling was carried out in interactive regime by comparison of density models with seismic sections and with results of depth to magnetic source estimations.

**RESULTS**

**Density model 1 “De Long Islands – Makarov Basin”**

The geometry of the density model (Fig. 2b) in general corresponds to that on the seismic section. Minor corrections of deep boundaries were done on the basis of gravity data.

Thick continental crust of the De Long Islands shelf was divided into two layers with essentially different density - 2.78 g/cm³ and 3.05 g/cm³ - to achieve the best correspondence of the gravity model with the seismic section. The density of the upper mantle within the shelf area, 3.3 g/cm³, is typical for continental lithosphere. The continent-ocean transition zone is marked by a large positive gravity anomaly corresponding to the rise of Moho from 43 to 20 km depth. The density of consolidated crust increases accordingly and its thickness drastically decreases to 5-16 km in the deep ocean.

The thickness of the sedimentary cover in the Podvodnikov and Makarov basins varies between 3 and
Figure 4. Seismic (Daragan-Suschev et al., 2002) (a), and density (b) models along line 4. Legend 1. Block number; 2. Sediments: I - (1.8-2.0 km/s), II - (2.2 km/s), III - (2.7-2.9 km/s), III-IV - (2.7-3.6 km/s), IV - (3.3-3.6 km/s), V - (4.2-4.6 km/s), VI - (5.2 km/s); 3-5. Seismic sequences in Franz Josef Land: 3. - 3.5 km/s, 4. - 4.5-4.7 km/s, 5. - 5.0-5.3 km/s; 6. Crust; 7. Faults; 8. Density (g/cm³); 9. Depths to magnetic sources.
7 km. At 600-870 km from the beginning of the profile there is a basement high characterized by enlarged thickness and density of crust. This buried high is almost not present in the sea bottom relief. A Mesozoic spreading centre was interpreted here from geochronological analysis of magnetic anomalies (Kovacs et al., 1999). As a result, this high may be considered as an “anomalous” oceanic structure composed of a thick layer of serpentinized peridotites (Gurevich and Maschenkov, 2000; Astafurova et al., 2002).

The profile crosses two more rises of basement at 1100 and 1370 km distance from its beginning. Both of them are revealed in the bathymetry and are characterized by an increase in thickness and a decrease in the density of the crust. The nature of these rises is debatable.

By density and thickness the deep structure of the Podvodnikov and Makarov basins differs from both continental and normal oceanic structure. It is characterized by less mantle density and thinner crust than normal continental crust. At the same time it has thick sedimentary cover and thicker crust than normal oceanic crust.

Density model 2 “Amundsen Basin–Podvodnikov Basin”

Results of forward density modeling along seismic section 2 (Fig. 3) show considerable difference between computed and observed gravity anomalies over the Lomonosov Ridge. Two additional blocks within its mantle and crust were simulated to compensate the residual positive gravity anomaly. The density of the upper mantle here amounts to 3.3 g/cm\(^3\) which is typical for continental lithosphere. The high density of the consolidated crust (3.0 -3.1 g/cm\(^3\)) beneath the Lomonosov Ridge may be explained by its tectonic restructuring during the Eurasia Basin opening. Seismic section in this area is characterized by drastic changes in the velocities.

Seismic data and density modeling show that the Amundsen Basin has thin (5 km) consolidated crust covered by a sedimentary layer with thickness about 2 km, and underlain by low-density mantle. All characteristics listed above agree with oceanic lithosphere.

The top and bottom of the lower layer of consolidated crust in the Podvodnikov Basin defined from seismic data was forcibly modified for better correspondence between observed and modeled gravity anomalies.

The modest accuracy of the curve fitting may be explained by poor gravity data coverage and correspondingly by the smooth shape of the observed gravity anomaly. The shape of this anomaly should to be more complicated as judged from the bottom relief from both the seismic section and the IBCAO grid (Jakobsson et al., 2000).

Density model 3 “Podvodnikov Basin–Mendeleev Basin”

In a similar manner to the first two gravity models, a few blocks with different densities were simulated within the mantle along seismic section 3 (Fig. 3). Density of the mantle below the Mendeleev Ridge is 3.3 g/cm\(^3\), whereas on the Mendeleev Basin side it is essentially reduced (3.23 g/cm\(^3\)). An abrupt, 8-10 km upwarp of the Moho indicated by the seismic model near the Podvodnikov Basin – Mendeleev Ridge transition was smoothed on the basis of gravity modeling. The other seismic boundaries were not modified.

In spite of relatively thick crust in the Mendeleev Basin, reduced density of underlying mantle may be evidence of its oceanic nature. This assumption requires additional study.

The result of the modeling does not provide any clear answer concerning the nature of the Mendeleev Ridge. A thin upper consolidated crust (about 4 km), increased thickness (about 23 km) and high density of the lower crust may be indications of an anomalous type of oceanic crust here. At the same time the total thickness of crust (up to 32 km) and the high density of the mantle indicate a continental nature. From seismic data alone the Mendeleev Ridge appears to be continental (Poselov et al., 2002).

Additional regular gravity and magnetic surveys are required to define more accurately the nature of the Mendeleev Ridge and the ridge-shelf transition.

Density model 4 “Frantz Josef Land–Lomonosov Ridge”

In contrast to density models along profiles 1-3 (Figs. 2b, 3b) the Moho and partly the basement relief along profile 4 were modeled from gravity and magnetic data. The modeled Moho depth beneath Frantz Josef Land (FJL) is 28-30 km whereas under the Lomonosov Ridge it is 23-26 km. The density of continental crust within both continental structures is less than that of the adjacent oceanic basins.

The continent – ocean transition is marked by a sharp rise of the Moho near both the FJL shelf and the Lomonosov Ridge (Fig. 4b).

The Nansen and Amundsen basins are underlain by thin (about 6-8 km) oceanic crust.

The modest accuracy of the curve fitting may be explained by poor gravity data coverage and correspondingly by the smooth shape of the observed gravity anomaly. The shape of this anomaly should to be more complicated as judged from the bottom relief from both the seismic section and the IBCAO grid (Jakobsson et al., 2000).

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Additional regular gravity and magnetic surveys are required to define more accurately the nature of the Mendeleev Ridge and the ridge-shelf transition.
Gakkel Ridge province is decreased with respect to the adjacent basins. Also, the crustal thickness near the axis of the ridge is decreased to 4.5 km.

Oceanic crust in our model was divided into two layers – basaltic and gabbroic – by analogy with standard models. Results of rock sampling in the Gakkel Ridge in 2001 (Cochran et al., 2003; Michael et al., 2003) show that the gabbroic layer here is either very thin or absent. Moreover, the shape of the mantle wedge in our model is rudimentary simulated without taking into account thermal structure of the lithosphere. The new map of magnetic isochrones in the Eurasia Basin that was recently constructed at VNIIOkeangeologia (Glebovsky et al., 2004) provides information to estimate the thermal expansion of mantle depending on the age of oceanic crust. We propose to redo the current model including thermal structure of the lithosphere by the same way as, for example, (Brevik et al., 1999).

CONCLUSIONS

1) The Amundsen and Nansen basins are underlain by thin (about 6-8 km) oceanic crust. The thickness and density of the crust increase with age.

2) The Podvodnikov and Makarov basins have crustal thicknesses varying from 10 to 24 km. The thickness of sediments here varies from 3 to 7 km. The consolidated part of the crust has thickness that varies from 4 to 16 km and is characterized by 2.86 g/cm³ density. Thus, according to thickness and density characteristics the crust of the Podvodnikov Basin is of intermediate type and may be considered as old oceanic or thinned continental. The first assumption is confirmed by results of magnetic anomaly field interpretation that revealed a Mesozoic spreading center in the Basin (Kovacs et al., 1999). The second assumption is based on analysis of seismic and bathymetry data (Poselov et al., 2002).

3) The modeling supports widely accepted points of view on the oceanic nature of the Gakkel Ridge crust and on the presence of thinned (26 km) continental crust under the Lomonosov Ridge.

4) The Mendeleev Ridge is slightly different from a typical continental structure by density characteristics, but has thick crust (about 32 km), that is not typical for oceanic regions. That is why we think it is continental rather than oceanic. Additional geological and geophysical data hopefully will help to solve the problem regarding the nature of the Mendeleev Ridge.

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GEOLOGICAL ORIGIN OF THE MAGNETIC ANOMALY FIELD IN THE CENTRAL AMERASIAN BASIN (ARCTIC OCEAN)

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ABSTRACT

Understanding the origin of the Alpha-Mendeleyev Rise is integral to understanding the Amerasia Basin and has critical implications for claims under the United Nations Convention on the Law of the Sea. It has been proposed that these structures are the result of an ancient mid-ocean spreading center; or an oceanic volcanic plateau with a hot-spot trace; or a fragment of modified continental crust. We propose through study of magnetic anomalies and comparison with analogue structures, that the Alpha-Mendeleyev Rise is continental crust modified by intraplate magmatism.

INTRODUCTION

The geological origin of the Alpha Ridge-Mendeleyev Rise system in the Central Amerasian Basin (Fig. 1) is controversial and critical under the United Nations Convention on the Law of the Sea Article 76 of 1982. The magnetic anomaly field has high amplitude of the interference type which regionally forms an ensemble of conformal anomalies subjected to the general structure of the magnetic field (Fig. 2). In the area of Greenland and Canada archipelagos these magnetic anomalies terminate abruptly at the continental slope. North of Ellesmere Island, magnetic anomalies are similar to anomalies of the Alpha Ridge.

In the south-eastern part of the Amerasian province in the area of the Chukchi Borderland block, local magnetic anomalies have no extension in the southern part of the eastern Arctic shelves. The magnetic field of the shelves of the eastern seas is distinguished by a predominance of weak intensity linear--mainly sub–latitudinal--anomalies against the background of negative field values. There are no analogues of the Amerasian type anomalies.

ORIGIN THEORIES

There are three alternative points of view on the
geological origin of the Alpha Ridge and Mendeleyev Rise magnetic anomalies.

According to one of them, the Alpha Ridge is an ancient spreading centre (Vogt and Ostenso, 1970; Hall, 1973; and Karasik 1980). This theory characterizes the Alpha Ridge as an ancient mid-oceanic ridge having a lengthwise zoning typical of mid-oceanic ridges.

These authors proposed that in the middle of the Cretaceous weakened zones in the Amerasian Basin were subjected to intra-plate volcanism and vertical tectonic movements which resulted in the formation of the modern Alpha Ridge. This explanation of the nature of the Amerasian province magnetic anomalies comes into conflict with data of crustal thickness in this ocean area. According to seismic data, a thick crust up to 40 km is established here that is not consistent with existing views of mid-oceanic ridges—the crust thickness under which, as a general rule, is no more than 7-8 km.

An alternative viewpoint holds that the Alpha Ridge and Mendeleyev Rise system is an oceanic volcanic plateau overprinted by a trace of an active hot spot. This view is shared by Vogt et al. (1986), and Forsyth et al. (1986). An increase of thickness of the so-called "lower basaltic layer" of up to 15 km and the presence of relatively high-velocity rocks (4.7-5.1 km/s) in the upper part of the section, have been established on the Alpha Ridge from the data of CESAR expedition. The thickness of these rocks is 8 km. To take account of these data, Forsyth et al. (1986) concluded that it can be correlated with the velocity section of Iceland and the Pacific Ocean rises of Manihiki and Ontong-Yava. He believes the Alpha Ridge crust appears to be ancient oceanic crust that was altered many times in later stages of its evolution. In the early stages it was an ancient mid-oceanic ridge with a dissected topography that existed in the area of the central Arctic rises. It is precisely this topography that defines the complicated appearance of the magnetic field above it. However, Jackson and Johnson (1986) note: “where hot spot traces have traversed previously formed oceanic crust as in large oceanic plateau, a distinct free-air gravity pattern with negative values on either side of 200 mGal positive is formed which is not observed in this region. Here values free-air gravity anomaly changes limit -20mGal - +30mGal”.

Disagreeing with Forsyth et al. (1986), Kiselyov (1986) comparing the same velocity section of the Alpha Ridge with the Greenland section, established their complete similarity and concluded that the structure of the Amerasian Basin has a continental nature.


In addition, according to the scheme worked out by Karasik (1980) for magnetized bodies, the area of the central Arctic rises is characterized by high values of effective magnetization (2 A/m) that is typical for continental formations. The magnetic section of the crust also has a two-layer structure. Magnetic sources at 1-4 km depths coincide with the acoustic basement topography composed of volcanogenic rocks. The second deeper-level source of magnetic anomalies corresponds to depths of 7-8 km. The two-layer magnetic model of the crust is also typical for the Lomonosov Ridge, whose continental crustal nature is accepted by many investigators (e.g. Kiselyov, 1986, Volk, 1992). The range of depth for the lower magnetic "horizon" at the Lomonosov Ridge and the Mendeleyev Rise is 8-10 km and based on seismic data correlates to the surface of crystalline basement.

SPECTRAL CHARACTERISTIC OF MAGNETIC ANOMALIES

We have conducted an investigation to "recognize image" by frequency-spectral characteristics of magnetic anomalies (continental and oceanic) of lithospheric blocks in the Amerasian Basin and to establish the geological nature of the Alpha Ridge and Mendeleyev Rise (Fig.2). To accomplish this task, we used a grid of magnetic anomalies, originally released by the Geological Survey of Canada (Verhoev et al., 1996) and refined by the Russian data on the Amerasian Basin. On the basis of this data, polygons for analysis were chosen. The magnetic anomaly structure of the fields of Mendeleyev Rise and the Alpha Ridge is visually similar. Three polygons have been chosen as such analogues: the Greenland-Faeroe Ridge, a zone of intensive linear anomalies above the Anabar Shield, and an area of similar magnetic anomalies of the Tunguska Basin.

Magnetic fields of these three areas have been corrected to facilitate correlation with the field of the Alpha Ridge and Mendeleyev Rise, where average depths of occurrence for the first magnetic "horizon" in the crust 2.0 - 2.5 km from marine seismic data. Then for each polygon an analysis was done of the anomalies spectral characteristics according to the standard
method. Results of the calculations are presented as amplitude and power spectrum histograms in Figure 3.

CONCLUSIONS

The spectral characteristics of magnetic anomalies of the Alpha Ridge and Mendeleyev Rise are equally correlated with both the Greenland-Faeroe Ridge, which is an oceanic morph structure complicated by a long period of "hot spot" volcanic activity (including modern), and with the Anabar Shield located in the core of the ancient craton (Fig. 2). More noticeable distinctions can be defined only in correlation with the Tunguska Basin (Fig. 2). We should point out, however, that the structure of magnetic anomalies within a relatively large Tunguska Basin polygon (400 km x 400 km) consists two approximately equal areas, one with a short-wave and one long-wave fields that are clearly distinguished even visually.

The correlation with the Tunguska Basin is of a great interest because this large element of the ancient Siberia Platform is characterized by intense manifestation of volcanism. The trappean complex of the Tunguska Syncline has been studied in considerable detail (Al'mukhamedov et al, 1999). It has been established from the results of drilling that the thickness of Permian-Triassic traps reaches 4 km and resulted from a very powerful impulse of midland volcanism. The areal distribution of the trappean
Figure 3. Amplitude and power spectrum histograms for magnetic anomaly field.
volcanism of the Siberian Platform is reasonably
correlative with the size of the province of magnetic
anomalies in the Amerasian Basin.

Indirect evidence for the continental nature of the
Alpha Ridge–Mendeleyev Rise can be found in the
analysis of Bouguer anomalies (Fig. 4) calculated for
the Amerasian Basin based on a three-dimensional
model by Mashchenkov (2000). In the central part of
the Amerasian Basin there are negative anomalies up to
30 mGal, typical for blocks with continental crust. The
thickness of the crust varies within the range 17 - 40
km (Volk, 1992).

On the basis of the data review and analysis, we
are inclined to recognize the continental nature of the
Amerasian Basin magnetic anomalies and to link their
appearance with the regional manifestation of midland
magmatism of the Cretaceous age. This is supported by
the similarity of magnetic anomalies of the Alpha
Ridge with their extension at the northern Ellesmere–
Greenland area.

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VELCITY STRUCTURE AND CORRELATION OF THE SEDIMENTARY COVER ON THE 
LOMONOSOV RIDGE AND IN THE AMERASIAN BASIN, ARCTIC OCEAN

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ABSTRACT

Russian (and previous Soviet) geophysical surveys of the high Arctic Ocean included wide-angle refraction and shallow reflection seismic profiling, gravity and magnetic surveys. During this work, special reflection seismic experiments were carried out over four widely separated areas to determine the velocity structure of the sedimentary cover. These morphological provinces are the East-Siberian Continental slope, the Arlis Rise, the Makarov Basin, and the Lomonosov Ridge.

Three main sedimentary units were recognized in all areas and correlated on the basis of previous seismic operations, particularly the profiles acquired on NP-28, which crossed the Arctic Ocean from the Podvodnikov Basin to near the North Pole and onwards to the Yermak Plateau. The three units – I, II and III, have interval Vp velocities of 1.7-1.9 km/s, 2.0-3.0 km/s and 3.0-4.0 km/s, respectively.

On the basis of correlation with the results of the IODP ACEX 2004 drilling on the Lomonosov Ridge, the three sedimentary units are inferred to be Neogene, Paleogene and Mesozoic (Cretaceous, perhaps older) in age. This work supports previous studies that have concluded that the Eurasian Basin opened in the Cenozoic and the Amerasian Basin partly in the Mesozoic. It also provides evidence of fragments of continental crust in the Amerasian Basin.

Key words:


INTRODUCTION

Geophysical research in the Amerasian part of the Arctic Basin and the Lomonosov Ridge were carried out annually by the Polar Marine Geological Research Expedition (PMGRE) between 1989 and 1992 (Fig. 1). This work was part of a general program to define the northern border of the Russian continental shelf. The research included about 1700 km of the wide-angle refraction and shallow reflection seismic profiles for investigating crustal and upper mantle structure (Gramberg et al., 1993; Sorokin et al., 1999; Zamansky et al., 1999; Ivanova et al., 2002; and Lebedeva-Ivanova et al., 2004). It also included complementary shallow reflection profiles along some of the wide-angle profiles for definition of the sedimentary units overlying acoustic basement.

During these major comprehensive geophysical investigations, four widely separated areas were selected for special reflection seismic experiments to determine the velocity structure of the sedimentary cover. These data provide the basis for subsequent interpretation of the velocity structure of the basement. The four areas were selected as being representative of four of the main morphological provinces in the high Arctic – the East-Siberian Continental slope (TRA(b)-91), the Arlis Rise (TRA(b)-89), the Makarov Basin (TRA(b)-90), and the Lomonosov Ridge (TRA(b)-92) (Fig. 1). Previous ice-station and other reflection seismic measurements, for example on NP-28 (Sorokin et al., 1999) allow some of the results of the four experiments to be directly compared.

On the basis of this reflection imaging and velocity data, an attempt is made to estimate the age and character of the Mesozoic and Cenozoic sedimentary sequences in the Arctic Basin.

EXPERIMENTAL METHODS

At each of the four locations for the detailed reflection measurements, a seismic recording array was set up and operated for long enough (a couple of weeks) to acquire a profile of about 100 km length. The ice-drift velocity generally was in the order of 5-10 km/day, but the direction was variable and unpredictable. Figure 1 shows the tracks of the drift of the four ice-station (TRA(b)-89, 90, 91 and 92).
The seismic receiver array was in the form of an orthogonal cross (Fig. 2A), the arms of which were either 1150 m (TRA(b)-89 and 90, over the Arlis Rise and the Makarov Basin) or 550 m (TRA(b)-91 and 92, over the East-Siberian Continental slope and the Lomonosov Ridge) in length. Receivers were distributed along each of the four arms of the array at a distance of 50 m. The shots were provided by 3-5 detonators (electric caps) placed in the centre of the array in water at a depth of 8 m. The shots were fired every 2-4 hours, depending on the rate of the drift of the ice-station. This frequency allowed data to be acquired at 0.5-1.0 km intervals.

To increase precision in the velocity determinations, one of the arms of the array was lengthened up to 7 km; 2 or 3 extra shots were detonated along that arm 2-3 times a day for increasing the shot-distance interval to 3-7 km (Figs. 2B and 2C).

The seismic recording equipment employed for these experiments was composed of SMOV-0-24 seismic stations, providing analog magnetic records. The length of the recordings was 12 sec.

The location of the array was determined using a MX-1502 receiver of the Transit SNS system and a MX-4400 receiver of the GPS system. The average coordinate error did not exceed 300 m.

**PROCESSING**

In 1995-1996, the analog seismic records were digitized at PMGRE and reformatted into SEG-Y files. The data have been processed using the ProMax system with parameters given in Table 1.
Velocity analysis was performed on those reflections which correspond to the main horizons identified on profile NP-28. Areas for this velocity analysis were chosen where the reflection geometry was relatively horizontal. The obtained normal moveout velocities (NMO) were converted to interval velocities using the Dix formula. Examples of the velocity analysis are shown in Figure 3. The mean interval velocity for units I, II and III has been calculated for each unit in all profiles. Interval velocities show a deviation of up to 4%, except for over the Lomonosov Ridge (TRA(b)-92), where unit I shows a deviation of 7%.

RESULTS

Time seismic sections (Figs. 4 and 5) were constructed along the profile in the four areas. Reflections were traced and marked in the sections. On all sections, the sea-bottom, acoustic basement and several intervening reflections are readily recognizable. After each of the profiles had been examined and the reflections traced and marked, the velocities for the different sedimentary units were calculated. Thereafter, on the basis of the seismic reflection data made along NP-28 and TRA-91 (Fig. 1 and Fig. 6), the most prominent reflections and sedimentary units were correlated (see below) and designated accordingly:

<table>
<thead>
<tr>
<th>Processing step</th>
<th>Details</th>
</tr>
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<tbody>
<tr>
<td>1</td>
<td>Sort to common source and common receiver gathers</td>
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<tr>
<td>2</td>
<td>Define zero time</td>
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<tr>
<td>3</td>
<td>Trace editing and muting</td>
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<td>4</td>
<td>AGC window 1 sec.</td>
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<td>5</td>
<td>Minimum phase predictive deconvolution</td>
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<td></td>
<td>- prediction distance- 8 ms</td>
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<td></td>
<td>- operator length - 200 ms</td>
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<tr>
<td></td>
<td>- white noise - 0,1 %</td>
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<td></td>
<td>- deconvolution gate - wide</td>
</tr>
<tr>
<td>6</td>
<td>Bandpass filter 15-18-45-55 Hz.</td>
</tr>
</tbody>
</table>

"d" – sea bottom;
Unit I – uppermost sedimentary unit;
"d_1" – first prominent reflection;
Unit IIa
"d_2" - second prominent reflection (not always present);
Unit IIb
"A" – main composite reflection;
Unit III – locally with intervening reflections A_1, A_2, A_3;
"Af" – acoustic basement.

These designations, where applicable, are marked on all the profiles along with the velocity data.

Description of Profiles

The four profiles are described briefly below and illustrated in Figures 4 and 5.

