Cretaceous–Tertiary magmatic and tectonic events in North Greenland and the history of adjacent ocean basins

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Soper, N. J., Dawes, P. R. & Higgins, A. K. 1982. Cretaceous–Tertiary magmatic and tectonic events in North Greenland and the history of adjacent ocean basins. – In: Dawes, P. R. & Kerr, J. W. (eds), Nares Strait and the drift of Greenland: a conflict in plate tectonics. – Meddr Grønland, Geosci. 8: 205–220.

Greenland emerged as a separate entity during the break-up of the Laurasian plate in Palaeogene time. Research in the last decade has revealed some of the magmatic and tectonic events which were associated with the development of Greenland's northern margin: late Cretaceous basic dyke swarms, an explosive volcanic province (Kap Washington Group) dated as end-Cretaceous, a thrust zone of late Paleocene – Eocene age and a later Tertiary period of fault reactivation related to regional uplift. The north-eastern (Wandel Sea) margin was affected by compressional and extensional deformations.

The evidence for this sequence of events is outlined together with chemical data which characterise the igneous activity as alkalic, of 'within-plate' type. A coherent pattern can be discerned between the onshore geological history and the spreading history of adjacent ocean basins, when the revised polarity time scale of Hailwood et al. (1979) is used as a link. We show how the geological evidence from North Greenland provides a number of important constraints on the timing and magnitude of displacements between Greenland and Canada in the Labrador Sea and Baffin Bay and also along Nares Strait.

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In North Greenland a Proterozoic – Lower Palaeozoic platform sequence passes northwards into a clastic trough which was subjected to deformation and metamorphism in mid-Palaeozoic time. This comprises the eastward extension of the Franklinian mobile belt of Arctic Canada. Superimposed on the Franklinian belt are structures of Tertiary age which delimit an important regional tectonic zone extending from the Canadian Arctic Islands (Eurekan orogeny: Thorsteinsson & Tozer 1970, Trettin et al. 1972) through North Greenland to Svalbard (West Spitzbergen orogeny: Harland & Horsfield 1974). The deformation in this zone is only locally intense. The Canadian and Svalbard segments have been extensively studied, but the Tertiary events in North Greenland have been documented only in the last decade. In 1969 a volcanic province (Kap Washington Group) was discovered on the north coast of Peary Land (Dawes & Soper 1970, 1973) and subsequently dated isotopically at about the Cretaceous–Tertiary boundary (Larsen et al. 1978). The volcanic rocks are in thrust contact with Palaeozoic metasediments of the Franklinian trough, and this important phase of compressive deformation perpendicular to the North Greenland continental margin is also of Palaeogene age. Weaker structures parallel to the Wandel Sea margin have been shown to affect the youngest (Paleocene) sediments of the Wandel Sea Basin cover sequence (Troelsen 1950, Håkansson 1979, Håkansson et al. 1981), and other volcanic and tectonic features of Tertiary age have come to light during the systematic mapping programme which is currently being undertaken by the Geological Survey of Greenland.

It is evident that these events affected the North Greenland continental margin during the inception and development of the adjacent ocean basins in late Mesozoic and Cenozoic time (Fig. 1). The spreading geometry of the Labrador Sea, Norwegian Sea and Eurasia Basin has been investigated intensively in recent years by magnetic anomaly matching (Kristoffersen & Talwani 1977, Talwani & Eldholm 1977, Phillips et al. 1978, Srivastava 1978, Feden et al. 1979), but incompatibilities exist between the various models proposed and no unique geometrical solution has emerged for the development of the oceans around Greenland in terms of rigid-plate theory. It is here that geological evidence of within-plate deformations can provide useful constraints on displacement models. The prime example is the very subject of this symposium — Nares Strait. Since the geological evidence from the lands bordering the Strait allows only tens rather than hundreds of kilometres wrench displacement (Christie et al. 1981, Dawes & Kerr, this volume) models which