The Lomonosov Ridge (TRA(b)-92), c. 84° N
Figure 4A shows the 30 km long NW-SE trending part of the Lomonosov Ridge profile (this profile bends back on itself for the rest of the acquisition period, see Fig. 1). Figure 4A covers the western part of the Ridge at depths of c. 2.5 km and a gentle slope down to 3.1 km towards the Amundsen Basin. Two prominent flat-lying reflections are conspicuous, the upper ("d_1") at c. 2.8 km in depth extending across the Ridge and being truncated by the slope into the Basin. The lower zone ("A") is composite and stronger; it occurs at a depth of 3.3 km on the Ridge and can be followed down the slope into the Basin to the end of the profile. Beneath these reflections at c. 4.0-5.5 km depth there is a less persistent and somewhat irregular level of reflections ("A_f") which either dies out before reaching the slope into the Basin or persists, weakly, beneath the latter.
The Makarov Basin (TRA(b)-90), c. 87-88° N

Figures 4B shows an approximately 90 km long NW-SE profile across the Makarov Basin, located immediately to the southeast of the Lomonosov Ridge. This remarkable profile shows a regular horizontal sea-bottom at 3.8 km depth; underlain by three parallel, approximately horizontal reflections marked "d1", "d2" and "A". The last of these is particularly strong and composite and transgresses an underlying "basin-ridge-basin" structure. The top of the ridge is defined by a strong composite reflection (A_F) that is inferred to be acoustic basement; there is much irregular reflectivity within the ridge. The basins to the northwest and southeast of the ridge are defined by gently concave reflections marked "A1", "A2", etc.; of particular interest is a low velocity (2.8 km/s) unit beneath the "A" horizon.

The Arlis Rise (TRA(b)-89), c. 83.5° N

Figure 5A shows a 120 km long NW-SE trending profile across the Podvodnikov Basin, where the latter narrows over the Arlis Rise; this feature marks the 500 m step down from the southern to northern parts of the Podvodnikov Basin. The profile is oriented at a high angle to the TRA-1989 transect and extends along the northern margin of the Arlis Rise, towards the Lomonosov Ridge, but does not reach the latter. Water depth along the profile decreases northwards from about 3.0 km in the south to about 2.5 km in the north. The sea-bottom and underlying reflections ("d1" and "A") are broken by young faulting near the northwestern end of the profile; further southeast similar down-stepping is accommodated by gentle folding of the sedimentary units. The section is highly reflective, with "d1" prominent and closely underlain by...
"A". The strong reflection packet "A_F" is less regular then "A" and various reflections ("A_1", "A_2") can be traced sporadically. Towards the northwestern end of the profile, "A" and "A_F" reflections converge above a basement high.

The East Siberian Slope (TRA(b)-91), c. 80° N
This c. 110 km long S-N profile (Fig. 5A), reaches from approximately 40 km north of the edge of the East Siberian Shelf northwards across the slope towards the southernmost part of the Podvodnikov Basin. The water depth increases northwards from around 1.0 km to 2.5 km and the reflections likewise dip gently northwards. The profile is dominated in the lower parts by strong multiples. The uppermost part of the section is highly reflective and horizons "d_1" and "d_2" are recognized, overlying a strong "A" reflection at about 5 km depth. Acoustic basement is obscured by multiples.

Description of the sedimentary units.
The ice-station NP-28 track, together with reflection data from TRA-91 (Fig. 6) provided a basis

Figure 4. Time section (A) along the TRA(b)-92 profile and (B) along TRA(b)-90 profile. Locations of profiles shown on Figure 1. Letters refer to the inferred surfaces of the main reflection boundaries within the sedimentary cover; V is the value of the interval velocities of P-waves; arrows show locations of the examples of shot gathers on Figure 3.
for correlating reflections from the East Siberian Shelf to the Makarov Basin and thereafter across the Lomonosov Ridge. The areas reported here lie close enough to the NP-28 or TRA-91 profiles for these to provide a basis for correlation. Figure 7 shows typical parts of sections, facilitating comparison of the different areas. Below follows a description of the different sedimentary units and their correlation.

**Unit I**

Between the sea-bottom and first prominent reflection ("d1"), Unit I generally has a Vp velocity of 1.7-1.9 km/s. It is seen nearly everywhere lying conformably on Unit II. Only along the Lomonosov Ridge profile, on the slope into the Amundsen Basin is this unit cut out. In general, Unit I has a fairly uniform thickness of about 300-400 m; towards the East Siberian slope it increases to more than double this amount.

**Unit II**

Unit II has a velocity Vp of 2.0-3.0 km/s. It rests conformably (e.g. in the Makarov Basin), or with very low angle discordance (e.g. in the Podvodnikov Basin) on the underlying reflection "A". In some areas, such as the Makarov Basin, reflections within this unit (e.g. "d2"), allow it to be subdivided. Unit II's thickness in the Makarov Basin is c. 1.3 km. Over the Lomonosov Ridge profile, on the slope into the Amundsen Basin is this unit cut out. In general, Unit I has a fairly uniform thickness of about 300-400 m; towards the East Siberian slope it increases to more than double this amount.

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*Figure 5. Time section (A) along the TRA(b)-89 profile and (B) along TRA(b)-91 profile. Location of profiles shown on Figure 1. Letters refer to the inferred surfaces of the main reflection boundaries within the sedimentary cover; V is the value of the interval velocities of P-waves; arrows show locations of the examples of shot gathers on Figure 3.*

184

Langinen et al.
Ridge, Unit II is about half this thickness. In the vicinity of the Arlis Rise, this unit is even thinner or absent, but further south across the Podvodnikov Basin, towards the East Siberian slope, it increases again to about 1.5 km.

**Unit III**

Beneath the composite reflection packet "A", Unit III, in general, has a velocity of 3.0-4.0 km/s. It rests unconformably on the irregular surface of the acoustic basement (reflection "A_F"), and its thickness and subdivision varies accordingly. In the Makarov Basin, Unit III is divisible into at least two subunits (by reflection "A_1" and "A_2") and its total thickness reaches 4 km. Over the Lomonosov Ridge, the thickness is up to 2 km. The Arlis Rise profile shows a thinning of Unit III from 1.2 km to 0.3 km towards the Lomonosov Ridge. Further south, in the southernmost part of the Podvodnikov Basin, approaching the East Siberian slope, this unit thickens again to at least 3 km (reflection "A_F" obscured by multiples). Of particular interest in the Makarov Basin and Arlis Rise area has AWI-91091, extend from the the Amundsen Basin scarp eastwards across the Ridge to the slope towards been the identification of low velocity subunits (2.8 km/s and 2.3 km/s, respectively) in Unit III directly below the "A" reflection.

**Acoustic Basement**

The physical properties of the acoustic basement have been treated elsewhere (Lebedeva-Ivanova et al., 2004), based on analysis of wide-angle refraction data. Various Vp velocities have been defined of around 5 km/s.

**Stratigraphic correlation of the units**

During the last fifteen years, since the reflection measurements described here were acquired, several expeditions to the high Arctic Basin have collected reflection seismic data (Kristoffersen, 2001, Jokat 2005 and references therein). Grantz et al. (2001) reported piston-core evidence of Cenozoic and late Mesozoic strata on the Eurasian slope of the Lomonosov Ridge. In 2004, IODP's ACEX expedition drilled the first holes at these high latitudes (88° N) on the Lomonosov Ridge (Backman et al., 2005). The drillings were located along seismic profile AWI-91090 (Jokat et al, 1992) (Fig. 1). This profile and a neighboring line AWI-91091, extend from the Amundsen Basin scarp across the Ridge to the slope towards the Makarov

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**Figure 6.** Combined time section of the TRA-1991 and NP-28 profiles, with similar space step for correlation of reflections from the continental slope to the Podvodnikov Basin. Location of profiles shown on Figure 1.
Basin. Thus, on this part of the Lomonosov Ridge, it has been possible to identify the sedimentary succession on the ridge and compare directly with the seismic reflections in the profile. It has also been possible to compare the AWI-91091 profile with the results from the Lomonosov Ridge profile described here at about 84° N and other Lomonosov profiles (Langinen et al., 2004). On line AWI-91090, the ACEX expedition drilled approximately 430 m through the flat-lying reflections on the Lomonosov Ridge (Backman et al., 2005) and penetrated the Cenozoic (Paleocene)-Mesozoic boundary near the base of the hole at 425 m. The reflection profile clearly shows the reflective Cenozoic strata to lie with marked unconformity on underlying Mesozoic (Cretaceous) strata. At the ACEX drill-hole location, the hole penetrated a first prominent reflection at 200 m, which proved to coincide with another unconformity, separating Early Miocene from Early Eocene strata. This hiatus of c. 20 million years marks a parallel unconformity that apparently transgresses the entire Ridge at this latitude.

The Lomonosov profile described here near 84° N is similar to that along AWI-91090, but the sedimentary succession above the angular unconformity (base of composite reflection "A") is about two and a half times thicker. This thickness agrees well with that obtained by Jokat (2005) along a neighboring profile at this latitude - AWI-985500.

The AWI-91090 and AWI-91091 profiles, referred to above, extend eastwards across the Lomonosov Ridge to the slope into the Makarov Basin and reach a location close to the end of the similarly-oriented Makarov profile, described here. AWI-91091 shows a prograding succession dipping into the Makarov Basin, overlain by an angular unconformity that can be correlated westwards to the Paleocene-Cretaceous boundary, described at the ACEX drill-hole on the Ridge (Backman et al., 2005). In the Makarov profile, described here, this unconformity is located at the base of the "A" reflection, which transgresses the underlying "basin-ridge-basin" structure referred to above. This basement ridge, with higher Vp velocities, is located down-strike south from the prominent linear basement feature of the Marvin Spur. The latter, further north, is apparently a slice of basement rifted off the Lomonosov Ridge (Langinen et al., 2004, Cochran et al. 2005). It can therefore be concluded that the basement ridge beneath this part of the Makarov Basin is probably Lomonosov-derived basement, flanked by Mesozoic sediments (Unit III) that were gently folded and faulted prior to erosion and early Cenozoic deposition.

The Cenozoic section (Units I and II) in the Makarov Basin is thicker than that on the Lomonosov Ridge and contains more prominent reflections,
suggesting that this succession may be more complete than that on the Ridge. The velocity Vp of the strata immediately below the basal Cenozoic unconformity (reflection "A") is lower (about 2.8 km/s) than normal below the inferred Cenozoic succession, suggesting significant variation in sedimentary facies and/or compaction of the Late Mesozoic strata.

From the Makarov Basin southwards to the Podvodnikov Basin, Units I and II have been traced via the NP-28 ice-station profile and TRA(b)-89-90. Over the Arlis Rise, Unit I retains its thickness, whilst Unit II is much reduced. Figure 6 shows that strata between the reflection "A" (inferred to be the base of the Cenozoic section) and the acoustic basement ("Ar") thickens away from the Arlis Rise, particularly southwards in the direction of the East Siberian Margin; indeed all units thicken in this direction. The composition of the acoustic basement beneath the Rise has been inferred, from magnetic data (Taylor 1981, Kovaks et al. 1999, Lebedeva-Ivanova and Gee, 2005), to be mafic igneous rocks related to Mesozoic sea-floor spreading. This hypothesis can be accommodated within the stratigraphic context proposed here, with late Mesozoic strata overlying the basement. The hummocky nature of the acoustic basement (Sorokin et al., 1999 and Fig. 6 herein) in the vicinity of the Rise could support this interpretation.

The southern part of the Podvodnikov Basin, separating the East Siberian Margin from the Arlis Rise, shows typical slope-rise geometry. The sedimentary succession of the East Siberian Margin has been described by Sekretov (2001), based on the geology exposed in the vicinity of the New Siberian Archipelago and on the De Long Islands; he inferred the existence of a nearly ten kilometer thick and complete Cenozoic and late Mesozoic (Cretaceous) succession. Underlying basement here is dominated by deformed Paleozoic and Early-Mid Mesozoic rocks, as inferred from the geology of the islands on the De Long Plateau.

CONCLUDING REMARKS

Controlling the velocity structure of the sedimentary cover in the Arctic Basin has been an essential prerequisite for interpretation of the deep structure of the Basin based on wide-angle refraction profiles. Four type areas were therefore selected for special investigation of the velocity structure of the sedimentary successions: the Lomonosov Ridge, the Makarov Basin, the Arlis Rise, and the continental slope of the East Siberian margin. The reflection seismic data from these four areas, presented and analyzed here, taken together with previous Russian and subsequent western acquisition, provide new evidence for the Cenozoic development of the Eurasian Basin and the Cenozoic-Mesozoic (and perhaps older) development of the Amerasian Basin.

In the four areas selected for special study, three main stratigraphic units have been recognized – I, II and III. The interval Vp velocity of Unit I is, in general, 1.7 – 1.9 km/s, that of Unit II, 2.0 to 3.0 km/s and Unit III, 3.0 – 4.0 km/s.

A major composite reflection "A" is widespread in the Amerasian Basin and on the Lomonosov Ridge, where, at the ACEX 2004 drill-site (Backman et al., 2005), it marks the base of the Cenozoic succession. An overlying prominent reflection marks a parallel unconformity separating Neogene (Miocene) from Paleogene (Eocene) sediments on the Ridge and, apparently, throughout the Makarov and Podvodnikov basins.

A prominent structural high beneath the Cenozoic sediments of the Makarov Basin appears to be a southerly continuation of the Marvin Spur, i.e. a fragment of continental crust rifted off Lomonosova (subsequently, today's Lomonosov Ridge) in the Mesozoic. Similar fragments of continental crust may exist beneath the Podvodnikov Basin and Alpha and Mendeleev ridges.

Towards the East Siberian shelf, the Cenozoic and Late Mesozoic successions thicken to two or three times those in the Podvodnikov Basin. Most of the Cenozoic part of this succession is thought to be dominated by turbidities, implying that the Cenozoic successions out in the Amerasian Basin are probably composed of distal turbidities and other very fine grained siliciclastics.

This account of the character of the sedimentary cover in the Arctic Basin supports previous interpretations of the structure and stratigraphy; it provides a foundation for further work, particularly piston-coring and drilling, to unambiguously define the nature of these successions and the Mesozoic-Cenozoic tectonic evolution of the Basin.

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DEVELOPING OUTER CONTINENTAL SHELF LIMITS IN THE ARCTIC OCEAN: GEOSCIENCE ENCOUNTERS UNCLOS ARTICLE 76

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ABSTRACT

Article 76 of the United Nations Convention on the Law of the Sea (UNCLOS) entitles a qualified coastal state to exercise certain sovereign rights beyond the usual 200 nautical mile limit, in a zone known generally as the outer continental shelf. To qualify, a coastal state must demonstrate that it possesses an authentic wide margin that satisfies the bathymetric and geological criteria defined in the Article. These criteria touch upon the depth and shape of the seabed and the thickness of underlying sediment, and are used to determine whether submarine elevations adjacent to the continental margin are “natural prolongations” of a coastal state’s land mass. While straightforward in principle, the concept can be difficult to apply in practice on account of the Article’s simplifying assumptions and ambiguous language.

This paper discusses key sources of ambiguity in Article 76, and considers how the outcome of a given implementation depends on the data bases that are available, and on the methodologies applied in their interpretation. It reviews and compares two recent implementations in the Arctic Ocean - one performed for academic purposes using public-domain data bases, the other developed as part of a formal continental shelf submission using classified data bases. The latter implementation was subjected to a review mandated by UNCLOS, and while only partially disclosed, the mixed outcome of that process suggests strategies that other Arctic states might consider adopting as they develop their own submissions.

INTRODUCTION

The Arctic Ocean is a semi-enclosed sea surrounded by the land masses of Canada, Greenland, Norway, Russia, and the USA (Fig. 1). In regions adjacent to the continental margins of all five coastal states, the seabed has characteristics that may, in

Figure 1. The Arctic Ocean, showing surrounding coastal states, their combined 200 nautical mile limits, and submarine elevations that could figure in the determination of the outer limit of the juridical continental shelf, according to the provisions of UNCLOS Article 76: Chukchi Cap, Alpha-Mendeleev Ridge, Lomonosov Ridge, Morris Jesup Plateau, and Yermak Plateau.
accordance with the provisions of Article 76 of the United Nations Convention on the Law of the Sea (UNCLOS), provide validation for the exercise of certain sovereign rights beyond the usual 200 nautical mile (nm) limit. The area in which these sovereign rights apply is known as the juridical continental shelf, which is not to be confused with the physiographic continental shelf.

Article 76 defines the bathymetric and geological criteria that a coastal state must satisfy in order to project elements of its national jurisdiction beyond 200 nm, and to define the outer limit of that projected jurisdiction. In general, this entails the collection and analysis of observations that describe the depth and shape of the seabed, as well as the thickness of underlying sediment. The outer limit that is so determined, along with supporting information, must then be documented in a submission that is presented to the Commission on the Limits of the Continental Shelf (CLCS), an elected body of 21 experts in the field of geology, geophysics, or hydrography; this must occur within ten years of the entry into force of UNCLOS for that particular state. The primary function of the CLCS is to review the contents of the submission, and to issue recommendations concerning the proposed outer limit.

If its continental shelf submission attracts no objections from the CLCS, a coastal state may begin to exercise significant sovereign rights within the extended region: jurisdiction over living and non-living resources of the seabed and subsoil; control over the emplacement and use of submarine cables and pipelines, artificial islands, installations, and structures; regulation of drilling; control and prevention of marine pollution; and regulation of marine scientific research. Article 76 is therefore a piece of international maritime law that has significant relevance for coastal states that qualify.

CRITERIA AND UNCERTAINTIES IN THE IMPLEMENTATION OF ARTICLE 76

To succeed in projecting sovereign rights beyond 200 nm according to the provisions of Article 76, a coastal state must satisfy several technical criteria that are outlined below, and which engender different classes of uncertainties.

Test of appurtenance. Characteristics of the seabed in the prospective outer continental shelf must comply with the specifications defined in Article 76. This involves a consideration of seafloor morphology for the purpose of assessing whether submarine elevations in the zone indeed comprise a “submerged prolongation of the land mass of the coastal state” (Paragraph 3 of Article 76). This question is potentially fraught with controversy, because in some situations there may be no clear-cut indication of the geological character and structure of the submarine elevation in question, nor of the nature of its connection to the land mass. Under these circumstances, the answer may derive from a potentially contentious interpretation of incomplete information.

Morphology. Successive locations of the foot of the continental slope, defined as “the point of maximum change in the gradient at its base” (Paragraph 4b of Article 76) must be determined with sufficient accuracy and appropriate spacing. These locations define points of reference for subsequent applications of the two formula lines of Article 76, i.e. the distance and sediment thickness rules. The location of the foot of slope can vary according to the data base that is used in the analysis, in the analytical techniques that are applied, and in the investigator’s degree of
Figure 3. The sediment thickness formula of Article 76, in principle and in practice. The upper diagram presents the simple model upon which the formula is based. The lower diagram presents a typical sediment profile off the Canadian Atlantic margin, where the ruggedness of the crystalline basement contrasts with the smooth oceanic crust of the conceptual model. This portends significant difficulties in locating unambiguously the point where sediment thickness is equal to one percent of the distance to the foot of slope. Rigour in their application (Fig. 2).

Sediment thickness. The location of points where sediment thickness is equal to one percent of the distance to the foot of slope must be determined with reasonable accuracy and appropriate spacing. In deep water where the sediment thickness formula is most likely to apply, seismic observations may be scarce or poorly distributed. Barraging the mobilization of a costly data gathering program, there may be a tendency to make do with existing data sets and to try building a case on reasonable assumptions - which may or may not meet with the approval of the CLCS. Even with a complete and well-distributed data set, inherent errors in seismic interpretation tend to hover around the ten percent level. Moreover, the analysis may be compromised by basement irregularities (Fig. 3) or by difficulties in distinguishing sedimentary from non-sedimentary material.

Bathymetry. In some situations, the 2500 meter (m) isobath must be determined with reasonable accuracy and continuity, to provide a reference for developing one of the cutoff or constraint lines of Article 76, i.e. a line drawn 100 nm seaward of that particular isobath. (The second cutoff or constraint line is an envelope of arcs constructed with a radius of 350 nm, and centred on the coastal state’s territorial sea baseline.) In spite of modern technological advances, the precise measurement of water depth remains a somewhat inexact science, with the highest accuracy estimated to be in the vicinity of two percent. Compounding the measurement aspect, there may be more than one 2500 m isobath to choose from, and the final selection could raise objections from the CLCS.

A COMPARISON OF RECENT ARCTIC IMPLEMENTATIONS

To date, there have been two independent implementations of Article 76 in the central Arctic Ocean: an academic investigation for the whole of the central basin, using public data bases (Macnab et al., 2001); and a formal submission by the Russian Federation for a continental shelf extension off that country’s Arctic margin, using data bases that remain classified for the most part (United Nations, 2002a). These are described briefly in the paragraphs that follow.

Academic investigation. This work was based on information extracted from two primary sources: the International Bathymetric Chart of the Arctic Ocean (Jakobsson et al., 2000), and a sedimentary thickness map (Jackson and Oakey, 1986). The authors of this study assumed that the Alpha-Mendeleev and Lomonosov Ridges constituted legitimate prolongations of the land masses of the adjacent coastal states. The derived outer continental shelf limit is portrayed in Figure 4, which shows that the combined shelves of the five Arctic coastal states occupy most of the Arctic Ocean, except for two “donut holes” where sovereign rights cannot be extended: a small trapezoidal zone in Canada Basin, and a larger elongated zone that encloses the Gakkel Ridge.

Russian submission. In developing their continental shelf limit, Russian investigators had access to original observations that were collected by their national agencies during numerous scientific and mapping expeditions over the past half-century. The existence of these observations has been publicly acknowledged...
Figure 4. Results derived by Macnab et al (2001), assuming that the Alpha-Mendeleev and Lomonosov Ridges constitute legitimate prolongations of the land masses of the surrounding coastal states. The combined continental shelves of these states occupy most of the Arctic Ocean, except for two 'donut holes'. The smaller opening is bounded by segments of the 350 nm limit and the 2500 m isobath plus 100 nm; these are the outer limits of Canada, Russia, and the USA. The larger opening is bounded by segments of the 200 and 350 nm limits, the 2500 m isobath plus 100 nm, and the lines constructed in accordance with the distance and sediment thickness formulae; these are the outer limits of Greenland, Norway, and Russia.

(HDNO, 1996; HDNO et al, 1999), however their distribution and contents remain classified because they were collected for defence purposes. An exception is the series of geotransects that were executed specifically for Article 76 purposes, and which yielded deep crustal profiles across seabed features that were deemed critical to the determination of the outer limit (Fig. 5). As was done in the academic investigation described previously, analysts considered that the Mendeleev and Lomonosov Ridges were prolongations of Russia’s land mass (Poselov et al, this volume). The derived outer limit is shown in Figure 6, where the most noticeable enclosure consists of a roughly triangular zone with its apex at the North Pole, and its base defined by the 200 nm limit north of eastern Siberia.

From an examination of Figures 4 and 6, it can be seen that the outer shelf limit illustrated in each diagram skirts the Arctic Mid Ocean Ridge, known also as the Gakkel Ridge. However a detailed comparison (Fig. 7) reveals that the two limits diverge appreciably in some places. Two possible explanations for this discrepancy are: (1) the use of different data sets for describing depth and sediment thickness; and (2) the application of different techniques and criteria for analysing and interpreting this information.

As explained earlier, the results shown in Figure 4 were derived from public-domain data sets, while the results of Figure 6 were based on mainly classified holdings of depth and sediment thickness observations. Without access to the original observations, it is not possible to rule out differences between the public and classified models that could generate significant divergences in the locations of parameters that are fundamental to the implementation of Article 76, i.e. the foot of the slope, the 2500 m isobath, and the line where sediment thickness is everywhere equal to one percent of the distance to the foot of the slope.

At the same time, it is conceivable that the different outer limits of Figures 4 and 6 could have resulted from variations in the interpretive styles and techniques that were employed by the two investigative teams, not to mention the levels of conservatism in their approaches. For instance and as illustrated at the beginning of this paper, determining the location of the foot of the slope is an exercise that offers considerable scope for subjective decisions on the part of the investigator, especially in situations that involve a choice between two or more plausible locations. The same could be said for the 2500 m isobath under certain circumstances. To further complicate the process, determining the one percent line of sediment thickness may offer a significant range of options, depending on one’s interpretation of where the base of sediment lies.

**ISSUES RAISED BY THE RUSSIAN SUBMISSION**

The Russian submission aroused concerns both within and without the CLCS. The deliberations of the
CLCS and its ensuing recommendations are not open to public scrutiny, but it has been reported unofficially that the CLCS considered there was insufficient justification for characterizing the Mendeleev and Lomonosov Ridges as natural prolongations of the Russian land mass. Hence it was recommended that Russia present a revised submission (United Nations, 2002b). This was consistent with the stance articulated in the Guidelines of the CLCS (United Nations, 1999), where it is stated that ridges need to be examined on a case-by-case basis, and with knowledge of the relevant details.

Complementing the concerns of the CLCS, three coastal states responded specifically to the submission’s central Arctic component. Canada and Denmark (the latter acting on behalf of Greenland) both indicated a need for additional supporting data in order to assess properly the submission, and advised that the recommendations of the CLCS would be without prejudice to the prospective delimitation of the continental shelf between themselves and the Russian Federation. The USA, meanwhile, disagreed strongly with the submission, claiming that it had “major flaws” and questioning whether its geological criteria and interpretations were “accepted as valid by the weight of informed scientific opinion” (United Nations, 2002c).

These or similar issues are likely to attach themselves to submissions presented by other Arctic states, therefore it would be appropriate for affected parties to assess the validity of the objections that have been raised (by the CLCS in particular) concerning the Russian submission. Then it should be possible to make an informed decision on how to circumvent the objections - either by refuting them, or by taking them into account when constructing new outer limits.

**SUGGESTED STRATEGIES FOR DEALING WITH THE ISSUES RAISED**

In the present geographic setting, the key requirement of this process would be to consider the case for characterizing the Alpha-Mendeleev and Lomonosov Ridges as natural prolongations of the adjacent landmasses. Initially, this would entail the assembly and analysis of all available data sets that could shed light on the tectonic and geological framework of the seabed elevations in question, e.g. bathymetry, magnetic and gravity anomalies, and seismic reflection/refraction. In this context, it is conceivable that the “evidence to the contrary” provision of Paragraph 4(b) of Article 76 could be invoked as an alternative to the morphological definition of the foot of the slope, i.e. the “point of maximum change in gradient.”

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**Figure 5.** Geotransects executed by agencies of the Russian Federation to measure the depth of sedimentary material and crustal thickness across three features considered key to Russia’s implementation of Article 76: Delong Islands to Makarov Basin; Lomonosov Ridge; and Mendeleev Ridge.
Figure 6. Shaded areas show the locations of the continental shelf extensions sought by Russia in the Barents Sea and in the central Arctic Ocean. The double black line is a provisional outer limit of the Russian continental shelf, its final position subject to negotiation with neighbour states. Other components seen in this figure represent elements that figures in the development of the Russian claim. Source: website of the UN Division of Ocean Affairs and the Law of the Sea (DOALOS).

Significant holdings of data sets from the Arctic remain classified, and so are unavailable to the international scientific community. However, four separate compilations (Figs. 8a-8d) have assembled and rationalized thematic, public-domain information for the Arctic region (Kenyon and Forsberg, 2001; Jackson and Oakey, 1986; Jakobsson et al, 2000; Verhoef et al, 1996). These provide a knowledge base for framework investigations, which could identify specific problems that required follow-up action in the form of targeted expeditions for collecting new data needed to resolve ambiguities in the existing data sets.

A further stage in the analysis would involve studies that clarified the tectonic evolution and framework of the central basin of the Arctic Ocean. This would require the development of models that treated the Lomonosov and Alpha-Mendeleev Ridges as double-ended natural prolongations, and which related their origins, natures, and positions to the two main opening episodes of the Arctic Ocean: (1) the lateral opening of the Eurasia Basin, with spreading centred on the Gakkel Ridge; (2) the presumed rotational opening of the Amerasia Basin, possibly accompanied by deformation or accretion in the adjacent land masses, and with one or more poles of rotation situated in North America. Opening scenarios for these two Basins and their potential implications will be discussed briefly in the paragraphs that follow.

Lomonosov Ridge. This feature is considered to be a sliver of continental crust that separated from the Barents-Kara margin during the opening of the Eurasia Basin. Morphologically, the Ridge’s extremities appear to be partially attached to the Siberian and Ellesmere-Greenland margins, suggesting earlier linkages that were moderately disrupted during the opening of the Basin. Therefore one would expect the Ridge to retain a strong geological affinity to the
adjacent continental margins, which would support its characterization as a natural prolongation. This scenario needs to be confirmed with a plate reconstruction, which should be relatively straightforward in this region on account of its uncomplicated spreading geometry (Fig. 9).

Alpha-Mendeleev Ridge. The nature and origin of this feature remain enigmatic, however judging from the character of the overlying magnetic field, the complex appears to have a volcanic origin. A mantle hotspot has been suggested as the most likely source of this volcanism. If a stationary hotspot were located beneath the spreading centre for the duration of the opening of the Amerasia Basin, it would conceivably cause a continuous extrusion of material through the upper crust: initially this material would penetrate the overlying plate, accreting to and modifying the edges of the spreading plate segments (i.e. the new continental margins). This would initiate the creation of the Alpha-Mendeleev Ridge, and as the end segments of this embryonic feature were drawn in opposite directions, new material would be injected into the central gap that was so created (Fig. 10). Performing a plate reconstruction for such a scenario would require more age data than is currently available, and the operation would likely be significantly more complicated than a reconstruction in Eurasia Basin. However, the outcome might provide a plausible explanation for the origin and formation of the Alpha-Mendeleev Ridge. If it also supported the notion that their formation was coeval with the initial breakup, the present end segments of the Ridge could then be perceived as “natural prolongations” of the adjacent continental margins.

CONCLUSIONS
Two significant lessons may be drawn from a consideration of the Russian submission and of the circumstances that surround it: (1) the locations of proposed outer continental shelf limits will vary according to the data bases, methodologies, and criteria used in their development; (2) there may be no simple or predictable outcome to the characterization of submarine ridges as “natural prolongations” because the CLCS deals with ridges on a case-by-case basis. These lessons apply to other continental shelf submissions, not only in the Arctic but in several other regions as well.

These lessons suggest five key points of a general strategy that coastal states could adopt to ensure success in the development of their continental shelf submissions, particularly where there were prospects of overlapping continental shelf claims between adjacent states:

Pursue regional cooperation with high levels of transparency. It is preferable that contending parties address overlap issues and achieve some degree of accommodation at the beginning of the delimitation process, rather than afterwards when positions may be fixed and less amenable to change.
Consolidate and rationalize existing databases. Where regional bathymetry and sediment thickness are described in an array of fragmented, incomplete, or contradictory data sets, it is desirable to assemble and reconcile all available holdings in order to develop a complete, common, and coherent perspective of the seabed conditions that relate to the implementation of Article 76.

Harmonize methodologies and criteria. To avoid diverse and at times contrasting interpretations that can arise from the application of dissimilar techniques and criteria to the analysis of a given data base, it is recommended that standard tools and specifications be adopted for handling and treating data.

Address collectively the ridge concerns of the CLCS. In anticipation of a critical assessment by the CLCS, it is essential to develop a robust justification for the inclusion of natural prolongations and ridges in a continental shelf submission.

Coordinate submissions, scientifically and strategically. Neighbour states may achieve substantial economies of effort, not to mention consistent treatment from the CLCS, if they agree to share information and to work closely together in developing their continental shelf submissions.

ACKNOWLEDGEMENTS AND DISCLAIMER
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Figure 9. A simplified model for a transverse opening of the Eurasia Basin, indicating zones of attachment that connect the ends of Lomonosov Ridge to the adjacent continental margins, and which might have undergone moderate distortion during seafloor spreading.

(DOALOS), and from numerous colleagues and informed parties; errors of understanding or interpretation in this respect are the author’s alone. The results and conclusions presented here do not represent the official views of any government or organization, or even necessarily the final views of the author.

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Figure 10. A hypothetical model for a rotational opening of the Amerasia Basin, indicating zones where the end segments of the Alpha-Mendeleev Ridge, emplaced at the onset of seafloor spreading by a stationary hotspot, could have accreted to the adjacent continental margins.


ABSTRACT
Seismic studies of the Russian sector of the Arctic Ocean make it possible to use the geological criteria of Article 76 of the UN Convention on the Law of the Sea in order to extend the continental shelf of the Russian Federation beyond the 200 nautical mile. The sedimentary cover was investigated by means of a systematic seismic reflection surveys that form a 25×25km grid supplemented by data from ice drifting stations. The deep structures of the main Arctic submarine rises were explored by geophysical (DSS as a main method) and geological (including sampling) methods along three profiles with a total length of 2300 km.

On the basis of DSS data together with seismic reflection, bathymetric and potential fields data, crustal models of the Lomonosov Ridge (LR) and Mendeleev Rise (MR) were constructed. In both cases, the data support models involving thinned continental crust. This modeling combined with sampling data and the results of geomorphological analysis show that the LR and MR are to be considered elevations of continental origin. Based on these results the outer limits of the continental shelf beyond 200 nautical miles of the Russian Federation’s continental shelf in the Arctic Ocean have been projected.

INTRODUCTION
The Russian Federation is the first coastal State to make a submission to the Commission on Limits of
Continental shelf regarding the outer limits or its continental shelf in accordance with Article 76 of the UN Convention on the Law of the Sea. The submission was made in December 2001 and included, amongst other areas, the presented continental shelf of the Russian Federation beyond 200 nautical miles in the Arctic Ocean.

THE DATA BASE

_Bathymetry investigations._ Systematic studies of the bottom relief of the Arctic basin began during the 1960ies. During the time from 1962 to 1988 echo sounding surveys and seismic reflection surveys were carried out by the Russian scientists in a pre-defined grid covering about 80% of the deep-water area of the basin. The investigations were conducted by Ministry of Defense “North” expeditions. At the same time bathymetric acquisition was carried out also by USSR nuclear submarines.

The sounding density varied with depth; on bathymetric highs the sounding density was one point per 5 km, while in deep waters and abyssal plains the distance between depth points was increased to 15 kilometers or more. In this way, it was possible to characterise the entire surveyed area by a bathymetric grid of 25 x 25 km (locally much better). A bottom relief map of the Arctic Ocean at a scale of 1:5,000,000 scale was published by Russian authorities in 1999. In 2000 the International Bathymetric Chart of Arctic Ocean (IBCAO) based on a of 2.5x2.5 km grid was compiled and published in an international cooperation project involving scientists from circum Arctic countries (Jacobsson et al., 2000), and which parts of the Russian survey data were included. Recently, Russian authorities published an updated and correct bathymetric map Central Arctic Basin in 1:2,500,000 scale (HDNO, 2002, S-Petersburg).

Morphological analyses, accomplished by Russian scientists, have shown that the large-scale terraces that average up to more than 100 sq. km, complicate the Arctic basin continental slope. Seventy nine bathymetric profiles were constructed normal to the continental slope at a 60 nautical miles spacing in order to delineate the foot of the continental slope and the 2,500 meters isobath in accordance with Article 76. The exact location of the foot of the continental slope points were calculated mathematically based on the geomorphology.

_Seismic investigations._ The acquisition of the seismic reflection, including deep seismic sounding (DSS) data, have been, and still is, the main geophysical methods to study the Earth crustal structure of the Arctic. Russian seismic reflection and refraction studies have been carried out mainly as systematic areal and/or line on-ice surveys and presently amount to a data base far exceeding those acquired high-latitude expeditions of other countries.

Seismic reflection measurements made from drifting ice stations NP (North Pole) - 21 (1973), NP 22 (1975-81), NP 23 (1978), NP 24 (1979-80), NP 26 (1983-85) NP 28 (1987-88) and NP31 (1988-89) add up to 14,400 line km of profile consisting of 35,600 individual seismic shots. The long-term Trans-Arctic program carried out during several summer seasons (1989-1992, 2000) provided a total 2,300 line km of deep crustal profiling in three different transects (two sub-latitude and one sub-longitudinal direction). In the SLO 89-91 (De-Long Island– North Pole profiles) and SLO 92 (Lomonosov Ridge profile) the recorder spacing was generally 10 km, except across the Mendeleev Rise where the recorder spacing was 5 km. Signals were recorded to 200 km two-way travel distance.

Areas seismic reflection surveys enabled the compilation of 20,900 km of composite seismic profiles documenting the sedimentary fill in a large part of the Arctic basin.

Seismic acquisition by other countries was performed mainly during 1950-1983 on year-round drifting stations (T-3, Alpha, Charlie, Arlys-II) and from seasonal ice stations (Lorex-79, Fram-I, - Fram-II, - Fram-III, - Fram-IV, Cesar-83). A total of 4,000 km of high-resolution seismic, 1,800 km of single reflection soundings, and 3,000 km of refraction profiles were acquired in this way. More recent important contributions of seismic reflection 2D data has been made through multinational expeditions in 1991, 1995, 1998 and 2001 utilising icebreakers _Polarstern_ (Germany), _Oden_ (Sweden), and _Healy_ (USA); amounting to almost 4,500 line km of seismic reflection data supplemented by about 100 sonobuoy measurements (Fig.1).

A major part of the deepwater Arctic Basin is covered by Russian aeromagnetic surveys with the total flight track length in excess of 2 million km. The shelves of the Barents, Laptev and south-western Kara Seas and their bounding archipelagos were studied by relatively dense coverage with line spacings of 5-10 km, rarely up to 20 km. In the deep Arctic Ocean, comparatively detailed surveys were performed over the Gakkel Ridge and within the De Long Islands – North Pole and Lomonosov Ridge geotransects. The shelves of the East Arctic seas and the zones of their transition into adjacent deep seabed, however, are quite insufficiently surveyed; the Russian flight tracks here spaced at 20-40 km or more and these flights also had large positioning errors in the order of 10-20 km (or...
Figure 2. Time section along NP28 drift line

Figure 3. Depth section along A03 seismic composite profile
Figure 4. Seismotomographic model of the lithosphere along the “TRA 89-91” geotransect with horizontal/vertical ratio ~ 10

Figure 5. Seismotomographic model of the lithosphere along the “TRA 92 - 2000” geotransect with horizontal/vertical ratio ~ 7.5
even greater) due to manual navigation method of the 1960-ies.

More than a half of the Arctic Ocean on the Greenland/North-American side and the respective continental margins are covered by more contemporary U.S. aeromagnetic surveys that were flown at a much higher precision level et al. (Brozena et al., 2003).

DISCUSSION

There are two geological criteria specified in Article 76 for the determination of the outer limit of the extended continental shelf of coastal States. The first one (paragraph 76.4(a)(i)) provides for establishing the outer limit based on the thickness of the sedimentary cover, and the second (paragraphs 76.3 and 76.6) provides for establishing the outer limits on the basis of submarine rises (elevations) that are natural prolongations (components) of the continental margin.

The seismic studies of the Russian sector of the Arctic Ocean support the use of both criteria for delineating the outer limits of the continental shelf of Russia beyond the 200 nautical miles.

The top of the acoustic basement that as determined on the composite sections (constructed from areal coverage single shots) has been confirmed by seismic reflection data collected along NP drift lines (Fig. 2) and seismic reflection lines of “Polarstern -91, -98”. All available velocity information was used for time-depth conversion of the composite seismic sections.

The formula line at 60 nautical miles distance from the foot of the continental slope, and the points where the sediment thickness is 1% of the distance back to the foot slope were determined from the composite seismic profiles in order to delineate the outer limit of the Russian continental shelf beyond 200 nautical miles (Fig. 3).

The deep structure of the Earth’s crust of the main morphological of the Arctic Basin was investigated by a combination of geological and geophysical methods (the main method being DSS) along three trans-Arctic geotransects. Models of the Earth’s crust along the De Long Island - Makarov Basin and Lomonosov Ridge - Mendeleev Rise profiles were calculated based on the DSS data integrated with seismic reflection and bathymetric data.

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Figure 6. The specified Area of the extended Continental Shelf of the Russia beyond 200 NM Exclusive Economic Zone (EEZ) in the Arctic Ocean
The maximum thickness of crust beneath the De Long Islands was calculated as 43 km, in the Makarov Basin as 22 km, beneath the Lomonosov Ridge as 26 km, and beneath the Mendeleev Rise as 33 km. A two-layer model for crystalline crust was established (Fig. 4, 5). The thickness of the upper crust in the Lomonosov Ridge and in the Mendeleev Rise ranges is comparable, i.e. up to 7 and 9 km, respectively. The thickness of the lower crust beneath in the Mendeleev Rise is up to 18 km, by far exceeding that of the upper crust. This corresponds to a model of thinned continental crust.

The velocity structure of the sedimentary cover of the Lomonosov Ridge is similar to that of the Alpha Ridge as determined from the sonobuoy data acquired by “Polarstern” (Jokat, 2003). A regional unconformity is covered by sediments with velocities of 2.1-2.3 km/s, and is underlain by two sediment units with velocities of 4.0-4.7 km/s and 5.0-5.4 km/s. It is assumed that the lowermost unit consists of consolidated sediments or carbonates.

The seismic cruises also included dredge sampling on the Mendeleev Rise. In our opinion, the rock samples recover from this dredging represents in situ materials and confirm the continental nature of the Mendeleev Rise. Paleontological evidence suggests a Paleozoic age for the sampled rocks similar to the platform formations recovered from Northwind Range (Grantz et al., 1998).

In summary, the DSS and seismic reflection data, the structural analysis of seabed morphology, the bottom sampling have shown that the main rises end elevation of the Amerasian Basin are of continental origin.

The key problem is the character of the junctions of the Lomonosov Ridge and Mendeleev Rise with their adjacent shelf. Some scientists believe that these are tectonic (Grantz et al., 1998). This is not confirmed by bathymetric and seismic data and must be regarded only a hypothesis.

According to the morphological analysis, the Mendeleev Rise and the Alpha Ridge are natural components of complicated continental margin. The rises and depressions of the Amerasian Basin exhibit terrace shaped bathymetric steps with bathymetric levels of 500-600 meters suggesting that the morphology of the region is mainly a result of the vertical movements.

Furthermore, no large earthquake epicenters have been in the area of the junction between the Lomonosov Ridge and the adjacent shelf. Thus, the Russian conception of outer limit of the extended continental shelf of the Russian Federation in the Arctic Basin is based on the interpretation that the Lomonosov Ridge, the Alpha Ridge and the Mendeleev Rise are natural submarine prolongations of the Ellesmere – West Siberian mainland continental margins.

**CONCLUSIONS**

Based on the results of an integrated interpretation of geological and geophysical data, a map of the extended continental shelf of the Russian Federation beyond 200 nautical miles in the Arctic Ocean has been constructed in accordance with the provisions of Article 76 of the UN Convention on the Law of the Sea (Fig. 6). The limit of the extended continental shelf is described by sections (Fig. 6) determined on the basis of various criteria specified in Article 76 of the Convention, as follow:

- **Section I.** This line will depend on the outcome of negotiations on the delimitation between the Russian Federation and Norway.
- **Sections II and IV.** The limit is delineated along the lines connecting fixed points where the thickness of sedimentary rocks exceeds 1% of the shortest distance from such points to the foot of the continental slope (paragraph 76.4 (a)(i)) and which do not extend beyond the constraint of 350 nautical miles from the baselines (paragraph 76.5).
- **Sections III and V.** The limit is delineated at a distance of 200 nautical miles from the coastal baselines.
- **Sections VI and VIII.** The limit is delineated along the lines connecting fixed points where the thickness of sedimentary rocks exceeds 1% of the shortest distance from such points to the foot of the continental slope (paragraph 76.4 (a)(i)) and which do not extend beyond the constraint of 100 nautical miles from the 2,500 m isobath (paragraph 76.5).
- **Sections VII and IX.** The limit is delineated at a distance of 60 nautical miles from the foot of the continental slope (paragraph 76.4 (a)(i)) and which do not extend beyond the distance constraint of 100 nautical miles from the 2,500 m isobath (paragraph 76.5).
- **Section X.** The limit is delineated along the line which will delimit the continental shelves of the Russian Federation and Canada (it will be defined in the course of bilateral negotiations) and along the line of delimitation between the Russian Federation and the USA.

The seabed of the Podvodnikov Basin is included in the juridical continental shelf of the Russian Federation by the criterion of the thickness of sedimentary cover (paragraph 76.4 (a)(i)).
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ABSTRACT
The Sverdrup Basin Magmatic Province (SBMP) is located in the east-central part of the Sverdrup Basin of the Canadian Arctic, and consists of hypabyssal intrusive sheets and dykes, flood basalts, and volcanoes that were emplaced episodically from the Early Cretaceous to the Paleogene. Igneous rocks belonging to the SBMP have previously been dated by primarily by K-Ar, with ages ranging from 129 ± 2 Ma to 59 ± 1 Ma, with a preponderance of ages in the Cenomanian to Turonian (99-89 Ma), suggestive of a magmatic peak. Seven new \(^{40}\)Ar\(^{-39}\)Ar dates of volcanic and intrusive rocks of the SBMP are presented. A dyke from Axel Heiberg Island gives an age of 128.2±2.1 Ma, while tholeiitic lava flows give two age groups with dates of 96.1±1.9 Ma, 92.3±1.1 Ma, 83.8±1.2 Ma and 80.7±1.1 Ma. This suggests at least three episodes of magmatic activity. The older ages correlate with dating from two sills of northern Ellesmere Island that give ages of 94.4±2.2 Ma and 97.2±1.4 Ma. The younger ages appear to conflict with biostratigraphically-determined Cenomanian-Turonian ages for overlying strata of the Kanguk Formation, highlighting the need for further investigation to solve this apparent discrepancy.

INTRODUCTION
The Sverdrup Basin Magmatic Province (SBMP) is located in the east-central part of the Canadian Arctic Archipelago, Nunavut, Canada (Fig. 1). The igneous province consists of hypabyssal intrusive sheets and dykes, flood basalts, and central volcanoes that were emplaced episodically from the Early Cretaceous to the Paleocene. Volcanic rocks mostly outcrop on Axel Heiberg Island and northern Ellesmere Island, and include flood basalts associated with voluminous, hypabyssal intrusive sheets; thin successions of ferrobasaltic lavas, and associated intrusive rocks; and bimodal volcanic successions of alkaline character containing pyroclastic deposits, and the associated plutons. Igneous rocks are predominantly basaltic (Williamson, 1988), a characteristic of large igneous provinces associated with continental break up (Coffin and Eldholm, 1992).

K-Ar and \(^{40}\)Ar\(^{-39}\)Ar geochronological data acquired since the mid 1970's (Table 1) suggest that the locus of volcanic activity shifted from areas close to the Mesozoic depocentre (Embry, 1991; Amund Ringnes Island, ca. 129 Ma) to the northern margin of the basin (ca. 92 Ma). Notably, the bulk of reported ages for igneous rocks from northern Ellesmere Island fall between the Cenomanian and Turonian (98.9-89.0 Ma), suggesting a peak of magma production and emplacement along the Sverdrup Rim (Meneley et al., 1975). Sporadic and volumetrically less significant magmatic activity took place throughout Early Cretaceous to the Paleocene (Embry and

Osadetz, 1988). Williamson (1988) proposed that age-progressive magmatism in the Sverdrup Basin, and a widening of the tectonic zone to the north were linked to the development of the adjacent rifted margin, and opening of the Arctic Ocean. In addition, the available data indicate that subalkaline basalts were the first and most voluminous magmas emplaced in the SBMP (Williamson, 1998). The recognition of older, rift-related structures in the western Arctic Islands (Balkwill and Fox, 1982) and of volcanism of Late Cretaceous age at the site of the Alpha Ridge (Van Wagoner et al., 1986; Muhe and Jokat, 1999) argue in favour of age-progressive igneous activity in the Sverdrup Basin.