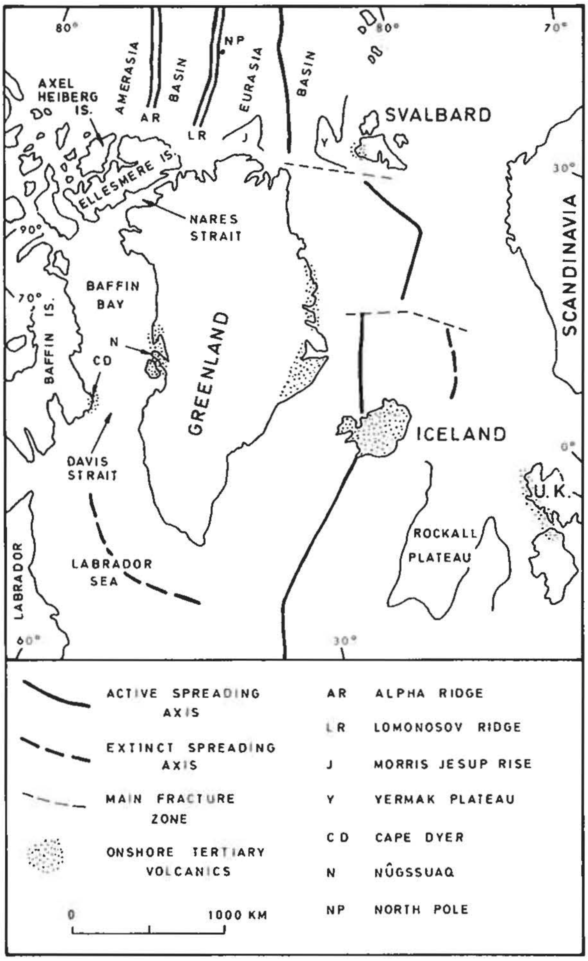

Fig. 1. Index map showing position of Greenland in the North Atlantic – Arctic Ocean system. Outline of submarine plateaux demarcated by 1000 m isobath.

require very large displacements can be eliminated. On the other hand, reconstructions that allow no movement along Nares Strait and treat the North American and Greenland continental masses as entirely rigid result in unacceptable overlap of continental rocks in some part of the North Atlantic region. In addition, credible plate tectonic models must provide appropriate geotectonic settings to account for geological features which would otherwise remain unexplained — for example the existence of an explosive alkalic volcanic province on the north coast of Greenland.

In order to relate the onshore geological record to the spreading history of adjacent oceans, a chronological framework is required and this is provided by the polarity time scale. By the mid-70s it had become apparent that the Heirtzler et al. (1968) time scale assigned ages to magnetic epochs in the Palaeogene which were discrepantly high by about ten per cent. The studies of spreading geometry listed above continued to use the Heirtzler scale and thus relationships with geologically dated events are not always apparent. Among the revised time scales recently proposed we adopt that of Hailwood et al. (1979) which is based in the Palaeogene on a biostratigraphically controlled reversal stratigraphy from North Atlantic sediment cores.

In this paper we show that the Cretaceous–Tertiary geological record in North Greenland and the spreading history of the adjacent ocean basins, in so far as they are presently understood, form a coherent pattern if viewed in the light of the revised polarity time scale, and that the parallel lines of evidence provide useful constraints on the interpretation of each other.

Cretaceous–Tertiary events in North Greenland

Much remains to be learned of the post-Palaeozoic history of North Greenland, but there is sufficient evidence to identify several major magmatic and tectonic features of late Mesozoic–Tertiary age (Fig. 2).

Basic dyke swarms

Dolerite dykes are extensively developed in the North Greenland fold belt where they cut the mid-Palaeozoic structures and they occur less frequently on the platform to the south (Ellitsgaard-Rasmussen 1955, Jepsen 1969, Dawes & Soper 1970). These dykes are restricted to the eastern part of the fold belt, extending as far west as the vicinity of Victoria Fjord at long. 48°W. In the fold belt approximately coast-normal dykes predominate (NNE to NW) and there is a marked concentration in a 100 km wide zone of the north coastal region (Fig. 2). On the islands in the central part of this zone the proportion of northerly-trending dykes reaches 20 per cent; large inclined dolerite sheets and scattered E–W-trending dykes also occur here. The northerly swarm dies out southwards towards the Harder Fjord fault zone, to be replaced in and south of the fault by WNW–ESE and NW–SE swarms of en echelon dykes extending SSE for about 120 km (Higgins et al. 1981). Some dykes in the Frederick E. Hyde Fjord area are aligned in the fault zone.