The objective of our study is to test this hypothesis by examining the timing of emplacement of lava flows and associated intrusives in the SBMP using geochronology. Here, we present new $^{40}$Ar/$^{39}$Ar age data for seven samples of igneous rock collected by Williamson (1988) on Axel Heiberg Island, and northern Ellesmere Island (4 basaltic lava flows; 1 dyke; 2 sills). The sampling locations are shown in Figure 1, and keyed to the results listed in Table 1.

### METHODOLOGY

Selected samples were processed for $^{40}$Ar/$^{39}$Ar analysis of whole rock by standard preparation techniques, including hand-picking of unaltered pieces in the size range 0.25 to 0.50 mm. Individual mineral separates were loaded into aluminum foil packets along with a single grain of Fish Canyon Tuff Sanidine (FCT-SAN) to act as flux monitor (apparent age = 28.03 Ma; Renne et al., 1998). The sample packets were arranged radially inside an aluminum can. The samples were then irradiated for 12 hours at the research reactor of McMaster University in a fast neutron flux of approximately $3 \times 10^{16}$ neutrons/cm$^2$.

Laser $^{40}$Ar/$^{39}$Ar step-heating analysis was carried out at the Geological Survey of Canada laboratories in Ottawa, Ontario. Upon return from the reactor, samples were split into several aliquots and loaded into individual 1.5 mm-diameter holes in a copper planchet. The planchet was then placed in the extraction line and the system evacuated. Heating of individual sample aliquots in steps of increasing temperature was achieved using a Merchantek MIR10 10W CO$_2$ laser equipped with a 2 mm x 2 mm flat-field lens. The released Ar gas was cleaned over getters for ten minutes, and then analyzed isotopically using the secondary electron multiplier system of a VG3600 gas source mass spectrometer; details of data collection protocols can be found in Villeneuve and MacIntyre (1997) and Villeneuve et al. (2000). Error analysis on individual steps follows numerical error analysis routines outlined in Scaillet (2000); error analysis on grouped data follows algebraic methods of Roddick (1988).

Corrected argon isotopic data are listed in Table 2, and presented (Fig. 2 and 3) as spectra of gas release or on inverse-isochron plots (Roddick et al. 1980). Each gas-release spectrum plotted contains step-heating data from up to two aliquots, alternately shaded and normalized to the total volume of $^{36}$Ar released for each aliquot. Such plots provide a visual image of replicated heating profiles, evidence for Ar-loss in the low temperature steps, and the error and apparent age of each step.

Neutron flux gradients throughout the sample canister were evaluated by analyzing the sanidine flux monitors included with each sample packet and interpolating a linear fit against calculated J-factor and sample position. The error on individual J-factor values is conservatively estimated at $\pm 0.6\%$ (2$\sigma$). Because the error associated with the J-factor is systematic and not related to individual analyses, correction for this uncertainty is not applied until calculation of dates from isotopic correlation diagrams (Roddick, 1988). No evidence for excess $^{40}$Ar was observed in any of the samples and, therefore, all regressions are assumed to pass through the $^{40}$Ar/$^{36}$Ar value for atmospheric air (295.5). If there is no evidence for excess $^{40}$Ar the regressions are assumed to pass through the $^{40}$Ar/$^{36}$Ar value for atmospheric air (295.5) and are plotted on gas release spectra. Blank were measured prior and after each aliquot and levels vary between $^{40}$Ar=2.5-3.6x10$^{-7}$ nm, $^{39}$Ar=4.2-13.3x10$^{-9}$ nm, $^{38}$Ar=0.4-1.7x10$^{-9}$ nm, $^{37}$Ar=0.4-1.7x10$^{-9}$ nm, $^{36}$Ar=0.7-1.3x10$^{-9}$ nm, all at $\pm 20\%$ uncertainty.

Nucleogenic interference corrections are ($^{40}$Ar/$^{39}$Ar)$_{K}$=0.025±0.005, ($^{39}$Ar/$^{38}$Ar)$_{K}$=0.011±0.010, ($^{40}$Ar/$^{37}$Ar)$_{Ca}$=0.002±0.002, ($^{39}$Ar/$^{37}$Ar)$_{Ca}$=0.00068±0.00004, ($^{38}$Ar/$^{37}$Ar)$_{Ca}$=0.00003±0.00003, ($^{36}$Ar/$^{37}$Ar)$_{Ca}$=0.00028±0.00016. All errors are quoted at the 2$\sigma$ level of uncertainty.

### VOLCANIC ROCKS

The Strand Fiord Formation exposed on western and northern Axel Heiberg Island consists predominantly of ponded lava flows, lapilli tuffs (rare), and volcanic breccia (lahar; rare) reaching a maximum thickness of 800 metres (Ricketts et al., 1985; Williamson, 1988; MacRae et al., 1996). The lava
Table 1. Ages reported for igneous rocks of the Sverdrup Basin Magmatic Province, and new data from this study.

<table>
<thead>
<tr>
<th>Location</th>
<th>Rock Type</th>
<th>Method</th>
<th>Age (Ma)</th>
<th>Ref.</th>
</tr>
</thead>
<tbody>
<tr>
<td>Sabine Peninsula,</td>
<td>Dyke</td>
<td>K-Ar w.r.</td>
<td>123 ± 6</td>
<td>[1]</td>
</tr>
<tr>
<td>Melville Island</td>
<td>Sill</td>
<td>K-Ar w.r.</td>
<td>152 ± 6</td>
<td>[1]</td>
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<td>Devon Island, Grinnell Peninsula,</td>
<td>Dyke</td>
<td>K-Ar w.r.</td>
<td>131 ± 6</td>
<td>[1]</td>
</tr>
<tr>
<td></td>
<td>Dyke</td>
<td>K-Ar w.r.</td>
<td>112 ± 5</td>
<td>[2]</td>
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<tr>
<td>Axel Heiberg Island</td>
<td>Buchanan Lake</td>
<td>Sill</td>
<td>129 ± 2</td>
<td>[4]</td>
</tr>
<tr>
<td></td>
<td></td>
<td>40Ar/39Ar m.</td>
<td>126 ± 2</td>
<td>[4]</td>
</tr>
<tr>
<td></td>
<td></td>
<td>40Ar/39Ar w.r.</td>
<td>113 ± 6</td>
<td>[4]</td>
</tr>
<tr>
<td></td>
<td>Strand Fiord Formation</td>
<td>Tholeiitic lava flow</td>
<td>40Ar/39Ar w.r.</td>
<td>100 ± 2</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Tholeiitic lava flow</td>
<td>40Ar/39Ar w.r.</td>
<td>95.3 ± 2</td>
</tr>
<tr>
<td></td>
<td>Twisted Ridge, Strand Fiord (1)</td>
<td>Tholeiitic lava flow</td>
<td>40Ar/39Ar w.r.</td>
<td>80.7 ± 1.1</td>
</tr>
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<td>Bunde Fiord Section (2)</td>
<td>Tholeiitic lava flow</td>
<td>40Ar/39Ar w.r.</td>
<td>83.8 ± 1.2</td>
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<td>Artharber Creek Section (3)</td>
<td>Tholeiitic lava flow</td>
<td>40Ar/39Ar w.r.</td>
<td>96.1 ± 1.9*</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Tholeiitic lava flow</td>
<td>40Ar/39Ar w.r.</td>
<td>92.3 ± 1.1</td>
</tr>
<tr>
<td></td>
<td>Lightfoot River (4)</td>
<td>Dyke</td>
<td>40Ar/39Ar w.r.</td>
<td>128.2 ± 2.1</td>
</tr>
<tr>
<td>Northern Ellesmere Island</td>
<td>Tanquary Fiord (5)</td>
<td>Sill</td>
<td>40Ar/39Ar w.r.</td>
<td>94.4 ± 2.2*</td>
</tr>
<tr>
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<td>Hassel Formation</td>
<td>Ferrobasaltic lava flow</td>
<td>40Ar/39Ar w.r.</td>
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<td>Wooton Intrusion</td>
<td>Gabbro</td>
<td>U/Pb zircon</td>
<td>92 ± 1</td>
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<td></td>
<td>Marvin Peninsula</td>
<td>Quartz Diorite</td>
<td>K/Ar hb.</td>
<td>93 (avg)</td>
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<td>Artharber Creek Section (3)</td>
<td>Sill</td>
<td>40Ar/39Ar w.r.</td>
<td>91 ± 2*</td>
</tr>
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<td>Same unit re-dated</td>
<td>40Ar/39Ar w.r.</td>
<td>97.2 ± 1.4</td>
<td>[11]</td>
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<tr>
<td></td>
<td>Hansen Point Volcanics</td>
<td>Lava flows (5)</td>
<td>Rb-Sr w.r</td>
<td>80.3 ± 1.5</td>
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<td></td>
<td>Philips Inlet</td>
<td>Lava flow</td>
<td>40Ar/39Ar w.r.</td>
<td>65 ± 3</td>
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<td>Basaltic dyke</td>
<td>Basaltic dyke</td>
<td>40Ar/39Ar w.r.</td>
<td>59 ± 1</td>
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<tr>
<td></td>
<td>Offshore</td>
<td>Alkali basalt</td>
<td>40Ar/39Ar w.r.</td>
<td>82 ± 1</td>
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</tbody>
</table>

1 Numbers in brackets are keyed to Fig. 1
All 40Ar/39Ar ages corrected for reference monitor FCT sanidine =28.03 Ma (Renne et al., 1998) except @ which does not contain information related to monitor used. Ages reported in this study (bold type) represent plateau ages except for *, which are inverse isochron ages (Roddick et al., 1980). Uncertainties reported at 2σ.
w.r = whole rock; p = plagioclase; m = biotite; hb = hornblende
<table>
<thead>
<tr>
<th>Power</th>
<th>Volume ( \times 10^{11} \text{ cc} )</th>
<th>( ^{36}\text{Ar}/^{39}\text{Ar} )</th>
<th>( ^{37}\text{Ar}/^{39}\text{Ar} )</th>
<th>( ^{38}\text{Ar}/^{39}\text{Ar} )</th>
<th>( ^{40}\text{Ar}/^{39}\text{Ar} )</th>
<th>( % 40\text{Ar} )</th>
<th>( ^{40}\text{Ar}/^{39}\text{Ar} )</th>
<th>Apparent Age (Ma)</th>
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<tbody>
<tr>
<td><strong>AX83-60 Whole Rock; J=0.0288490 (Z7402)</strong></td>
<td></td>
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<td></td>
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<tr>
<td><strong>Aliquot: A</strong></td>
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<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
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<tr>
<td>2.8</td>
<td>0.872</td>
<td>0.0411±0.0036</td>
<td>1.466±0.030</td>
<td>0.210±0.013</td>
<td>22.585±0.299</td>
<td>54.4</td>
<td>10.300±0.793</td>
<td>16.8</td>
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<tr>
<td>3.0</td>
<td>0.7323</td>
<td>0.0252±0.0031</td>
<td>2.573±0.041</td>
<td>0.116±0.012</td>
<td>23.177±0.300</td>
<td>32.2</td>
<td>15.710±0.370</td>
<td>14.1</td>
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<td>3.5</td>
<td>0.5709</td>
<td>0.0169±0.0039</td>
<td>3.145±0.066</td>
<td>0.082±0.011</td>
<td>21.026±0.317</td>
<td>23.3</td>
<td>16.123±0.409</td>
<td>11.2</td>
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<td>4.6</td>
<td>1.4151</td>
<td>0.0186±0.0017</td>
<td>3.144±0.075</td>
<td>0.128±0.012</td>
<td>21.353±0.497</td>
<td>25.7</td>
<td>15.876±0.501</td>
<td>27.2</td>
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<tr>
<td>5.5</td>
<td>0.7838</td>
<td>0.0205±0.0029</td>
<td>6.422±0.117</td>
<td>0.115±0.012</td>
<td>21.670±0.367</td>
<td>27.8</td>
<td>15.650±0.478</td>
<td>15.1</td>
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<td>12</td>
<td>0.8252</td>
<td>0.0195±0.0034</td>
<td>12.956±0.200</td>
<td>0.077±0.011</td>
<td>21.916±0.334</td>
<td>26.1</td>
<td>16.188±0.934</td>
<td>15.9</td>
</tr>
</tbody>
</table>

**BND83-16 Whole Rock; J=0.0288300 (Z7403)**

**Aliquot: A**

| 2.8 | 0.7242 | 0.0432±0.0038 | 2.190±0.052 | 0.163±0.013 | 26.032±0.419 | 49.4 | 13.161±0.704 | 9.6 | 67.18±3.53 |
| 3.5 | 1.1396 | 0.0207±0.0021 | 3.063±0.056 | 0.094±0.011 | 22.327±0.397 | 27.2 | 16.262±0.405 | 15.1 | 82.66±2.86 |
| 4.6 | 1.1413 | 0.0161±0.0019 | 4.467±0.057 | 0.085±0.011 | 20.727±0.203 | 20.8 | 16.420±0.337 | 18.7 | 83.44±1.67 |
| 5.5 | 0.4429 | 0.0157±0.0059 | 11.414±0.287 | 0.091±0.012 | 21.489±0.609 | 20.4 | 16.123±0.409 | 11.0 | 82.02±2.49 |
| 12 | 0.7743 | 0.0121±0.0032 | 24.368±0.401 | 0.059±0.011 | 20.634±0.370 | 16.7 | 17.193±1.284 | 10.3 | 87.28±6.36 |

**AX85-75 Whole Rock; J=0.0288050 (Z7404)**

**Aliquot: A**

| 2.4 | 0.5717 | 0.0779±0.0059 | 1.320±0.023 | 0.238±0.016 | 43.489±0.660 | 54.3 | 19.884±0.825 | 5.2 | 100.48±4.06 |
| 2.8 | 0.6263 | 0.0405±0.0053 | 1.806±0.023 | 0.238±0.016 | 30.293±0.469 | 40.1 | 18.138±0.680 | 5.7 | 91.88±3.36 |
| 3.0 | 1.0266 | 0.0317±0.0032 | 2.751±0.037 | 0.145±0.011 | 27.556±0.301 | 34.2 | 16.125±0.410 | 9.3 | 91.81±2.02 |
| 3.5 | 0.8305 | 0.0280±0.0045 | 3.264±0.045 | 0.132±0.014 | 26.130±0.373 | 31.8 | 16.817±0.816 | 7.5 | 90.29±4.03 |
| 4.2 | 1.1994 | 0.0175±0.0029 | 4.472±0.046 | 0.099±0.011 | 23.524±0.250 | 21.9 | 18.373±0.500 | 7.0 | 93.04±2.47 |
| 5.5 | 0.9717 | 0.0106±0.0033 | 10.161±0.165 | 0.096±0.013 | 21.402±0.318 | 14.4 | 18.323±0.639 | 8.8 | 92.79±3.16 |
| 12 | 1.3858 | 0.0203±0.0026 | 27.474±0.231 | 0.143±0.011 | 23.943±0.240 | 25.0 | 17.949±1.430 | 12.5 | 90.94±7.07 |

**AX85-97 Whole Rock; J=0.0288760 (Z7405)**

**Aliquot: A**

| 2.8 | 1.4633 | 0.0530±0.0017 | 2.031±0.042 | 0.201±0.013 | 33.538±0.250 | 46.1 | 18.084±0.474 | 13.2 | 91.61±2.34 |
| 3.0 | 0.9433 | 0.0188±0.0020 | 3.755±0.077 | 0.106±0.012 | 23.469±0.368 | 21.4 | 18.454±0.549 | 8.5 | 93.43±2.71 |
| 5.5 | 1.4734 | 0.0158±0.0016 | 10.300±0.192 | 0.111±0.012 | 22.260±0.354 | 19.3 | 17.954±0.701 | 13.3 | 90.97±3.46 |
| 12 | 0.5664 | 0.0264±0.0050 | 27.429±0.499 | 0.137±0.015 | 24.666±0.506 | 28.3 | 17.692±1.960 | 5.1 | 89.67±9.69 |

**TABLE 2 Ar-Ar data**
### TABLE 2 Ar-Ar data (continued)

<table>
<thead>
<tr>
<th>Power(^a)</th>
<th>Volume 39 Ar x10(^{-1}) cc</th>
<th>36 Ar/39 Ar</th>
<th>37 Ar/39 Ar</th>
<th>38 Ar/39 Ar</th>
<th>40 Ar/39 Ar</th>
<th>(f_{39}^{39})</th>
<th>(% 40) Ar (\pm)</th>
<th>Apparent Age (Ma)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Aliquot: A</td>
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<td></td>
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<td></td>
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<tr>
<td>2.8</td>
<td>0.4295</td>
<td>2.2425±0.0062</td>
<td>3.516±0.100</td>
<td>0.743±0.022</td>
<td>84.528±1.778</td>
<td>81.161±1.648</td>
<td>10.2 81.70±8.18</td>
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<tr>
<td>3</td>
<td>0.5677</td>
<td>2.284±0.091</td>
<td>0.761±0.031</td>
<td>64.355±2.171</td>
<td>60.4 25.492±1.964</td>
<td>13.4 127.70±9.50</td>
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<tr>
<td>4</td>
<td>0.7939</td>
<td>1.578±0.034</td>
<td>0.793±0.020</td>
<td>35.402±0.643</td>
<td>28.25 48.3±9.76</td>
<td>17.2 127.66±4.72</td>
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<tr>
<td>3.7</td>
<td>1.4069</td>
<td>2.848±0.037</td>
<td>0.624±0.014</td>
<td>33.275±0.360</td>
<td>22.5 25.785±0.533</td>
<td>33.2 129.12±2.57</td>
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<tr>
<td>5.5</td>
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<td>0.034±0.0032</td>
<td>1.577±0.038</td>
<td>0.714±0.031</td>
<td>37.599±1.176</td>
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<td>7 126.52±7.71</td>
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<td>4.449±0.066</td>
<td>0.262±0.011</td>
<td>20.137±0.205</td>
<td>5.6  20.14±0.34</td>
<td>7.2 96.14±1.68</td>
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<td>0.004±0.0009</td>
<td>1.061±0.023</td>
<td>0.701±0.011</td>
<td>20.751±0.228</td>
<td>0.8  20.49±0.39</td>
<td>4.4 103.40±1.86</td>
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<td>73.1 15.746±2.43</td>
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<td>9.157±0.328</td>
<td>0.183±0.020</td>
<td>35.32±1.088</td>
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<td>0.258±0.014</td>
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<td>15.2 19.67±0.58</td>
<td>8.8 99.90±2.86</td>
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<td>17.5 18.64±1.12</td>
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<td>0.392±0.019</td>
<td>26.530±0.694</td>
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<td>30.4 18.45±2.71</td>
<td>5.3 93.85±13.47</td>
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**AX85-37 Whole Rock; J=.00287740 (Z7407)**

**EL84-218 Whole Rock; J=.00289380 (Z7408)**

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\(a\): As measured by laser in \(\%\) of full nominal power (10W)

\(b\): Fraction 39 Ar as percent of total run

\(c\): Errors are analytical only and do not reflect error in irradiation parameter J

\(d\): Nominal J, referenced to FCT-SAN=28.03 Ma (Renne et al., 1994)

All uncertainties quoted at 2\(\sigma\) level
flows are uniformly basaltic, and consist of non-
-channelized sheets of aa and pahoehoe flows that
display complex intraflow structures as a result of
ponding and quenching. The basaltic lavas typically
show grey, vesicular tops, suggesting rapid extrusion
rates. In some cases, the presence of invasive flows
and sedimentary interbeds indicate lacustrine or
vegetated terrestrial conditions between volcanic
eruptions.

A total of three samples from the Strand Fiord
Formation were submitted for $^{40}$Ar/$^{39}$Ar dating.
Sampling localities are shown on Fig. 1 (Loc.1-3), and
described in Williamson (1988). A sample from the
type section of the Strand Fiord Formation on western
Axel Heiberg Island (AX83-060; Twisted Ridge,
Loc.1) consists of massive basaltic lava, sampled from
the central portion of a succession of ten lava flows.
No sills or invasive flows were observed. This sample
shows early-step Ar loss, but a well-defined plateau
that gives an age of 80.7 ± 1.1 Ma (Fig. 2a). Another
sample from the Strand Fiord Formation at Celluloid
Creek, near Bunde Fiord (BND83-16; Loc. 2) consists
of a portion of massive basaltic lava sampled from
the second flow in a succession of eight lava flows. No
sills, invasive flows, or sedimentary interbeds were
observed at the sampling locality. This sample showed
a very low K content, requiring analysis of a second
split of sample from the same irradiation package.
Although the first split showed a poorly-defined plateau
because of the larger proportion of background
correction, the second split displayed a well-defined,
multi-step plateau. Combining the data from the two
splits gives an age of 83.8 ± 1.2 Ma (Fig. 2b), just
overlapping at error limits, the age obtained from the
portion of basaltic lava sampled at Locality 1.

Two samples of tholeiitic basalt from the
Artharber Creek section, along the south shore of
Bunde Fiord (Loc.3, Fig. 1), were collected near the
base and top of the volcanic succession (Flow #1,
AX85-75; Flow #12, AX85-97). The sample from the
base of the volcanic pile gives results that are typical of
all the analyzed samples with a well-defined plateau. A
replicate analysis was made of this material that gave
reproducible results. Combining data from both
analyses gives an age of 92.3 ± 1.1 Ma (Fig. 2c). A
sample from the top of the succession displays clear
evidence of an excess $^{40}$Ar component and gives an
apparent old age on standard gas-release spectrum (Fig.
2d). However, use of inverse isochron plots (Rodick
et al, 1980) allows for correction for the excess $^{40}$Ar
and resolution of an age of 96.1 ± 1.9 Ma (Fig. 2e) for
AX85-97. Given these results, it is reasonable to
postulate that the volcanic succession was rapidly
emplaced at ca. 91-98 Ma.

**INTRUSIVE ROCKS**

The oldest sample dated in this study belongs to
the Lightfoot River dyke swarm located on the
northeastern coast of Axel Heiberg Island (Loc. 4, Fig.
1). These intrusions were first described by McMillan
(in Fortier et al., 1963), and reported to cut all the
bedrock formations in this area. Williamson (1988)
mapped a total of ten dykes and five sills that occur
along a 25 km wide belt, and can be traced for up to 40
km on air photographs. The intrusions are
deochemically indistinguishable from the Strand Fiord
lava flows, and could represent feeders for the flood
basalts. The sample chosen for age dating consists of
the fine-grained chilled margin of a 60 metre wide dyke
(AX85-37). Results for this sample indicate Ar-loss in
the lowest temperature part of the spectrum, followed
by a rapid rise into a well-defined multi-step plateau
with an age of 128.2 ± 2.1 Ma (Fig. 3a). The sample
also displays a typical pattern of rising Ca/K
concomitant with rising heating temperature, indicative
of Ar degassing from multiple phases in the rock. The
consistent plateau irrespective of analysis temperature
is indicative of a robust crystallization age.

Two samples collected from intrusive rocks near
Tanquary Fiord and in the Piper Pass area of northern
Ellesmere Island were selected for Argon age dating
(Fig. 1). A sill located along the Macdonald River,
near the head of Tanquary Fiord, gives an age of 94.4 ±
2.2 Ma (Fig 3b, c) and evidence for excess $^{40}$Ar (EL84-
218; Loc. 5). This sill shows sharp, conformable
contacts with Triassic-Jurassic strata of the Sverdrup
Basin, and has the most extreme Ca/K values of any of
the samples dated.

Sills in the Piper Pass area vary in thickness
between 10 and 90 m, are uniformly massive,
columnar-jointed, and display a gradual coarsening of
the gabbroic mineral assemblage from the margins to
the centre of units. A sill from this area (EL84-50; Loc.
6) has a reported $^{40}$Ar/$^{39}$Ar age of 91 ± 2 Ma (Muecke
et al, 1990). Redating the same sample results in a
plateau age of 97.2 ± 1.4 Ma (Fig. 3d). The slightly
older age may be due to a variety of factors, such as
analysis of much smaller volume of material (and
therefore ability to more carefully choose unaltered
fragments) or use of a different reference age for flux
corrector.

The Tanquary Fiord sills are geochemically
indistinguishable from thin successions of ferrobasaltic
lavas at Lake Hazen, and intrusive complexes of the
Figure 2. Gas release and inverse isochron plots (for samples with evidence of excess $^{40}$Ar) from volcanic rocks of the SBMP. Steps used in calculation of age on gas release plots are marked by arrowed line above the steps. All errors are quoted and plotted at 2σ level of uncertainty.
Figure 3: Gas release and inverse isochron plots (for samples with evidence of excess $^{40}$Ar) from intrusive rocks of the SBMP. Steps used in calculation of age on gas release plots are marked by arrowed line above the steps. All errors are quoted and plotted at 2σ level of uncertainty.
Piper Pass area (Williamson, 1988). The age data summarized in Table 1 suggests emplacement of the volcanic-intrusive complex during the Cenomanian and Turonian.

CONCLUSIONS

New laser $^{40}$Ar/$^{39}$Ar geochronology for volcanic and intrusive rocks on Axel Heiberg Island and northern Ellesmere Island confirms the age range of magmatism in the Sverdrup Basin Magmatic Province obtained from earlier studies. Two major pulses of igneous activity occurred in the east-central Sverdrup Basin and along the Sverdrup Rim during the Cretaceous: Voluminous magmatism of predominantly intrusive character peaked at 127-129 Ma, as indicated by the age of sills and dykes in the Buchanan Lake and Lightfoot River areas; and again at ca. 92-98 Ma, during the widespread eruption of flood basalts and emplacement of associated intrusions on central and northern Axel Heiberg Island; and of ferrobasaltic lavas and gabbros on northern Ellesmere Island. Moderate amounts of excess $^{40}$Ar in some samples could indicate the presence of a unique magmatic reservoir for flood basalts and their associated intrusive rocks, an observation reached independently on the basis of geochemical studies. The Santonian-Campanian ages obtained for two tholeiitic lava flows from the Strand Fiord Formation (85.7-81.2 Ma) appear to conflict with biostratigraphic ages reported for the overlying Kanguk Formation (Cenomanian-Turonian; e.g. Embry, 1991). However, both samples show comparable attributes in terms of age, presence of increasing Ca/K with temperature of analysis, and depressing of age in the first heating step, suggesting a similar crystallization history. The age data are consistent with the depositional model proposed by Nunez-Betelu and Hills (2001) for the Kanguk Formation (Cenomanian-Turonian; e.g. Embry, 1991). Furthermore, basaltic volcanism of Campanian age is documented from bimodal successions located on northern Ellesmere Island (Table 1; e.g. Merrett and Muecke, 1989). The possibility that Late Cretaceous volcanism occurred simultaneously in central and northern Axel Heiberg Island, on northern Ellesmere Island, and at the site of the Alpha Ridge, is being investigated.

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THE EARLY CRETACEOUS ARCTIC LIP: ITS GEODYNAMIC SETTING AND IMPLICATIONS FOR CANADA BASIN OPENING

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ABSTRACT

Recently, more data on the composition and age of the Early Cretaceous Arctic large igneous province have become available. Using published data and preliminary results of our study of basalts from Franz Josef Land and Bennett Island (De Long Archipelago), we review the origin and tectonic setting of this widespread province. Although the age of the Arctic province is not yet well constrained, existing data, primarily K-Ar dates, point to their formation between 130 and 100 Ma with a peak age between 115 and 110 Ma. The Arctic basalts are tholeiites with variable concentrations of major and trace elements, ranging from low-Ti and low-FeO tholeiites of the Franz Josef Land, to subalkaline, incompatible-element-enriched tholeiites from Bennett Island. Nb, Y and Zr data suggest a similarity to pre-break-up basalts from East Greenland, and the geochemistry of the basalts is consistent with the involvement of a plume-like mantle. We suggest that the Arctic occurrences of the Early Cretaceous basalts represent parts of Large Igneous Province dismembered and scattered around Arctic Ocean during the opening of the Canada and Eurasia oceanic basins.

INTRODUCTION

The importance of LIPs (Large Igneous Provinces) for studying the Earth’s interior and the initial breakup of continents is well recognized (Coffin and Eldholm, 1994). Worldwide, a significant period of basaltic volcanism occurred in the mid-Cretaceous, between about 125-85 Ma, when many continental flood basalts and oceanic plateau were emplaced (Larson, 1991). Occurrences of extensive outcrops of Cretaceous basalts have been recognized in the High Arctic since the first polar expeditions at the end of 19th century. They are scattered around the ocean basins, as outcrops on several islands, most notably eastern Svalbard (Kong Karls Land), Franz Josef Land, the De Long Archipelago and the Queen Elizabeth Islands in the Canadian Arctic Archipelago (Fig. 1). Data concerning the age, composition, and tectonic setting of the various parts of the Arctic flood basalt province are gradually accumulating (e.g., Maher, 2001). In the 1980’s, D.A. Forsyth et al. (1986) and H.R. Jackson et al. (1986)
demonstrated the similarity between the crustal structure of the Alpha Ridge and typical hotspot-related aseismic oceanic ridges, which has led to suggestions that other parts of the Arctic LIP may have been created by a mantle hotspot (see Lawver and Müller, 1994 for an overview). However, this concept has not been considered fully in context with all of the continental occurrences of the basalts. Furthermore, a comprehensive understanding of the Arctic basalts, in the context of their plate tectonic setting, is crucial for understanding Northern Hemisphere plate evolution, in particular the opening history of the Canada Basin.

GEOLOGICAL SETTING, COMPOSITION AND AGE OF THE CRETACEOUS ARCTIC FLOOD BASALTS

Franz Josef Land

The subaerial outcrop of flood basalts and related intrusives found in Franz Josef Land (FJL) is one of the largest in the Arctic. The basalts were first described during geological observations of the islands between 1895 and 1897 (see Bailey and Brooks, 1988; Embry, 1994; and Solheim et al., 1998 for overviews). Major contributions to geological investigations of the archipelago have been made by the Russian geologist V. D. Dibner and his co-workers (Dibner, 1998). Recently, additional data on the composition and age of the FJL basalts have been published by A.F. Grachev (2001) and T. Ntaflos and W. Richter (2003).

The FJL basalts are tholeiites, quartz tholeiites and basaltic andesites, with SiO$_2$ ranging from 48 to 54 wt%, MgO from 3.5 to 6.7 wt%, TiO$_2$ from 1.3 to 3.8 wt%, Na$_2$O from 2.1 to 3.6 wt%, and K$_2$O from 0.1 to about 1.5 wt% (Bailey and Brooks, 1988; Amundsen et al., 1998; Ntaflos and Richter, 2003; unpublished data of the authors). Many of the basalts show moderate iron enrichment, with total iron ranging from 10 to 17 wt%; the more enriched varieties are clearly ferrobasalts. There are broadly positive correlations between Fe, Ti, P, and K, and a negative correlation between these elements and MgO, consistent with low pressure fractionation of olivine, pyroxene and plagioclase. Scattered correlations between K and Ti indicate that there has been some post-magmatic addition of K in these rocks. Even if those samples with anomalously high K values are removed for the data set, the K$_2$O-TiO$_2$-P$_2$O$_5$ diagram shows the FJL basalts straddling the fields for oceanic and continental tholeiites, and the (Na$_2$O+K$_2$O)-total FeO-MgO diagram shows their tholeiitic affinities (Fig. 2A-C). The trace element data reveal some similarities between the FJL basalts and
ocean island basalts (Amundsen et al., 1998; Ntaflos and Richter, 2003), and some continental tholeiites, such as the initial pre-rift basalts in Greenland (Bailey and Brooks, 1988). This is further illustrated by the Zr/4-2Nb-Y and Zr/Y-Nb/Y plots (Figs. 3A and 3B). The high K and Ba contents of some FJL basalts and basaltic andesites suggests that the parental magmas may have been contaminated by continental lithosphere (lithospheric mantle or crust), although the relatively low initial $^{87}$Sr/$^{86}$Sr (0.7034-0.7059), high initial $^{143}$Nd/$^{144}$Nd (0.51264-0.51285), and low La/Nb (~1.1) suggests that large amounts of assimilation did not occur (Ntaflos and Richter, 2003).

The age of the FJL basic volcanism has not been reliably determined. Old published K-Ar dates are highly variable, from 200 to 60 Ma, while the biostratigraphic evidence indicates a Cretaceous age for the main volcanic phase (Dibner, 1998). Recent K-Ar dating of 22 samples of the FJL igneous rocks (Grachev, 2001) indicates a more tightly constrained period of 116±5 Ma for the basaltic event. This is in agreement with an unpublished $^{40}$Ar/$^{39}$Ar age of 117±5 Ma on FJL basalt (Pumhösl, 1998).

**Eastern Svalbard (Kong Karls Land)**

The first record of the basaltic rocks outcropping on the islands of Kong Karl Land (KKL) was made by the Swedish Arctic Expedition (1898) and subsequently reported by A.G. Nathorst (1899), A. Hamberg (1899) and H. G. Backlund (1907) (see Bailey and Rasmussen, 1997, for overview). In 1992, more field data were collected by the Norwegian Petroleum Directorate (Grogan et al., 1998). The basalts crop out on mountain tops, for example on Svenskøya Island, as either lavas or intrusive formations, and extensive sub-horizontal sheets have been recognised in seismic profiles to the east and south of Kongs Karl Land (Grogan et al., 1998).

According to J.C. Bailey and M.H. Rasmussen (1997), the KKL basalts are tholeiites with relatively high total FeO (13.1-18.2%) and high to moderate TiO₂ (2.7-3.5%), but low Al₂O₃ (10.5-13.3%), contents. The major element composition of the KKL basalts have an oceanic affinity (Fig. 2A-C) and correspond to the high FeO and low Na₂O MORB. However, the concentrations of incompatible trace elements are not typical of normal MORB, with Ba ranging from 44 to 355 ppm, high Nb abundances and low Zr/Nb (ca 13), similar to the FJL basalts. This results the KKL and FJL basalts plotting in similar fields on Figures 3A and 3B. Like the FJL, the KKL magmas may have been contaminated by addition of continental crust, as suggested by J.C. Bailey and M.H. Rasmussen (1997), but the amount of contamination is probably small (e.g., La/Nb ~ 1.1, indistinguishable from FJL basalts).

Based on to biostratigraphic dating, the KKL basalts have been considered as Late Jurassic to Early Cretaceous (Bailey and Rasmussen, 1997), but preliminary $^{40}$Ar/$^{39}$Ar dating by P. Grogan et al. (1998) has constrained their age to the Barremian-Albian time interval (131.8-97 Ma).

**Bennett Island**

Horizontal basaltic flows were found on Bennett Island by De Long’s 1881 expedition to the eponymous archipelago. The basalts were first sampled by E. V. Toll’ in 1902 (Wollosowitsch, 1909), and Toll’s samples were subsequently analyzed by B.M. Kupletskii (1930). Later the Bennett Island basalts were sampled and studied in late 1980’s (Drachev, 1987, unpublished; Fedorov et al., 2002). To this day, the Bennett Island basalts remain the least studied of the Arctic Cretaceous flood basalts.

The basalt formations on Bennett Island consist of 7 to 10 flat-lying flows of about 250-300 m total thickness. There are two distinct groups of the basalts: Mg-rich melanocratic picritic basalts occur at the bottom of the basaltic formation, whereas the leucocratic Al₂O₃-rich varieties form the upper parts of the section. Compositionally, the upper basalts resemble evolved continental tholeiites with elevated contents of Na₂O+K₂O, TiO₂, Al₂O₃, P₂O₅, Nb, Sr and Ba, and low MgO (3-4 wt%), Ni and Cr.

The age of the basalts is uncertain. One of us (SD) has performed K-Ar radiometric dating, in cooperation with the Institute of Precambrian Geology and Geochronology (Russian Academy of Sciences). The dates (119, 119, and 112±5 Ma) constrain the age of the basaltic magmatism to the Aptian stage, confirming Vol'nov and Sorokov's (1961) earlier estimate.

**Canadian Arctic Archipelago**

Volcanic rocks from the Cretaceous succession of the northern Ellesmere and Axel Heiberg islands (the Queen Elizabeth Islands, QEI) were first reported during Operation Franklin by the Geological Survey of Canada in 1955. In subsequent years, numerous occurrences of Cretaceous flood basalts were mapped and studied in detail (see Embry and Osadetz, 1988, for an overview), and age and composition data of the basalts have recently been published (Tarduno et al., 1998; Villeneuve and Williamson, 2003; Villeneuve and Williamson, 2006).
The QEI basalts occur mainly within the Isachsen (Valanginian to Aptian) and Strand Fiord (Albian to lower Cenomanian) formations, as well as within the upper Albian to lower Cenomanian Hassel Formation on NE Ellesmere Island. Basalts of the Isachsen Formation on the western Axel Heiberg Island are up to 230 m thick, with interbedded fluvial clastics (Embry and Osadetz, 1988). The Strand Fiord Formation basalts are nearly 800 m thick on Axel Heiberg Island, and are represented by mainly subaerial flows, underlain and overlain by marine shales (Ricketts et al., 1985; Embry and Osadetz, 1988). Basalts of the Hassel Formation on NE Ellesmere Island, interbedded with non-marine sandstone and shales, are up to 45 m thick (Osadetz and Moore 1988).

The QEI basalts are tholeiitic rocks relatively enriched with TiO$_2$ (1.6 to 3.7 %), Sr, K, Ba, Nb, P, Zr, as compared to normal MORB; their chemistry is similar to continental flood basalts. The Hassel Formation basalts are characterized by relatively high contents of total Fe (14.0-16.9 %), TiO$_2$ (3.0-3.8 %), and K$_2$O (0.8-1.5 %).

The majority of the isotopic ages for the QEI basalts, including recent $^{40}$Ar/$^{39}$Ar results by M. Villeneuve and M.C. Williamson (2003), fall between 99 and 89 Ma (Cenomanian to Turonian). These include basalts which are interbedded with marine sedimentary rocks that are of Aptian to Barremian age (115-125 Ma). These anomalously young ages could be interpreted as resetting ages, which coincided with the eruption of the younger (Late Albian to Early Cenomanian) basalts of the overlying Strand Fiord Formation (Trettin and Parrish, 1987; Tarduno et al., 1988). As consistent with the recent $^{40}$Ar/$^{39}$Ar data (Villeneuve and Williamson, 2006), there are two peaks of volcanism: 127-129 and 92-98 Ma. Presence of older flood basalts is in a good agreement with paleomagnetic data by P.J. Wynne et al. (1988), who discovered M0 and M1 reversed chron for the basalt lavas of the late Barremian to Aptian Walker Island Member of the Isachsen Formation that corresponds to 118-123 Ma time interval.

**DISCUSSION OF RESULTS**

Existing data on both the composition and age of the High Arctic Cretaceous flood...
basalts are sparse, with limited coverage of several of the regions. We therefore regard the following interpretations and conclusions as preliminary.

Although many of the Arctic Cretaceous basalts have compositions typical of continental tholeiites, the data indicate the existence of at least two distinctive groups of rocks. One group includes the FJL, KKL and some QEI tholeiites. Their compositions form a distinct trend on K2O-TiO2-P2O5 and AFM diagrams, straddling the divide between oceanic and continental tholeiites (Fig. 2), whereas the Zr/4-2Nb-Y diagram reveals their MORB like (between normal and enriched, P-MORB) compositions (Fig. 3A). (Note, however, that this diagram does not distinguish between MORB-like basalts and continental flood basalts.) The second group consists of the Bennett Island and some Ellesmere Island basalts. They are enriched with incompatible elements and display an intracontinental affinity. The latter is manifested on the K2O-TiO2-P2O5 and Zr/4-2Nb-Y diagrams (Fig. 3A, B).

All of the basalts form a well-pronounced and compact trend on the K2O-TiO2-P2O5 diagram from primitive FJL tholeiites to the most enriched Bennett basalts. These variations in the composition of the Arctic basalts are a result of variations in the degree and pressure of source melting, extent of low-pressure fractionation, and extent of crustal contamination, but a more exhaustive treatment of the data, and radiogenic isotope data, are required to fully evaluate the relative importance of these processes.

More details on the origin of the Arctic basalts can be obtained considering their Nb-Zr-Y composition, which is relatively insensitive to the variations in degree of mantle melting, low pressure fractionation, and crustal contamination (Fig. 3B). On the Nb/Y-Zr/Y diagram, many of the Arctic basalts fall into the part of the graph corresponding to enriched Icelandic basalts (Fitton et al., 1997). This may be because of contamination by continental crust, but could also be due to removal of clinopyroxene from the cooling magmas. Note that many of the basalts are evolved, with low MgO, although the samples plotted on figure 3B were initially filtered to remove samples with <5 wt% MgO.

Considering the tectonic setting of the Arctic Early Cretaceous basalts, we briefly compare them with the basalts from the North Atlantic Tertiary LIP. Based on Nb-Zr-Y systematics, the latter were discriminated by P.D. Kempton et al. (2000) into three main groups: (1) pre-breakup, (2) breakup and (3) post-breakup (Fig. 3B). FJL, KKL and Bennett basalts fall into the field of pre-breakup East Greenland basalts. This provides a basis to suggest a pre-breakup, plume-related setting for the Arctic suites. The latter has already been assumed for KKL, FJL and Sverdrup Basin Early Cretaceous basalts by J.C. Bailey and M.H. Rasmussen (1997) who applied the A. Grantz et al. (1979) plate tectonic reconstruction to show their relation to the rifting stage of opening of the Canada Basin. However, since the applied reconstruction assumed that the De Long domain had been attached to Siberian continent, these authors did not include Bennett basalt occurrence into the pre-breakup volcanic area.

We have developed a new plate tectonic reconstruction reassembling the Arctic continental massifs prior to opening of the Canada Basin (Fig. 4). This reconstruction, as many others, is based on a rotational mechanism proposed by S.M. Carey (1958), and to some extent resembles the reconstructions by A. Grantz et al. (1979) and D.B. Rowley and A.L. Lottes (1988). However, our reconstruction differs from the

Figure 4. Plate tectonic reconstruction of the Arctic on time preceding Canada Basin formation showing position of the Arctic Early Cretaceous LIP (outlined with thick dashed line). Black spots show flood basalt occurrences (KKL – Kong Karls Land, FJL – Franz Josef Land, QEI – Queen Elizabeth Islands, BI – Bennett Island). NW (Northwind Ridge), ChP (Chukchi Plateau) and AP (Arlis Plateau) denote continental fragments constituting present Chukchi Borderland. Lom – Lomonosov Ridge.
latter in the configuration of the North Alaska-Chukchi Block whose southern edge is assumed to coincide with South Anyui Suture and its western flank continuing to the southern New Siberian Islands (Natal’ in and Parfenov, 1983; Drachev and Savostin, 1993). The position of three small and, as we believe continental, blocks constituting the present Chukchi Borderland (Arlis, Chukchi Plateau (Cap) and Northwind Ridge) is, to some extent, similar to that proposed by A. Grantz et al. (1993) for the Chukchi Borderland. One of essential implications of the proposed reconstruction is that the De Long region forms an outer part of the North Alaska-Chukchi Block, which in this reconstruction was close to the Queen Elizabeth Islands. Therefore, the proposed pre-opening reconstruction shows all the occurrences of the Arctic flood basalts assembled in a tight cluster, which we consider to mark the Early Cretaceous Arctic LIP (Fig. 4).

CONCLUSIONS

The increasing amount of data on the composition and age of the High Arctic Cretaceous flood basalts have enabled us to review their origin and tectonic setting. Using published data and preliminary results of our study of the Franz Josef Land and Bennett Island basalts, we suggest the following:

1. The Arctic Early Cretaceous basalts are mostly evolved tholeiites with variable concentrations of the major elements from low-Ti and low-FeO tholeiites, to tholeiitic Bennett basalts enriched with incompatible elements. All are variably enriched in elements such as Ba and Sr, consistent with their emplacement in a continental setting.

2. Though the ages of the various parts of the province are not yet well constrained, existing data (primarily K-Ar dates) point to their formation between 130 and 100 Ma with maximum around 115-110 Ma, although there is well-defined basaltic magmatism of about 95 Ma in Axel Heiberg Island (Trettin and Parrish, 1987; Tarduno et al., 1998; Villeneuve and Williamson, 2006);

3. The composition of the basalts is consistent with the involvement of plume-like mantle in their formation. Nb/Y-Zr/Y systematics suggests their similarity to the East Greenland pre-breakup basalts;

4. Geologic data suggest that the De Long Plateau was a part of Chukchi-North Alaska Block before the opening of the Canada Basin. Restoring the Circum-Arctic blocks back to their pre-drift positions as required by rotational model of the opening of the Canada Basin, we find that all of the Cretaceous flood basalt occurrences are confined to a fairly compact area of about 1000 x 1500 km, which we consider as an Arctic Early Cretaceous LIP.

According to available geological and geochemical data, briefly overviewed in the present paper, we may conclude that the Arctic occurrences of the Early Cretaceous basalts represent parts of an Early Cretaceous LIP dismembered and scattered around Arctic Ocean due to opening of the Canada and Eurasia oceanic basins. Our results are consistent with a counterclockwise rotational model for opening of the Canada Basin and suggest a plume-related origin for this basin. As a pre-breakup setting of the Arctic LIP is a likely, more precise dating of the basalts is of critical significance to constrain the timing of the Canada Basin Opening.

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CORRELATION OF AEROMAGNETIC SIGNATURES AND VOLCANIC ROCKS OVER NORTHERN GREENLAND AND THE ADJACENT LINCOLN SEA

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ABSTRACT

Geological field studies carried out in 1994 and closely spaced aeromagnetic surveys flown in 1997 and 1998 were correlated to improve our understanding of the ice-covered continental margin of North Greenland and the adjacent Lincoln Sea. The anomaly patterns of the total magnetic field can be subdivided into three different magnetic units covering the North Greenland coastal area in the southeast and in the Lincoln Sea to the northwest. A broad magnetic low over the mountains of northern Greenland coincides with thick Paleozoic metasediments of the Franklinian Basin at the northern margin of the Greenland shield. Within this magnetic unit a number of small scale anomalies can be correlated with outcropping N-S-trending Cretaceous basaltic dykes. To the northwest, the magnetic anomaly pattern exhibits anomalies and anomaly chains running parallel to the coast over onshore and offshore areas. Onshore the magnetic unit coincides with the occurrence of a thick volcano-sedimentary succession of the Upper Cretaceous to Lower Tertiary Kap Washington Group. It is concluded that these volcanic rocks continue offshore parallel to the coast. Both the onshore and offshore volcanics are possibly associated with strike-slip movements parallel to the coast. The nature of the large, long-wavelength anomalies over large areas of the Lincoln Sea remains unclear. They may be caused by continental flood basalts on the passive margin off North Greenland. The plate-tectonic relationships of the volcanic activity that preceded the opening of the Arctic Ocean is discussed.