The dykes are normally straight discordant dolerites with chilled margins, usually ranging up to 25 m in thickness but occasionally attaining 200 m. Particularly towards the north coast there is evidence of shearing and retrogression (Dawes & Soper 1979), and within a few hundred metres of the Kap Cannon thrust the dykes develop schistose fabrics and greenschist-facies assemblages.

Petrographically the undeformed dykes are ophitic textured dolerites and olivine dolerites, occasionally feldsparphyric or vesicular. Samples occur with relatively fresh olivine, but there is a gradation in the degree of alteration to complete replacement of the olivine by

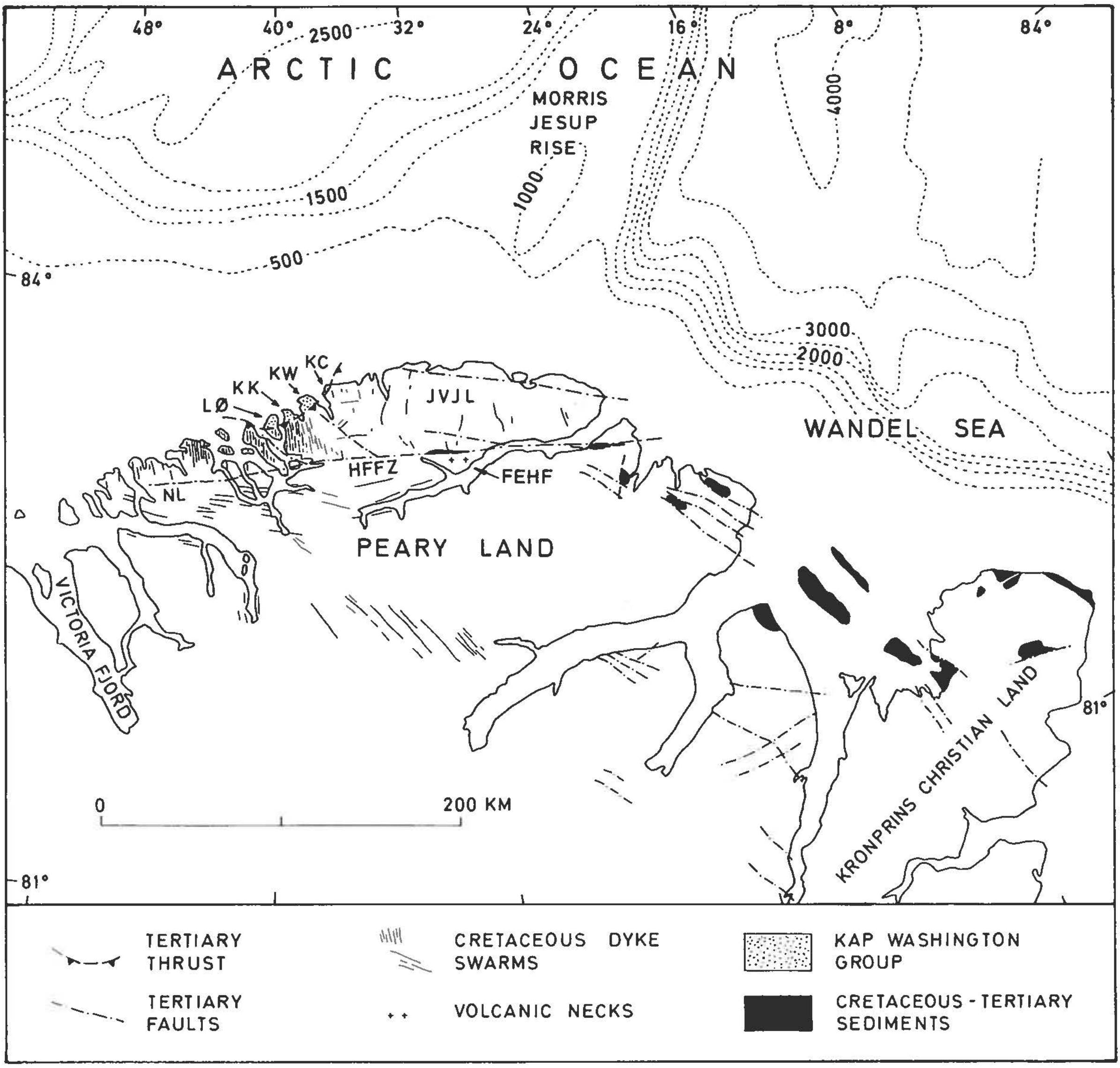

Fig. 2. Cretaceous–Tertiary magmatic and tectonic features of the northern margin of Greenland. Bathymetry in metres from Johnson et al. (1979). NL, Nansen Land; JVJL, Johannes V. Jensen Land; LØ, Lockwood Ø; KK, Kap Kane; KW, Kap Washington; KC, Kap Cannon; HFFZ, Harder Fjord fault zone; FEHF, Frederick E. Hyde Fjord. Data from the authors' field work and from references quoted in the text.

magnetite, iddingsite, etc. A single pyroxene is present, evidently a strongly titaniferous augite, and both the pyroxene and feldspar usually show some degree of alteration. Ilmenite is abundant in many samples.

A selection of major and trace element analyses of dykes is given in Table 1; average compositions of dyke swarms and sills are given in Table 2. Here, comment is confined to the essential chemical characteristics of the suite. Its alkalic character is indicated by exceptionally high TiO_2 (usually > 4 per cent), high Na_2O + K_2O (> 4 per cent) and high P_2O_5 and total iron. All analyses fall in the alkalic field of an alkali–silica plot (Fig. 3) although there is a scatter extending into the subalkaline field as defined by Irvine & Baragar (1971). All the dykes are olivine-normative; some contain nepheline and