INTRODUCTION

Since 1992, the BGR has been involved in research on the lithosphere of the European and American Arctic. Originally, the CASE project (Circum-Arctic Structural Events) was initiated to compare the evolution and timing of the Tertiary West Spitsbergen Fold-and-Thrust Belt and the Eurekan Fold Belt in North Greenland. In 1994, geological structures and volcanic rocks were studied along the northern coast of Greenland during CASE 2 (Tessensohn and Piepjohn, 1998; Estrada, 1998; Estrada et al., 1999; Gosen and Piepjohn, 1999; Lyberis and Manby, 1999; Paech, 1999). In 1997 and 1998, aeromagnetic surveys over the Lincoln Sea and adjacent areas of northern Greenland were conducted in a joint Canadian-German project (PMAP-CASE) (Damaske et al., 1997; Forsyth et al., 1998; Nelson et al., 1998) (Fig. 1). To obtain a better understanding of the plate-tectonic development of the Arctic, the CASE studies have been extended since 1998 from the margins of North Greenland and the Barents Shelf towards the Canadian Arctic (CASE 4-8). An integrated onshore and marine program was carried out during a joint German-Canadian geological-geophysical cruise in Nares Strait in 2001. The main aim was to obtain new multidisciplinary data that would help to solve the controversial tectonic problem of the relative motion of Greenland and Ellesmere Island (Dawes and Kerr, 1982; Tessensohn and Piepjohn, 1998).

The opening of the Arctic Ocean, Baffin Bay and the North Atlantic since Late Jurassic/Early Cretaceous times was accompanied by intrusion and extrusion of igneous rocks. In this paper, onshore magnetic anomalies are correlated with geological structures, volcano-sedimentary basins and mafic dykes on land.

Figure 1: The Arctic region with the study areas of the PMAP-CASE and CASE 2 projects. MJR – Morris Jesup Rise, YP – Yermak Plateau, LH – Lake Hazen, JDP – Judge Daly Promontory.
The known onshore anomalies can then be traced onto the shelf, providing a basis for correlating the offshore anomalies with geological features on land.

**THE AEROMAGNETIC SURVEY**

In polar regions, aeromagnetic surveys provide an efficient tool to obtain data over a considerable area on the geological features underlying ice-covered areas of land and sea. Ideally, magnetic anomalies should be correlated with geological units; however, this is only possible if the flight lines are spaced sufficiently closely. If the line spacing is approximately the same as the depth of the magnetic sources to be investigated, then the anomaly will not be distorted (Bosum, 1981). However, a depth-to-source to line-spacing ratio of 1:2 to 1:4 is also acceptable. In the case of the Lincoln Sea and northern Greenland, the chosen distance between survey lines of 3 km and a flight elevation of 300 m above terrain are minimum requirements. A Convair 580 aircraft equipped with wing and tail stinger sensors was used as a platform for the surveys. The total magnetic intensity was calculated for each sensor, and the use of a combination of sensors permitted the horizontal and vertical gradients to be determined. Positioning was done with a GPS-system improved by referencing to a base station located close to the runway at the Canadian station Alert on Ellesmere Island, from where survey flights were carried out. Both surveys in 1997 and 1998 produced 30,000 km of aeromagnetic line data covering an area of 73,000 km² at an altitude of 300 m above ground and a line spacing of 3 km (Fig. 2).

After correcting for diurnal variation and statistical leveling of the profile lines and tielines using a GEOSOFT-OASIS software package, the data were reduced with the IGRF (1995 model). Here we present a portion (Fig. 3) of the 1997 anomaly map of the total magnetic field first presented as a poster at the ICAM III conference in Celle, Germany (Forsyth et al., 1998), showing the northern Greenland survey area between Mascart Inlet in the west and Sands Fjord in the east, as well as part of the Lincoln Sea to the north.

**MAGNETIC ANOMALIES AND THEIR CORRELATION WITH KNOWN GEOLOGICAL FEATURES**

The survey area can be divided into three zones (hereafter referred to as “magnetic units”, units 1 to 3 from SE to NW in Fig. 3) showing different magnetic anomaly patterns.

The southeastern magnetic unit (unit 1, Fig. 3) is characterized by a broad magnetic low (magnetic values below –200 nT) over the mountains of northern Greenland. It coincides with thick metasediments of the Paleozoic Franklinian Basin (e.g. Dawes and Soper, 1973; Surlyk, 1991) on the northern margin of the Greenland shield. Within this area (mainly in the north and northwest part), a number of small anomalies can be correlated with outcropping mainly N-S-trending basaltic dykes (Fig. 4).

The magnetic pattern of the adjacent magnetic unit to the northwest (unit 2, Fig. 3) reveals individual anomalies and rows of anomalies running parallel to the coast over onshore and offshore areas. This magnetic unit stretches from Jewell Fjord in the southwest, where the largest anomaly in the area is located (anomaly D, amplitude 1100 nT), to at least 50 km east of Kap Washington and extends seaward for about 25-30 km (north of Kap Morris Jesup). This magnetic unit can be subdivided into two parallel, NE-SW striking rows of anomalies. The onshore portion of the southeastern row of anomalies (anomaly E, maximum amplitude 785 nT) coincides with the volcano-sedimentary sequence of the Kap Washington Group of Late Cretaceous to earliest Tertiary age (Dawes and Soper, 1970; Batten et al., 1981; Brown and Parsons, 1981; Larsen,
DISCUSSION OF THE MAGNETIC ANOMALY UNIT OF THE COASTAL REGION

The Kap Washington Group is overthrust from the southeast along the Kap Cannon Thrust Zone by Paleozoic metasediments belonging to the North Greenland fold belt (Dawes and Soper, 1970; Brown and Parsons, 1981; Soper and Higgins, 1987; Gosen and Piepjohn, 1999) (Fig. 4). This is evidenced by the fact that the unit-2 magnetic anomaly pattern extends to the southeast of this thrust zone (Figs. 3 and 4). The southeastern row of anomalies also extends along the coast – with attenuated amplitudes – to about 40 km east of Kap Cannon, and east of the apparent NE end of the thrust zone (see Figs. 3 and 4). This suggests that intrusive rocks (probably intrusive equivalents of the Kap Washington volcanics) exist at depth beneath the Paleozoic sediments. The differences in amplitude as well as the totally different direction of the row of anomalies make it rather unlikely that the source bodies could be dykes extending from magnetic unit 1 to the coast.

The northwestern row of anomalies in magnetic unit 2 runs offshore parallel to the coast (Fig. 3). It has a similar magnetic signature to the southeastern row of anomalies dealt with in the last paragraph. We infer from this that the sources of these anomalies are also volcanic rocks.

A similar magnetic pattern can be found in the
Nares Strait region: Damaske and Oakey (in press) report on magnetic anomalies over known Tertiary basins on Judge Daly Promontory (northwestern coast of Nares Strait, Fig. 1) (Miall, 1981). The continuation of the magnetic anomalies from here to the NE makes it likely that Tertiary basins extend all along the northern part of Nares Strait into the Lincoln Sea. The exposed pull-apart basins were formed by sinistral strike-slip tectonics (Mayr and deVries, 1982; Tessensohn and Piepjohn, 1998; Tessensohn et al., in press). Erosion products of 61–58 Ma old alkaline volcanism in the Nares Strait region (Estrada et al., in press) were deposited in these pull-apart basins. The detrital volcanic material in the Tertiary basins is the source of the magnetic anomalies.

A similar situation can be inferred for the Kap Washington area. Likewise, the magnetic anomalies of unit 2 provide a hint of strike-slip movements parallel to the coast. Possibly, the Kap Washington Group and its offshore equivalents were also formed in pull-apart basins.

If the Kap Washington volcanism is related to strike-slip tectonics, these movements would have already started during the Late Cretaceous. Plant microfossils in intercalated shales indicate a Campanian or Maastrichtian age of the Kap Washington Group (Batten et al., 1981), while radiometric dating on volcanic and pyroclastic rocks yields earliest Tertiary ages (about 64 Ma; Larsen, 1982; Estrada et al., 1999). Onshore strike-slip movements in the Kap Washington area are unknown. However, they may be overprinted by the later (?Eocene) thrust event which carried Paleozoic metasediments of the North Greenland fold belt northward over the Kap Washington Group (Fig. 4).
DISCUSSION OF THE LINCOLN SEA ANOMALIES

Magnetic unit 3 is confined to the continental shelf/slope area (water depths not exceeding 300 m; Perry and Fleming, 1986) off North Greenland. Owing to the difficult ice conditions, neither marine surveys nor drilling has been attempted in this area. The only seismic line in the Lincoln Sea is in the northern portion close to the Lincoln Sea Plateau (Forsyth et al., 1994). The magnetic anomalies over this area are different in size and shape in comparison with those of the southeastern Lincoln Sea and thus do not provide a reliable basis for interpreting magnetic unit 3. Some distance away from our study area, to the northeast, over the northern flank of the Morris Jesup Rise, a seismic line exists in an area of magnetic anomalies comparable with our magnetic unit 2 (Jokat et al., 1995). However, no structures to which the large magnetic anomalies might be attributed were discovered within the Morris Jesup Rise basement below a sedimentary cover of about 100-200 m. The high amplitude, long-wavelength magnetic anomalies over the Morris Jesup Rise have been known for some time (Feden et al., 1979). The Morris Jesup Rise is interpreted as an oceanic plateau formed in connection with the Yermak Plateau during the early opening of the Eurasian Basin (Dawes, 1990 and references therein).

The large, long-wavelength anomalies of magnetic unit 3 are similar to the anomalies over the Morris Jesup Rise. If the rise consists of volcanic rocks, it is likely that the anomalies of unit 3 over the continental shelf are caused also by thick volcanic sequences (? continental flood basalts). This idea is supported by the continuation of the anomaly pattern (with lower amplitudes) towards northeastern Ellesmere Island (Forsyth et al., 1998), where Cretaceous basalts and mafic dykes are exposed onshore in the Lake Hazen region (Fig. 1) (Embry and Osadetz, 1988; Osadetz and Moore, 1988; Embry, 1991). They are geochemically comparable to continental flood basalts (Estrada and Henjes-Kunst, 2004).

THE VOLCANIC ROCKS AND THEIR PLATE-TECTONIC RELATIONSHIPS

Several different igneous events are recorded or inferred in the study area:

- intrusion of basaltic dykes (onshore),
- Kap Washington volcanism (onshore and near-coast offshore),
- inferred thick volcanic sequences, probably continental flood basalt volcanism (offshore).

The N-S trending basaltic dykes in Paleozoic metasediments causing the small narrow anomalies (magnetic unit 1, see above) are related to the basalts of the Kap Washington Group (see below) in view of their similarity in petrology and chemical composition (Soper et al., 1982; Estrada, 1998). Rb/Sr isotopic ages of 103, 93, and 92 Ma have been obtained on biotite separates from some of these dykes (Lyberis and Manby, 1999). Since the dykes are of Early Cretaceous to early Late Cretaceous age, they precede the Kap Washington volcanism, which is of Late Cretaceous to earliest Tertiary age.

Early Cretaceous to Cenomanian basalts are known from the Canadian Arctic Islands (Embry and Osadetz, 1988; Embry, 1991). These tholeiitic basalts are chemically comparable with continental flood basalts (Estrada and Henjes-Kunst, 2004). Early Cretaceous basalts, basaltic andesites and mafic dykes are also known from Franz Josef Land (e.g. Campsie et al., 1988; Bailey and Brooks, 1988; Dibner, 1998; Ntaflos and Richter, 2003) and Svalbard (Burov et al., 1977). All these volcanic and subvolcanic rocks can probably be related to an initial stage of volcanism preceding the opening of the Arctic Ocean (Canada Basin ?; Fig. 5a). Embry (1991) assumes that the rifting of the Canada Basin took place between about 138 and 92 Ma and seafloor spreading was active from about 92 Ma to 67 Ma.

The Kap Washington Group is of Late Cretaceous to Paleocene age and comprises bimodal, partly alkaline volcanic and pyroclastic rocks (Dawes and Soper, 1970; Batten et al., 1981; Brown and Parsons, 1981; Larsen, 1982; Soper et al., 1982; Brown et al., 1987; Estrada et al., 1999; see Fig. 4). Petrographically and geochemically, the bimodal volcanic rocks of the Kap Washington Group compare well with the Late Cretaceous Hansen Point volcanics described by Trettin and Parrish (1987) from the northern coast of Ellesmere Island (Estrada et al., 2000). Another alkaline volcanic province can be reconstructed using volcanic pebbles preserved in Paleocene sedimentary basins along Nares Strait. 40Ar/39Ar dating on feldspar, amphibole and whole-rock fractions of the volcanic pebbles resulted in good plateau ages of 61–58 Ma (Estrada et al., in press). All the Late Cretaceous to Paleocene alkaline volcanic suites show intra-plate geochemical signatures and were probably formed in several branches of a large continental rift system preceding the opening of the Eurasian Basin (Estrada and Henjes-Kunst, 2004). The seafloor spreading of the Eurasian Basin started about 56 Ma ago (Srivastava and Tapscott, 1986).

We interpret the alkaline volcanism of the Kap Washington area and of the Nares Strait region to be related to strike-slip movements (Fig. 5b), which is also
Figure 5: Plate tectonic situation and volcanism a) during Early Cretaceous to Cenomanian and b) during Late Cretaceous to Paleocene. Map 5a) after Vogt et al. (1982). Map 5b) based on a paleogeographic reconstruction for Chron 31 (68 Ma) after Srivastava and Tapscott (1986), modified by Tessensohn and Piepjohn (1998). Seafloor spreading in the Labrador Sea south and southwest of Greenland between Chron 34 (about 84 Ma) and Chron 24 (about 55 Ma) causes northeastward movement of Greenland and Eurasia with sinistral strike-slip motion along Nares Strait (Srivastava, 1978; Peirce, 1982). References: (1) Embry (1991); (2) Ntaflos and Richter (2003); (3) Lyberis and Manby (1999); (4) Burov et al. (1977); (5) Estrada and Henjes-Kunst (2004); (6) Batten et al. (1981); (7) Larsen (1982); (8) Estrada et al. (1999); (9) Estrada et al. (in press).

CONCLUSIONS

A closely spaced aeromagnetic survey over the North Greenland coast and the adjacent Lincoln Sea shows that the volcano-sedimentary Kap Washington Group continues offshore. We interpret the magnetic pattern as due to a chain of small basins formed by a strike-slip system parallel to the coast, in which volcanic and volcaniclastic rocks were deposited.

The basins are probably related to a larger system of strike-slip movements along the present northern margin of Greenland as a result of the Late Cretaceous to Paleocene plate-tectonic development. The sea-floor spreading in the Labrador Sea between anomalies 34 (84 Ma) and 24 (55 Ma) caused northeast directed movement of the Greenland-Eurasia plate. This motion was accompanied by sinistral strike-slip tectonics in the Nares Strait region (Srivastava, 1978; Peirce, 1982) (Fig. 5b).

The nature of the extensive anomalies of the Lincoln Sea remains unclear. We infer that they are caused by continental flood basalts on the passive
margin off North Greenland. The sparse recent geophysical surveys and geological data are insufficient for a conclusive explanation. Marine seismic surveys or ice-flow based seismic experiments are necessary. Finally, drilling into the shelf will be the only way to obtain direct geological evidence.

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The assemblage of basic igneous rocks on Franz Josef Land Archipelago comprises Na-tholeiites and andesitic basalts and their derivative subalkaline rocks. A collection of flood basalts and dolerites were accumulated from 15 islands of the archipelago during several field seasons. Bulk chemical and trace elements analyses were executed in laboratories of VNIIOkeangeologija and the University Oslo. Thirty four samples were selected for analysis from northwest-striking dikes, sills and a series of lava flows out of more than 300 samples of basic rocks. The raw samples selected were taken from a northwest transect oriented across the most active tectonic faults and zones of initial magmatic activity.

We have studied the distribution of major oxides in northwest-trending dykes and a series of lava flows in exposed sections on the islands of the archipelago. A consistent differentiation trend was discovered that was apparently controlled by an igneous center near Hocker, Brady and Hall Islands (Fig. 1) in the southwestern part of the archipelago. The center of igneous activity of the archipelago typical has relatively more basic rocks containing low SiO$_2$ and high MgO. A gradual increase of SiO$_2$ and decrease of MgO content (Fig. 2) and an abrupt increase of Ti, Fe, Na and K contents were discovered northeast and southwest of the center (Fig. 3). The distribution is characteristic of the progressive differentiation of basaltic melt under decreasing crystallization temperature (Amundsen et al., 1998; Ntaflos and Richter, 2003).

The geological investigations in natural exposures shows different interrelations of sills, dikes, basal flows:

1) subalkaline basalts are overlain by tuff, then intruded by a sill and neck, which is in turn overlain with tholeitic basalt (Fig.4--A);
2) cranked system “dike-sill” (Fig.4–B);
3) dike cutting basalt sheet (Fig. 4–C);
4) dolerite sill cutting basalt flows (Fig. 4–D);
5) multiple intrusive complex “dike in dike” (Fig.4 – E);
6) dike cutting sill (Fig. 4–F);
7) sill cutting basalt sheet (Fig. 4-G).

The distribution of trace elements such as Ni and Cr in basalts and dolerites supports the location of the serious oceanic ridge that was the source of the volcanics of Franz Josef Land Archipelago (Table 1).
magmatic center. Ni and Cr contents vary proportionally in the dykes and in the lava flows and reach the lowest values in the intrusive beds (Fig. 5). This trend provides evidence that the latter originated from the residual melt at the concluding stages of fractional crystallization of the initial magma. It is important to note that dykes cutting the sills have low contents of Ni and Cr, comparable to contents of those elements in the sills (Figs. 4-D and 4-F). The contents of these elements decrease both in dykes and lava flows across the “Hooker-Hall” central zone to the northeast and southwest towards the zones margins.

The composition of magmatic rocks also varies moderately along the strike of the northwest trending zone. The SiO₂ and Na and K contents decrease gradually to the northwest (Figs. 2 and 3).

The compositional variation of major oxides and trace elements of basalt flows in layered sections, documents the stages of fractional crystallization of the initial magma (Figs. 2 and 3 – basalt floors in southwest trend). A section on Hall Island shows a succession opposite to that of the typical layering of magma in an intermediate (inside crust) chamber, where light subalkaline lava from the top of the chamber constitute the base of the outcrop section, and tholeiitic lava from the lower portion of the chamber forms the top of the section (Fig. 4-A). The successions are different on Hooker, Mabel, Jackson, Payer and Greely Islands, where the upper portions of the sections contain more alkaline basalt with higher Fe content. This chemical evolution is in accordance with a concept of progressive differentiation of the initial melt at relatively deep levels in the lithosphere. The basalts in the Hooker, Mabel, Jackson, Payer and Greely Islands sections show stable contents of CaO, Al₂O₃ and TiO₂, indicative of the absence of clinopyroxene and plagioclase crystal differentiation at the P-T conditions in an intracrustal intermediate chamber. The magmatic sequence on Mabel Island comprises two cycles. Both cycles are topped with more alkaline basalt. The increase of Fe, Na and K contents in the younger basalt sheets on both Hooker and Mabel Islands is followed by the increase of Cs, Rb, Ba, Nd, U, Th, K, Nb, La, Ce, Sr, Zr, Sm, Eu, Ta, Mn contents and the decrease of Ni, Cr, Cu and Sc contents.

Stages of the evolution of the basaltic magma were manifested regionally, as is evident from the MgO content in the residual melt. A high MgO-content zone (Hooker–Mabel), a medium MgO-content zone (Champ–Graham–Bell), and a low MgO-content zone (Hooker–Rudolph) have been identified.

A combination of plume (Grachev, 2001) and spreading models is hypothesized to control the evolution of basic igneous activity in the region. The diffusive spreading axes identified as early magmatic manifestations were located within a linear zone transecting the region in a northwest direction in the Hooker and Hall Islands area. The existence of such spreading is supported by the existence of “dike in dike” structures (Fig. 4–E). The spreading was initiated by a plume centered at the site of the least differentiated traps with the highest content of Mg, Ni
Figure 4. Geological interrelation of sills, dikes and basalt flows and Cr, Ni, and Ti contents in some of them. A-subalkaline basalts are overlain by tuff, then intruded by sills and necks the over lain by tholeitic basalt; B-cranked system “dike-sill”; C- dike cutting basalt sheet; D-dolerite sill cutting basalt flows; E-multiple intrusive complex “dike in dike”; F-dike cutting sill; G - sill cutting basalt sheet.
and Cr, in the southern portion of the above spreading zone.

The age of Franz Josef Land magmatism is Late Jurassic–Early Cretaceous. This is confirmed by geological (Ntaflos and Richter, 2003) and radiological data (Stolbov, 2006). The extensional setting was succeeded by regional compression related to the opening of the Eurasian Basin in the Early Cretaceous–Paleocene. The compression was documented by the tectonic separation of the initially continuous igneous bodies and the sense and scale of relative motions of their fragments. For example, the northeast-striking fault on Wilczek Land Island cuts the dikes and sills (Fig. 1).

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Figure 5. Ni and Cr varies proportionally in dykes and in lava flows and reach the lowest values in sills.
Anomalous uplift and magmatism has affected a region encompassing the North Atlantic Ocean and fringing continental margins during Cenozoic times. These phenomena most likely result from temporal and spatial variations in Iceland Plume activity. A mantle plume generates uplift in two ways. First, magmatic material intruded into the crust produces permanent uplift. Secondly, the abnormally hot mantle and associated upward flow of the plume generates dynamic support which uplifts the overlying plate. As the mantle plume moves away from an area, or cools, the buoyant effect of this hot plume material is removed and the area subsides. This second type of uplift is temporary, but such transient vertical motions can be recorded in the stratigraphic record of the region overlying the plume, enabling the size and planform of the mantle plume head to be mapped out through time. Here, we describe the record of activity of the southern half of the Iceland Plume head, a region encompassing the British Isles, southern Greenland and eastern Canada. This record is based on the stratigraphic record of continental basins combined with

![Figure 1. Backstripped subsidence histories and residual plots for 2 wells offshore British Isles. (a: Outer Moray Firth Basin, well 14/25-2; b: Porcupine Basin, well 35/19-1.) Map of British Isles showing well locations. The well plots illustrate the estimation of Paleocene dynamic support from extensional basin stratigraphy. Two plots shown for each well. Left-hand plot: Backstripped data points show observed subsidence history, solid line is theoretical subsidence curve from inverse subsidence modelling of data points up to the end of the syn-rift period; discrepancy between the two (residual subsidence), shown in right-hand plot. Grey shading is syn-rift period constrained by seismic reflection profiles and vertical error bars reflect uncertainty in water depth at time of deposition. The transient uplift event was generated by dynamic support from the underlying edge of the Iceland Plume head during the Paleocene. The amplitude of dynamic support was greater in the west of the British Isles (b), than in the east (a).](image)
observations of oceanic crustal thickness and structure. Plume-related uplift commenced in earliest Paleocene time. The plume initiation phase culminated with widespread transient uplift at the Paleocene/Eocene boundary (Nadin and Kusznir, 1995). In the Porcupine Basin, off SW Ireland, peak dynamic uplift was 500-800 m (Jones et al., 2001). In the Outer Moray Firth Basin, part of the North Sea Rift System, peak dynamic uplift was 180-425 m (Figure 1; Mackay et al., 2005). Fossil coastlines and transient anomalous uplift measured within sedimentary basins constrain the contemporary planform and amplitude of convective support (Jones and White, 2003). Estimates from continental stratigraphy compliment convective planforms obtained from distribution of plume-related magmatism. Widespread latest Paleocene uplift is most likely explained by rapid lateral injection of a hot layer beneath the lithosphere. Our results suggest that the planform of the plume head was not circular, it is most likely to have been lobate.

Both oceanic crustal thickness measurements and continental anomalous subsidence records suggest that the large starting plume head shrank through Early-Mid Eocene time. V-shaped oceanic crustal thickness anomalies suggest that plume flux varied with a principal periodicity of 5–6 Myr and at smaller periodicities during Oligocene-Recent times (Jones et al., 2002). This fluctuating dynamic support has acted to raise and lower the bathymetrically shallow Greenland-Scotland Ridge, periodically inhibiting the flow of oceanic currents between the Arctic and North Atlantic Oceans (Wright and Miller, 1996; Poore et al., 2006).

The stratigraphic records of the extensional basins which fringe the North Atlantic and Arctic margins yield valuable information about the temporal and spatial evolution of the Iceland Plume during Cenozoic times. Such records of mantle plume evolution will provide important new constraints for models of mantle convection.

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Given that significant geomorphological impact in the coastal regime occurs during storms, a storm identification algorithm was established that applies a series of decision rules to station observed wind speed data. Using the results from the algorithm a database of storm events was assembled. From this database, climatologies were assembled for the following parameters, by station by month: mean storm count, mean storm speed, mean storm duration, and mean storm power. These parameters were then aggregated according to sector definitions established by the Arctic Coastal Dynamics Project (ACD).

Results indicate a difference between those regions proximate to the influence of the mid-latitude oceans (the Atlantic and, to a lesser extent, the Pacific), and those of the north arctic coast. Regions influenced by the oceans show a mid-latitude profile typified by dominant winter-time activity with a minima during summer. Arctic north coast regions, however, show a strong activity peak in the late summer/early fall, coincident with the height of the open water season. Storm potential power, defined as the square of the mean speed multiplied by duration, possesses some of the largest values in the Chukchi Sea zone in late open water season.

INTRODUCTION

Coastal regions are particularly sensitive to the impact of high magnitude weather events. Wind energy drives powerful wave action and water level surges, which have a variety of effects. These include significant geomorphological, infrastructure, and ecological impacts, and consist of, for example, wave action causing erosion and sedimentological work (e.g. Solomon et al., 1994, Harper et al., 1988), surge-induced inland flooding resulting in significant vegetation kill (Reimnitz and Mauer, 1979), and infrastructure damage due to the strength of positive buoyancy of wooden or hollow objects and structures (Hume and Schalk, 1967). The ocean is able to move large quantities of material both to and from the coastal zone, i.e. accretionary and erosional coasts (e.g. Grigoriev et al., 2002 and Hume and Schalk,1967), with the net result that a given coastline can undergo significant change on an annual basis. A number of researchers report that most geomorphological work done to a coast occurs during significant events (Reimnitz and Mauer, 1979 and Solomon et al., 1994).

To provide arctic researchers with more information about storm activity fifty years of storm event occurrences (defined below) were extracted from observational wind data for coastal regions of the circumpolar arctic (Fig. 1). Climatologies, consisting of aggregate totals and means by region, will be presented. Data over the period 1950 – 2000 were utilized.

METHOD

Station selection and data preparation

Initial station selection was conducted to satisfy the requirements of the Arctic Coastal Dynamics project; for this work observational data from 59 stations were retained and processed (Fig. 1). Data pre-processing was necessary to handle the wide variety of
observing regimens noted amongst the stations. All stations were standardized to a 6-hourly observational regime. Where there were insufficient data to adequately represent the 6-hourly regime for each month of the year, a station-year was excluded. This prevented a potential artificial reduction in storm counts.

Storm identification

The identification of a “storm”, in terms of coastal issues, is focussed on the right combination of wind speed, duration, and direction such that the wave-generating and/or surge generation potential is maximized. Depending on the specific application a range of threshold speeds, durations, and form of wind profile (e.g. MacClenahan et al., 2001) must be considered. For this project the interest was in storms that can produce waves and/or surges that are of sufficient magnitude to cause damage to habitats or infrastructure, and/or to perform geomorphological work. Only storms with winds above a certain speed that are maintained for a certain period of time are able to create this sort of impact (e.g. MacClenahan et al., 2001 and Solomon et al., 1994). An “event threshold” for wind speed was thus set at 10 m/s. Although this speed is lower than what a storm would traditionally be considered capable of generating, this speed was selected based on precedents that have been set in arctic geomorphology work (e.g. Hudak and Young, 2000 and Solomon et al., 1994). In this sense the interest is not purely meteorological but partly geomorphological. Others, such as, MacClenahan et al. (2001), set their speed threshold with reference to the more powerful storms to reach the Irish coast. The “duration threshold”, that is, the minimum length for a storm event, for the project described here was set to six hours, as based on various other researchers (e.g. Hudak and Young, 2000 and Solomon et al., 1994).

For this project the following, multi-stage algorithm for identifying a storm event was utilized. To begin with, any wind speed exceeding event threshold is tagged. The tagged observations are then assessed for grouping into discrete storm events. In performing this task two morphological features in the presentation of a storm event in a wind speed profile are recognized: “lulls” and “shoulder events” (Fig. 2). A lull is a temporary decrease in wind speed during a synoptically contiguous storm event. In this case lulls were defined to be the occurrence of a single wind speed observation that dropped below threshold, with tagged observations immediately previous to and following the lull. A shoulder event was defined as a wind speed occurring before the first, or after the last, tagged observation in a contiguous set of tagged observations, that was just a bit below event threshold, i.e., and most likely associated with the synoptic storm event.

To evaluate lulls and shoulder events a secondary threshold, termed the “continuity threshold”, was defined. This value was arbitrarily set at 0.7 x event threshold. If the lull dropped below this value the tagged events on either side were considered to belong to separate storm events. If a shoulder dropped below this value then the event count tagging was stopped for that event. Without use of lulls the algorithm returned too many storm events that were too short in duration. Without use of shoulder events the synoptic duration of storm events was overly shortened.

After the addition of lulls and shoulder events to the storm events the six-hourly wind speed observations were linearly interpolation to a one-hour frequency. This served to refine estimates of duration and mean speed. Counting began when the hourly wind speed estimate rose above the continuity threshold, and similarly ceased when it dropped below.

This resulted in the generation of a storm event database for each station used in the study. This database served as the basis for the analyses described next.

**Figure 2:** Schematic representation of storm representation in the wind speed profile with various components indicated.
The analyses were based on the aggregation of station results into sector boundaries established by the ACD (Rachold et al., 2003) and modified slightly for this project (Fig. 1). The first set of analyses dealt with the preparation of fifty-year climatologies of selected parameters for each sector, by month. The parameters included, by station: mean event count, mean core wind speed, mean core duration, and mean total power (defined below). The aggregation method varied slightly depending upon the parameter being considered. Determination of a sector mean event count was performed in the following manner. First a mean event count by station, by month was obtained. From station counts the sector mean count was computed. Storm event durations by sector by month were calculated in a manner similar to that used for the frequency count, i.e. a mean duration by station was first determined, and then a mean duration by sector was calculated.

From the parameters retained for each storm event a derived parameter, storm wind power, was calculated. This parameter is designed to provide a rough indication of the total power available from a storm event, and is defined as the square of speed multiplied by the duration. Calculations for wind power were based on a subset of storm events selected on the basis of mean direction, that is, if the storm mean direction was from the north sector the event was retained. This is a very rough representation of the fact that many of the coastal observing stations are situated on a coastline oriented east-west and are exposed to water in the north and land in the south. Storm events of interest for derived impacts studies are thus those with a prevailing northerly direction.

RESULTS AND DISCUSSION

The long-term mean distribution of storms through the year (Fig. 3) revealed a mid-latitude, northern oceanic influence in zone 1, situated closest to the Atlantic. The low point in July, coupled with the steady rise into the fall, is a typical mid-latitude storm activity cycle. Zone 1 is also situated northeast of the Icelandic Low, a region of strong atmospheric low-pressure system activity in the winter, and is also in an area of strengthened east-west flow that directs weather systems into the region (Wallén, 1970: p26-27). The pattern in zone 2 is similar in form to that in zone 1 but in each month the mean annual number of storms is greater, and there is a definite late open-water season peak. This pattern is deemed to be a mixing of two signals, one being a vestige of the mid-latitude signature, and the other being the increase in storminess potential due to temperature differences between the land and the sea. By zone 3 the mid-latitude signature in the storm count pattern is gone. Instead this area sees more activity in the summer with a small drop into the fall. In zones 3, 4, and 5 the storm counts reach a high level in August. Even though open-water has not reached its greatest extent by August, the temperature of the land is at its highest level, thus providing the strongest land-sea thermal gradient of the annual cycle. This is able to compensate for a relative lack of open water, resulting in a storm count peak. Counts in September and October remain high in zones 3, 4, and 5 due to the rapidly increasing open water extents, even though the thermal gradient is dropping. Zone 5 has lower counts in June and July most likely because the storm tracks that develop in the summer channel systems towards zones 2-4 and not 5 (Lydolf, 1977). Zone 6 comes under the influence of systems that can move up from the Pacific Ocean either through the Bering Straight or across Alaska. Increasing open water amounts over the five-month period coincide with a general increase in storm activity. Zone 7 is located farther north than the previous 6 zones. It does not have the same open water season, and the northwest edge of the archipelago can have virtually no open water season. Here general storm activity is low compared to other regions. Its peak occurs in July and August as a result of changes in prevailing synoptic pattern; specifically, in the summer, weather from the south can penetrate into this region more readily than at other times of the year.

Zones 1-4 were similar for storm mean speed (Fig. 4), with summer (July and August) representing a speed low point, June representing a secondary peak, and September stronger leading up to October, which possessed the strongest mean winds in all zones for the open water season. In Zone 5 the June secondary peak was reduced to a level below July and August. Zone 6 had low speeds in general, especially for June and July. Zone 7 had consistently strong winds.

Storm durations (Fig. 5) in zones 1 through 4 exhibit the mid-latitude signature. Storms in zone 2 exhibited the longest durations of any sector for each month. The pattern breaks down in zone 5 because June storm durations are significantly lower relative to the other months. The pattern is apparent again in zone 6, although durations are short compared to other zones. Zone 7 does not have as strong an increase into September and October as other zones, and durations are generally short here as well.

Storm potential power (Fig. 6) was calculated as described. A rise in mean power values from a low point in July (zones 1,2,3,5,6) or August (zones 4,7) leading up to October was apparent in all zones, as was a large variability in June.
Figure 3: Long-term (1950-2000) mean distribution of storms over the year by station, aggregated by sector. The box represents a rough open-water season.

Figure 4: Long-term (1950-2000) mean distribution of speeds in storm cores by station, aggregated by sector. The box represents a rough open-water season.

Figure 5: Long-term (1950-2000) mean distribution of duration of core winds (hours) by station, aggregated by sector. The box represents a rough open-water season.

Figure 6: Long term (1950-2000) mean distribution of storm power (speed$^2$ x duration) by station, aggregated by sector. The box represents a rough open-water season.
Overall, zones 1, 2, 3 and 4 exhibited consistently high mean power values, zone 5, a large relative difference between June and October, and zones 6 and 7, consistently lower power values.

CONCLUSION
The annual cycle of storm frequency, duration, wind speed, and power vary by region over the circum-Arctic domain, and reflect the influence of the oceans in regions near the oceans, with an increasingly important Arctic coastal signal appearing for regions far from the oceans. Specific observations include:
➢ The greatest power values were observed not in the Atlantic zone but in zone 5, the Chukchi zone, late in the open water season.
➢ The Kara Sea zone (2) is a very active region, having many storms with long durations.
➢ A strong storm count peak that appeared in June in the Kara Sea zone (2) and the Laptev Sea zone (3) was noted and is possibly linked to the frequent occurrence of early open water off the mouths of the Ob and Yenesey Rivers in the Kara and the Lena River in the Laptev, caused by voluminous June discharges from these rivers (Lammers and Shiklomanov, 2000) and/or polynyas in both of these Seas.

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ABSTRACT

In August 2001 a large low-relief iceberg, loaded with rocks and difficult to make out in the foggy night against dark water was encountered by the icebreaker CCGS Louis S. St-Laurent in Kane Basin between Canada and Greenland (Latitude 80° N). The debris cover was sampled and the specimens compared with the regional geological units in nearby Greenland and Ellesmere Island, to ascertain the possible source glacier. Preliminary analysis of these samples and consideration of prevailing ocean currents suggest calving of the iceberg from the Petermann Glacier, which drains the Greenland ice sheet and derivation of the related sediment load from adjacent rock outcrops of Washington Land or western Hall Land. It is speculated that this unusual occurrence is the result of relatively high summer temperatures and heavy rainfall that may have enhanced mass wasting from the steep valleys onto lateral moraines, and to a larger than usual ice breakup from Petermann Glacier. These rock-loaded, low-profile, near neutral-density icebergs may pose a serious threat to navigation in northern waters.

INTRODUCTION

On the foggy night of August 17, 2001, during the Canadian-German Nares Strait Geocruise 2001 expedition on board the icebreaker CCGS Louis S. St-Laurent, we had a startling encounter with a large tabular iceberg weighted down with dark rocks, difficult to discern at a distance against the dark ocean, with a low profile very different from typical icebergs. The experienced crew indicated having never before observed such a large rock-covered, low-profile iceberg. The locality (Latitude: 80° 07.705′ N; Longitude: 69° 53.908′ W) was near cape Hayes, northeast Ellesmere Island (Fig. 1). In subsequent days, several large tabular icebergs, and many small, low-profile, rock-loaded icebergs were seen in the same region, most likely from the same source.

Currents are the major factor controlling drift and transportation of Arctic icebergs (Murray, 1969). A dominating and relatively constant southward flow of cold water and ice from the Arctic Ocean through the Nares Strait to Smith Sound (Fig. 1) dispenses sea ice and icebergs into Baffin Bay (Melling et al., 2000). The extent to which the West Greenland Current, an opposing north-west current following Greenland’s coastline affects the southerly flow, is not fully understood. Mixing of the two currents is thought not to occur any farther north than Smith Sound (Melling et
al., 2000). Therefore, current direction at the head of Kane Basin where the iceberg was encountered is in an area of predominantly southward flow.

METHODS

Using the bubblers the ship was positioned beside the berg and two of the authors (MZ and JCH) were lowered in a basket to its surface (Figs. 3 and 4). Rock specimens were collected in order to ascertain the possible origin of the iceberg and its rock cover (Zentilli and Harrison, 2002). Sampling was not systematic, but rather to collect a wide variety of lithologies, fossils, and bituminous rocks that could possibly be used as tracers. Hand specimens were labeled and described with the naked eye and under a hand lens and binocular microscope; fossils were tentatively identified. The size and characteristics of the berg were analyzed using photographs taken from the ship. Recently, RockEval® pyrolysis organic geochemical analyses were performed on a few samples (Crealock, 2004).

RESULTS and DISCUSSION

The iceberg shape (Figs. 2, 3 and 4) was roughly rhomboidal, its dimensions 90 by 70 m, with an average elevation of ca. 5 to 10 m, locally up to ca. 18 m. The cover of loose rock formed an irregular layer from 0 to 0.75 m in thickness.

The majority of the rock fragments were angular, but about 3 percent were rounded pebbles and boulders. A few angular blocks were up to 2 m in size, with the majority less than 0.5 m high. The largest erratic was an angular fossiliferous limestone block of about 12 m³ (Fig. 4). Gravel, sand and dark brown silt were observed in small puddles, and rock fragments were seen lying loosely on the ice surface; few rocks were observed embedded in the ice in deeper layer.

On first inspection 97% or more of the clasts were limestone. Bituminous specimens and well-preserved fossils called our attention and may be over represented. A list and description of all the specimens is available in Harrison and Brent (2002, pp. 72-78) and Crealock (2004).

The nearest candidate outlet glaciers capable of discharging sediment covered icebergs include the immense Petermann and Humboldt Glaciers on the NW Greenland coast (Fig. 1). By comparison, outlet glaciers on the Ellesmere coast are insignificant. These include ice tongues that reach tidewater on Dobbin Bay, Richardson Bay and Rawlings Bay. The north arm glacier of Richardson Bay and the glacier at the head of Rawlings Bay are unlikely sources as these are apt to yield mostly Cambrian sandstone debris (unpublished bedrock geology maps and field observations of U. Mayr and J.C. Harrison, 1998-2000; Harrison et al., 1999). Likewise the Humboldt Glacier is an unlikely source for the iceberg.

Figure 2. Photograph of the rock-covered iceberg on approach by CSSS Louis S. St-Laurent, showing the large boulders in the background and submerged undercut ledges (Photo A.M.Grist).

Figure 3. View of the iceberg on its longest side. Geologists J.C.H. and M.Z. (center left) serve as scale (Photo A.M.Grist).

Figure 4. Sampling the rock on the iceberg. Notice the angular shape of predominantly carbonate blocks (Telephoto A.M.Grist).
source because associated rocks of southern Washington Land and northeasternmost Inglefield Land are either Precambrian shield or Lower and Middle Cambrian clastics and dolostones, respectively (Dawes et al., 2000; Jepsen et al., 1983). All these compositions are exceedingly rare on the iceberg.

The general composition of the iceberg materials and the collected clasts would be consistent with derivation from glaciers emptying into Dobbin Bay and the south arm of Richardson Bay (e.g. Harrison et al., 1999). However, most of the rocks on the Canadian side display a penetrative slaty cleavage, a structural feature not identified in samples from the iceberg. Similarly the eastern Ellesmere outlet glaciers cannot have been a source for the few Precambrian granitoid rocks and gneisses.

The mostly likely source for the iceberg and its sediment load is Petermann Glacier and the adjacent valley outcrops that include Lower to Upper Cambrian strata in the southeast, and Lower Ordovician through Upper Silurian carbonates in the northwest, close to the Hall Basin outlet (Dawes et al., 2000; Jepsen et al., 1983). While shield rocks do not occur at surface on Petermann Glacier, it is likely that the Precambrian shield is present under the inland ice to the southeast (Dawes et al., 2000; Henriksen, 1989). Wisconsinan till carrying shield material could also have been deposited in northern Washington Land and then been reworked and re-deposited by recent incision of Petermann Glacier. Location and examination of lateral moraine on the Petermann Glacier could narrow the provenance of the sampled iceberg to either a northeastern Washington Land source or an adjacent source on the western edge of Hall Land. The organic geochemical analyses of 3 samples of bituminous debris from the iceberg by RockEval® pyrolysis reported by Crealock (2004) yielded a mean T_max value of 426.3°C. These values are comparable to those published by Christiansen and Nørh-Hansen (1989) for rocks of coastal Washington Land and Hall Land. The best match for the debris from the iceberg are strata exposed on either 70 km, essentially the floating frontal match for the debris from the iceberg are strata exposed coastal Washington Land and Hall Land. The best values are comparable to those published by Christiansen and Nørh-Hansen (1989) for rocks of coastal Washington Land and Hall Land. The best match for the debris from the iceberg are strata exposed along the lower 70 km, essentially the floating frontal portion (e.g. Higgins, 1990), of the Petermann Glacier.

Whereas normally icebergs have 7/8 of their volume submerged, rock loaded icebergs could be more deeply or totally submerged. Such neutral buoyancy icebergs could pose a threat to navigation and therefore a risk to the environment in northern waters. If the iceberg sediment load is to some extent the consequence of rock fall and landslide activity onto floating glacier ice prior to calving, or if rock-loaded icebergs result from higher than usual rates of calving, it is important to know whether this activity is on the increase as a result of climate change. Higgins (1991)

has described large lateral moraines on the Petermann Glacier, and calculated an average calf ice production to be 0.59 km^3/y - one of the highest in the region - and subject to multi-year cycles of accelerated calving.

The hypothesis that the Petermann Glacier is producing unusual icebergs should be investigated by comparing archival air photographs and recent satellite images of some representative glaciers, as was done by Higgins (1991). Warming and increased precipitation during the spring of 2001 may be accountable for the production of rock loaded and low profile icebergs in the area. The precipitation in the Thule airbase, NW Greenland during July and August 2001 exceeded 8 inches of rain (Weather Station Operations, Thule Air Base, Greenland). On August 2nd, 2001, the river that flows through the Thule airbase flooded due to persistent and heavy rains, destroying culverts, bridges and pipelines, effectively closing the airport for several days (e.g. Jackson et al., 2002). Long term Thule residents had never experienced such a flooding in at least 50 years. In the course of field work in NW Greenland during early August 2001, the Nares Strait Geocruise 2001 expedition scientists came across knee-deep thawed out permafrost, and red sediment loaded glacial outflow streams. It is likely that such conditions enhanced rock falls and landslides on steep valley walls, or may have simply accelerated calving (e.g. Higgins, 1991) at the Petermann Glacier, hauling anomalously large loads of lateral moraines and debris into the Nares Strait.

The northwest Greenland glaciers are known to produce thousands of icebergs annually but generally only small percentage reach lower latitude and become hazardous to shipping (Murray, 1969). The southerly flowing Ellesmere and Baffin Bay currents are responsible for transport and deterioration of icebergs on their journey to the Labrador Sea. It is unlikely that such a rock covered iceberg will drift to far from its source before rolling and dispensing loads of erratics to the ocean floor. However, the dense and equally distributed cover of rock may protect the ice from solar and wave erosion, thus insulating and stabilizing the iceberg, allowing it to survive longer in open water.

Width to height ratios of above water characteristics has been shown by Allaire (1972) to accurately predict iceberg stability in the field (e.g. Robe et al., 1977); stability of this tabular iceberg is assumed to fall below the minimum stable ratio 6:1, suggesting unstable conditions. However photos of the iceberg indicate no physical characteristics commonly observed on unstable icebergs: the sediment load still rest on the surface of the ice, and undercutting caused by wave erosion are present only on some faces and are parallel to the water line (Fig. 2) indicating that no
tilting or rolling have occurred (the iceberg has been sinking). Open water deterioration and erosion probably significantly decreased the size of this iceberg within weeks. Yet it is important to recognize this occurrence such that other exploration and north traveling vessels are aware of this potential danger. For instance, Tangborn et al. (1998) reported the collision between the oil tanker Overseas Ohio and a rock-covered, neutral-buoyancy iceberg in Prince William Sound, Alaska, in January 1994, which could have become an environmental disaster of magnitude. Furthermore, in 2002, rock-covered icebergs, believed to originate in northwest Greenland, were observed near the coast in offshore Newfoundland (personal communication, March 17, 2004, Ingrid Peterson, Coastal Ocean Science, Bedford Institute of Oceanography, Fisheries and Oceans Canada).

CONCLUSIONS
1. Preliminary analysis of rock debris carried by the tabular iceberg in question indicates calving of the iceberg from Petermann Glacier, which drains the Greenland ice sheet, and derivation of the related sediment load from adjacent outcrops of either northern Washington Land or western Hall Land.
2. Circumstantial evidence allows speculating that heavy rains and relatively warmer summer temperatures in northwest Greenland in 2001 may have led to fast breakup of the floating ice in the Petermann Glacier carrying anomalous amounts of rock debris from lateral moraines and recent products of mass wasting.
3. Tabular, rock-loaded, near neutral buoyancy icebergs from rapidly calving glaciers may possibly pose a threat to navigation in northern waters and should be further investigated.

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THE INFLUENCE OF REGIONAL WARMING ON THE TREELINE OF A SUBARCTIC MOUNTAIN RANGE – A FIRST APPROACH TO FIELD RESEARCH

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ABSTRACT

Recent climate change is evident but it is not necessarily due to global warming because ground temperatures differ regionally. Since subarctic regions are very sensitive they are a good object of investigation, especially in mountainous areas where life is at its limits. One possible indicator of global warming is vegetation. Investigating the vegetation changes across subarctic zones can be a difficult venture, because there is no exact limit, but a latitudinally wide zone of transition which differs due to pedo-hydrological factors. So it was decided to examine altitudinal zones, especially in arctic and subarctic environments because there they are reduced to only a few tens of meters, and it might be possible to probe the treeline for clues of global warming.

On the Kola Peninsula the increase in ground temperature is provable for the last 30 years. Field works were carried out to point out that there is a reaction of the vegetation on the regional warming at the treeline in the Chibiny subarctic mountain range.

The species Betula tortuosa forming the timberline as well as the treeline was observed during the investigations. Preliminary results show that there is an remarkable increase of younger trees in the ecotone between both mentioned lines. The main question is to find a real link between the warming and the increase of species at the tree line ecotone. The ongoing investigations will focus on site heterogeneity with special reference to soil, climatic conditions and dendrochronology.

INTRODUCTION

The transition from the early Holocene until present is well examined and we now have another transition, which may not take a thousand years, but may occur within a few tens of years. Many researchers postulate a circum-polar expansion of the forests (i.e. Cramer, 1997) that is based on rising temperatures. Something seems to be happening concerning warming in view of articles by Cotton and Pielke (1995),

![Figure 1: Global temperatures from 1860 to 2001 (IPCC 2001:3).](image-url)
Houghton et al. (1995), IPCC (1997), Angell (2000), Harvey (2000) and Jones et al. (2000). Though the cause may be slightly controversial—it is the consensus that mankind is contributing to global warming of the atmosphere (von Storch and Flöser, 1999). However, due to the short duration of this change, its cause is not possible to reliably prove. Because of the complex interaction between the atmosphere and water in its different states, warming is not the same everywhere (Watson et al., 1998). This is especially true in the Arctic region (Wadhams et al., 1996; Oechel et al., 1997)—and particularly in Northern Europe and the

Figure 2: Contribution of the potential-natural vegetation in Northern Europe and Northwest Russia (Model BIOME 1.1) A: simulated on the basis of actual climate data. B: simulated on the basis of actual climate data with a $3\times$CO$_2$-scenario (from: Cramer, 1997, slightly modified).
Kola Peninsula—because in these regions elevation of surface temperatures is not observable everywhere. The Gulf Stream or the North Atlantic Drift, is a crucial factor (Rahmstorf, 1997).

This ocean current causes ice-free winter conditions in the harbour of Murmansk, and therefore also influences the surface temperature on the Kola Peninsula. By comparison, Hudson-Bay in Canada, which is considerably more southward, and even East Greenland, are completely ice-covered during winter, but in the Kola region the polar ice lies just at the eastern coast of the peninsula and penetrates the White Sea. The strength and temperature of the North Atlantic Drift are not constant due to changes in the thermohaline circulation between Greenland and Newfoundland. Although the complex mechanisms are not completely understood (Rahmstorf, 1995; 1996), it appears that if there is a higher influx of freshwater from melting ice in North America, less of the heavy saltwater sinks in the North Atlantic Ocean. Therefore, the warm surface water from southern areas cannot flow northward. The result is to cut off the warm surface water from the Norwegian and Barents seas. When this southward shift in the warm surface water happens, it seriously influences temperatures in Northern Europe and on the Kola Peninsula.

Investigations in the last 30 years show increasing temperature in the northern hemisphere. The intense increase from the middle of the 19th century to about 1940 is clearly shown in Figure 1. It is followed by a decrease up until the 1960s and another continuous increase to date. These changes are also shown in the
regional focus or by a quarterly examination: the warming is observable from Alaska, Northwest Canada and Scandinavia to the Yamal Region in North Russia, as well as, in every quarter since 1970 (compare Maxwell, 1997; Jones et al., 2000).

The obvious question is, what effects warming has on the boreal (coniferous) forests—as Kellomäki et al. (1997) asked by entitling their publication “More timber from boreal forests under changing climate?” The “Potsdam-Institut für Klimafolgenforschung” (PIK – Potsdam Institute for Climate Impact Research) follows the same line which is shown in Figure 2. The simulated borderlines based on actual climate data shown in Figure 2A are consistent with our own investigations. But, due to edaphic and hygrical conditions, these zones are quite difficult to investigate.

Nevertheless, assuming this trend is real, the question arises about the nature of the consequences, especially in mountain-ecosystems. There is no clear evidence at this time. The PIK investigations indicate effects but are very cautious about a forecast due to the uncertainty of the climate models (Plöchl and Cramer, 1995; Cramer, 1996; and Lindner, 2000). The recent investigations of Lucht et al. (2002) using the “LPJ Dynamic Global Vegetation Model” show the incremental increase of vegetation in the boreal zone during the last 20 years.

Which changes are to be considered? Beniston (2000) shows a dramatic scheme of possible changes in Figure 3. This example is for the Alps assuming the increasing temperature 3.5°C for the next 50 years. Applied to the 0.2-0.5°C of the last 30 years in the Kola region, the result might be less drastic. By modeling North Europe and parts of Russia, similar small-scale changes can be shown, compared to Figure 2. Most important for such investigations is the timberline ecotone which is a dispersing area where the dense forest breaks up and changes gradually into tundra. This timberline ecotone is a “combat zone” where individuals or even species die out or immigrate according to the actual changes in climate—the more extreme the area of investigation, the more distinct the changes. Subarctic mountain ranges have very extreme climate and light conditions with a timberline at about 400 m above sea level (a.s.l.) and a vegetative period between May and September.

MATERIAL, METHODS, AND AREA DESCRIPTION

The Chibiny Mountain Range located in the middle of the Kola Peninsula (Fig. 4) arise rather steeply from the low undulating ground moraine at 200 m a.s.l. to 1200 m a.s.l. Even though this area is situated clearly north of the Arctic Circle within the

Fig. 6: The mean annual temperature of the meteorological stations Murmank and Kirowsk.
zone of the Northern Taiga, it is too far south of the coast to be influenced by the Gulf Stream.

The transitions of the Northern Taiga—here a mixed forest dominated by spruce, to forest tundra dominated by birches, to tundra, are quite obvious. Figure 5 shows the described area and examples of tundra, forest tundra, and Northern Taiga.

Sites 1, 2 and 3 were chosen because they are areas with south-east and south-west aspect and get the longest period of light and warmth. Site 1 shows a typical situation in the timberline ecotone. While climbing up one leaves Northern Taiga with spruce, to come into dense forest dominated by birches. With increasing altitude this forest becomes lower and less dense and is followed by an open ecotone with a distinct borderline. Within this region the possible changes might show clearly (cf. fig. 8 and 9).

Small trees or bushes, especially Betula tortuosa, Populus tremula, Salix reticulata, resp. Salix ssp., sporadic Sorbus aucuparia and Juniperus communis make up the tree layer, whereas Betula nana (partly or completely covering), Empetrum hermaphroditum, Vaccinium myrtillus, Calluna vulgaris and sporadic Vaccinium vitis-idaea make up the field layer. The site is situated on rough debris and is therefore mainly dry. The area is basically characterized by raw soils and podbur. The timberline is at about 430 m a.s.l.

At Site 2, the slope is not steep compared to Site 1 and the vegetation seemed fresher. The soils were mainly podzols without any coniferous trees, which is usual with podzols. This will be discussed further on. This site unlikely has any anthropogenic influence (felled conifers and some relict podzol). Beside the fact that there were no trunks, it is unlikely that there were any trees since it is situated at 490 to 540 m a.s.l., 50 to 100 m above the timberline.

Site 3 is different from the other sites. Changes should be seen in the species of spruce in investigations below the timberline. There, the plant community Piceetum hylocomiosum, being typical Northern Taiga according to Walter & Breckle (1994:481), should be found. Distinctly below the timberline the spruce thin out gradually. It is possible to eliminate anthropogenic/zoogenical usage of the timberline ecotone by reindeer in these parts of the Chibiny Mountain Range (Sites 1-3). There are no longer any reindeer, and not even excrement could be found.

To judge the regional warming we looked at data from the meteorological stations of Murmansk (northern part of the Kola Peninsula) with about 22,000 data records and from Kandalakscha (southern part) with 30,000 data records (which are not entirely analysed). In the summer of 2002, climate data could be provided directly out of the investigation area. Figure 6 shows recordings of the avalanche station of the former combine “Apatit”, with weather stations in quarries.

The Murmansk data show an obvious decrease in temperature until 1970, followed by an increase (using non-linear regression)—a warming trend of more than 1°C. The Kirowsk data show generally lower temperatures due to the location of the station at approximately 900 m a.s.l., but indicate a warming of about 0.8 °C. The fact that the average annual temperature never sinks below certain values—since 1977 Murmansk is never below -4°C and since 1971 Kirowsk was not below -6°C—is very important because the absence of extreme cold has formed better conditions for vegetation for the last 25-30 years.
North of the Arctic Circle the altitudinal zones spread only a few decameters vertically (Fig. 7). Methods like dendrochronology together with site-investigations should be able to document these changes. If there were shifts, they must have taken place within tens of meters, not hundreds of meters.

The developments being considered are taking place within the fringe of the original timber- and treeline.

By using the phrase “timber-/treeline” the authors do not intend to revisit the discussion about definitions of timberline or treeline as “combat zone” (Heikkinen et al. 1995:7). Tranquillini (1979), Treter (1984), and Kullmann (1990) discussed the problem without finding a resolution. Regardless, the differences between timberline and treeline seem to be clarified for the purposes of our discussion.

The process of timberline-procession can be described as a narrow compaction of trees below the treeline. Scott (et al., 1997) investigating actual germination of spruces, show that warm summers cause added seed formation and germination of spruces (Picea glauca), which can be correlated to the post-Little Ice Age phases. If these warm summers continue, more plants can survive. The warm periods in the late 1960s and early 1970s could have caused an increase at the timberline that is now near 30 years old and should be visible. Figure 7 shows a photo taken in the Chibiny Mountain Range in August 2002. It cannot verify the increase but shows similarities to the situation described by Scott (1977).

Many investigations of the changes at the timberline are mostly about the Holocene and therefore do not deal with post-glacial changes (Frenzel et al., 1993; Eronen et al., 1999; and MacDonald et al., 2000). Other investigations are showing an increase of the timberline due to climatic changes such as MacDonald et al. (1998) and Hicks (1993) using pollen analysis and δ¹³C/δ¹⁸O method investigations in Finnish Lapland, and MacDonald et al. (1993) and Wolfe et al. (1996) in Canada. Global climatic changes influencing forestry were investigated by Kellomäki and Kolström (1994), White (1994), and Jarvis (1998). They also deal with measurement of timberland and forest harvesting.

Heikkinen et al. (1995) showed that the summer temperature and the length of the vegetation period influence the position of the timberline. Therefore, increasing summer temperatures and longer vegetation periods can cause the advance of the timberline. Hofgaard (1997) advises not to

<table>
<thead>
<tr>
<th>Perimeter</th>
<th>Age</th>
</tr>
</thead>
<tbody>
<tr>
<td>&lt;2-5 cm</td>
<td>~ 3 a</td>
</tr>
<tr>
<td>&gt;5-10 cm</td>
<td>~ 7 a</td>
</tr>
<tr>
<td>&gt;10-15 cm</td>
<td>~ 15 a</td>
</tr>
<tr>
<td>&gt;15-20 cm</td>
<td>~ 20 a</td>
</tr>
<tr>
<td>&gt;20-25 cm</td>
<td>&gt; 30 a</td>
</tr>
</tbody>
</table>

Table 1: Circumference-age-relation with *Betula tortuosa*, sampling site 1.
overestimate the climatic change effects. Her investigations in Norway show an advancing timberline due to recessive grazing instead of climatic causes. This example shows the importance of recognising the anthropogenical factors. Mattsson (1995) gives similar information concerning anthropogenical involvement but shows a withdraw of the timberline caused by extensive grazing, tree felling and anthropogenic caused forest fires.

The following question still remains unanswered: “Is it possible to show the actual verifiable climatic change via the timberline?” The information given in literature range from “impossible” (Holtmeier, 1993) to detailed descriptions of changes that took place (Huntley, 1991; Grabherr et al., 1994; Kellomäki and Kolström, 1994; Hofgaard, 1997; Beniston, 1998), which might be depend on the particular view and questions of the scientists. This short synopsis might be unsatisfactory but shows the urgent need of further investigation. Frenzel et al. (1993) and Holtmeier (1994) also point out the lack of research.

RESULTS

Vegetation Growth

In the opinion of the authors, changes at the alpine timberline are easier to detect than the polar timberline because of the reduced altitudinal zones that might show the effects in a narrow space. Apart from pasturing, anthropogenic influence might be less since tree felling and fires are quite infrequent in the mountains.

This treeline presents a situation indicating incremental growth, as shown in Figure 8. There are adult birches (Betula tortuosa) which made up the former treeline. In recent years, young growth (different age classes) appeared around them as well as above. Obviously, conditions have changed allowing the growth of birches above the original treeline. The young birches are about 3-6 years old.

To quantify the increase in vegetation, five measuring-tapes were laid parallel through the trees at Site 1 – from the timberline to the treeline 50 m above. Along the measuring-tapes all individuals of Betula tortuosa were mapped along with their circumference. It is not possible to deduce the absolute age from the circumference because there were only a few age determinations per age class (see Table 1). Nevertheless, a trend is shown—the thicker the older. Furthermore, the young incremental growth differs from older individuals in the treeline ecotone.

Figure 9 shows the results of the mapping. The frequency of younger and older individuals close to the timberline is distinct. The older individuals diminish from the middle of the transect. The increase of young birches above the endmost adult birch is remarkable.

It was not possible to map the Picea abies ssp. obovata at their species limit at sampling Site 3 in the same way as the birches were mapped at theirs. The
Cyffka and Zierdt

Fig. 11: Extreme increase in aspen above the timberline at site 2. Photo: B. Cyffka, 24. 08. 2002.

Fig. 12: Podbur at site 1, above the timberline. Photo: B. Cyffka, 22. 08. 2002.

Birch forest has been too dense with only a few widespread spruces in between. Therefore, a new scheme was applied to describe the distribution of the spruces (Fig. 10).

At about 420 m a.s.l., which is the normal occurrence of *Picea obovata*, there are many adult and juvenile individuals. Similar to *Betula tortuosa*, the mapping shows a significant decrease of adult spruces (in only 15 m distance), which are associated with an equal stand of young spruces. This indicates an increase along the species boundary. Furthermore, these spruces show no signs of being crippled, whereas it is quite usual for individual trees in extreme locations to be crippled.

Rather unintentionally the increase of aspen, *Populus tremula*, was analysed. The sampling of 1-2 year old trees is observable: below the timberline at forest clearances as well as above in the timberline ecotone. At site 2 at 450 m above sea level, there were 75, 1-2 year old, aspens counted in 9 square metres (Fig. 11). Aspen are often indicated as a pioneer tree.

Whether these aspen accumulations at Site 2 are an anomaly or caused by warming is not yet known. Nevertheless this is another piece of evidence concerning the vegetation increase.

The amount of increase at the different sampling sites:

Sampling site 1: Increase in trees, mainly *Betula tortuosa*, *Populus tremula*, and *Salix ssp.* above the timberline and partly also above the former treeline

Sampling site 2: No real increase, partly in *Populus tremula*, below the timberline

Sampling site 3: Increase in *Picea obovata* at its species limit as well as an increase in *Populus tremula* and *Betula tortuosa* above the timberline.

Soils

The reason for the differences in these sites is not obvious at the first glance because the climate and location of the sampling sites on the slopes were nearly the same. However, there is a “deeper” site condition to be observed—the soil. The main soil type of the Northern Taiga is podzol; in forest tundra and tundra, raw soils with development up to minor arctic brown soils (podbur) are found. So podzols at Site 3 and podburs (Fig. 12) at Site 1 were trenched. Even in the dense birch forest below the timberline only podburs were found. Some marginal differences in the soil chemistry are shown in Table 2, which does not indicate a podzolisation. The attention was turned to the iron-mobilisation. Therefore, the Fe has been solubilized in a pH 4 buffered extract and diagnosed with ASS. The Fe-content of Site 1 was low with differences between the soil profile above the

Table 2: Analyses of iron content of two podbur profiles at sampling site 1.

<table>
<thead>
<tr>
<th>Horizon</th>
<th>Extract buffered at pH 4 [mg/kg]</th>
<th>Extract in H₂O [mg/kg]</th>
</tr>
</thead>
<tbody>
<tr>
<td>Above timberline (forest tundra)</td>
<td>O 2</td>
<td>2</td>
</tr>
<tr>
<td>A₁h</td>
<td>2</td>
<td>2</td>
</tr>
<tr>
<td>BC</td>
<td>18</td>
<td>1</td>
</tr>
<tr>
<td>Below timberline (birch forest)</td>
<td>O 18</td>
<td>2</td>
</tr>
<tr>
<td>A₁h</td>
<td>26</td>
<td>2</td>
</tr>
<tr>
<td>BC</td>
<td>30</td>
<td>1</td>
</tr>
</tbody>
</table>
timberline and the soil profile of the birch forest. The
birch vegetation works as acid donor. Nevertheless,
the soils are classified (chemically and diagnostically)
as podburs. Essential is the low iron mobilization.

Site 2 shows unexpected results. In the upper parts
of the timberline ecotone above the treeline at 520 m,
530 m and 540 m a.s.l., trenching revealed that pozols
have developed (Fig. 13). Podzols developed without
dense vegetation or without coniferous trees contradicts
findings at other sites and leads to the conclusion that
this must be a relict podzol. But it is not possible for

Table 3: Analyses of iron content of three podzol
profiles at sampling site 2.

<table>
<thead>
<tr>
<th>Horizon</th>
<th>Extract buffered at pH 4 [mg/kg]</th>
<th>Extract in H2O [mg/kg]</th>
</tr>
</thead>
<tbody>
<tr>
<td>Uppermost position in forest tundra</td>
<td>O-Ah 18</td>
<td>4</td>
</tr>
<tr>
<td></td>
<td>E 27</td>
<td>5</td>
</tr>
<tr>
<td></td>
<td>B 9</td>
<td>1</td>
</tr>
<tr>
<td></td>
<td>C 6</td>
<td>1</td>
</tr>
<tr>
<td>Middle position in forest tundra</td>
<td>O-Ah 25</td>
<td>7</td>
</tr>
<tr>
<td></td>
<td>E 31</td>
<td>4</td>
</tr>
<tr>
<td></td>
<td>B 48</td>
<td>1</td>
</tr>
<tr>
<td></td>
<td>C 20</td>
<td>1</td>
</tr>
<tr>
<td>Lower position in forest tundra</td>
<td>O-Ah 17</td>
<td>5</td>
</tr>
<tr>
<td></td>
<td>E 34</td>
<td>3</td>
</tr>
<tr>
<td></td>
<td>B 66</td>
<td>2</td>
</tr>
<tr>
<td></td>
<td>C 29</td>
<td>1</td>
</tr>
</tbody>
</table>

The podsolization at this site may be due to the
terrain situation. As opposed to Site 1 this site has a
southeast exposure with moderate declination and a
good water supply – described as “fresh” in geobotanic
terms. The timberline is visibly irregular. The degree of
coverage of the field layer is close to 100 % with a high
share of Empetrum hermaphroditum and Vaccinium
vitis-idaea, but also Vaccinium myrtillus and Betula
nana. It is likely that these species are able to produce
enough acid to start the podzolisation.

The soil chemistry of Site 2 is different than the
other sites (Table 3). The podzols observed in the field
proved chemically to be degenerating podzols. This is
indicated by the fact the B-horizon in the uppermost
sample is not the one with the highest Fe-share. Rather
the highest iron content is in the Ah through E-
horizons. The reasons are still unexplained. If the forest
had increased, then it is possible that the suddenly
missing vegetation could be responsible for
this result. Whether dense crowberry- and cranberry-
cover can cause this result must be investigated.

Finnish Site

On the way back from the Chibiny Mountain
Range there was an opportunity to study similar
phenomena in Finnish-Lapland. The location of
investigation (Figure 15) is the only natural spruce
existence around Ivalo/Saariselkä. The site is situated
about 2 km west of Kaunispää and species range from
the typical Piceetum hylocomiosum to a distinct birch
forest up to the timberline at the heights of about 420 m
a.s.l. in hilly terrain. Neither the soils (Table 4) nor the

Table 4: Analyses of iron content of two exemplary
podzol profiles from the Northern Taiga in Finland,
near Ivalo (investigations not shown here in detail).

<table>
<thead>
<tr>
<th>Horizon</th>
<th>Extract buffered at pH 4 [mg/kg]</th>
<th>Extract in H2O [mg/kg]</th>
</tr>
</thead>
<tbody>
<tr>
<td>Birch forest</td>
<td>O-Ah 6</td>
<td>4</td>
</tr>
<tr>
<td></td>
<td>E 14</td>
<td>2</td>
</tr>
<tr>
<td></td>
<td>B 83</td>
<td>10</td>
</tr>
<tr>
<td>Spruce forest (Northern Taiga)</td>
<td>O-Ah 8</td>
<td>5</td>
</tr>
<tr>
<td></td>
<td>E 21</td>
<td>3</td>
</tr>
<tr>
<td></td>
<td>B 100</td>
<td>5</td>
</tr>
</tbody>
</table>
Fig. 14: Terrain situation at site 2: timberline ecotone at a slope with moderate declination towards south-east exposition, and with fresh vegetation. Photo B. Cyffka, 23. 08. 2002.

Fig. 15. Location of Finnish sampling site 2 km west of Kaunispää.

Fig. 16: Increase in of Picea abies ssp. obovata in the birch forest of the Finnish site.

vegetation gave comparable results with Russia. They raised more questions than answers. Typical podzol profiles were found in the spruce forests at about 320 m a.s.l., as well as, in the birch forest at about 360 m a.s.l. in a north-east exposure. Iron-eluviation in the E-horizon and iron-enrichment in the B-horizon appears in both cases. The strong development of these soils in the birch forest is somewhat unexpected because generally the deepest podzols are supposed to be below acid coniferous litter.

In the birch forest there was a distinct increase of Picea abies ssp. obovata very close to the actual timberline (Fig. 16). This is all the more amazing because in a lower altitude there was a spruce forest with less of an increase of young spruce and surprising there was no young birch growth at all. No birches were found in either of the Finnish sites which were younger than 5 years. The reason is probably the intense reindeer-grazing all over Finnish-Lapland. Presumably the reindeer prefers the birch seedlings to the spruce seedlings.

Using the treeline and timberline for understanding climate becomes problematic to impossible in Finnish-Lapland because of the anthropogenical/zoological influence. To compare the neighbouring sites is quite difficult as well. The slightly hilly mountain tundra in Finnish-Lapland differs from the mountain range of the Chibiny Mountain Range in Russian-Lapland in climate and in the pedo-hydrological conditions. Furthermore, the species spectrum differs--while in Russia the main species are Picea abies ssp. obovata and Betula tortuosa, the cardinal species in Finland are Pinus sylvestris and Betula pubescens. The latter hybridize very intensely and therefore, there are many varieties that do not seem to react to climate warming. The timberline has a southwest exposure at 430 m a.s.l. about 20 km south of Saariselkä. Betula pubescens var. apressa and some sporadic adult trees of Pinus sylvestris form this line (Fig. 17) at the Kiilopää mountain (546 m a.s.l.). There is definitively no observable altitudinal increase in the forest.
timberline. Whether this is caused by the missing species-specific reaction to warming, or reindeer grazing up the young growth is still unknown.

CONCLUSIONS
In retrospect it is easy to see the still unsolved problems which have been shaped more precisely because of these first investigations. Further research can be more goal-oriented. The following results can be worked out:

• There is an increase in trees above the timberline and partly also above the former treeline in the Chibiny Mountain Range. Further investigations are needed to see if this increase is due to the latest phase of global/regional warming (since 1965).
• The increase differs regionally. There seems to be a dependence on the species and site conditions, as well as, the anthropological/zooological influence by reindeer grazing:
  o In Russia Betula tortuosa, Populus tremula, and Picea obovata show a clear increase
  o In Finland B. pubescens, B. pubescens var. appressa, and Pinus sylvestris show no increase, while Picea obovata shows some increase, but this species is no longer the main tree species due to forestry.
• Soils are important to characterize site conditions. But there are many new questions connected with them (e.g. podzol above timberline).
• All tentative results show no model is able to predict what really happens at the timberline.

REFERENCES


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Dartmouth, Nova Scotia, Canada
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