hypersthene in the norm. This is thought to be a consequence of alkali and silica mobility during alteration. The alkalic affinity of the suite is confirmed by the immobile element distribution. For example, on a TiO_2–Zr/P_2O_5 plot, adopted by Floyd & Winchester (1975) for the discrimination of alkali and tholeiitic basalts, the 40 dykes for which Zr has been determined show a pronounced 'vertical' trend characteristic of alkalic compositions (Fig. 4).

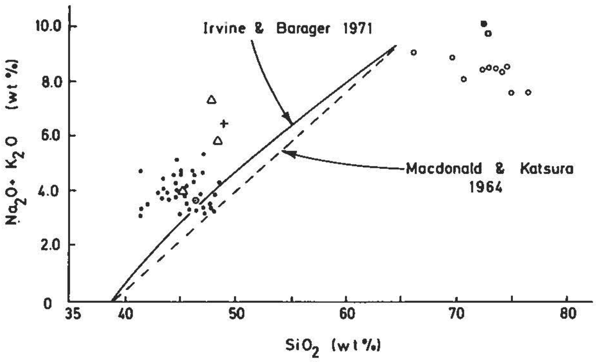

Fig. 3. Alkali–silica diagram for 50 igneous samples from the northern margin of Greenland. Full dots: dolerite dykes and sills; triangles: samples from a single composite dyke; open circles: Kap Washington rhyolitic volcanics; cross: Kap Washington basalt lava. Circle with dot represents average of 28 alkalic basalts from Hawaii (Macdonald & Katsura 1964), open square is average of four peralkaline rhyolites (pantellerites) from Ethiopia (Barberi et al. 1975) and the solid square is average of nine peralkaline rhyolites (comendites) from Kangerdlugssuaq, East Greenland (Brooks & Rucklidge 1976). Full and dashed lines divide the alkaline and tholeiitic basalts of Macdonald & Katsura (1964) and alkaline and subalkaline fields of Irvine & Barager (1971).

The dyke swarms were emplaced in continental crust, and show trace element distributions believed to be characteristic of 'within-plate' magmatism (Fig. 5). We infer a continental rift environment for the dyke swarms and return to the geotectonic implications of this in a later section.

The age of the main swarm is defined on geological evidence. Steeply inclined N–S-trending sheets and dykes cut Carboniferous–Permian and early Cretaceous sediments on western Lockwood Ø and Kap Kane but fail to penetrate the Kap Washington Group volcanic rocks (Brown & Parsons 1981) which are dated at about the end of the Cretaceous (Larsen et al. 1978, Batten et al. 1981). The main dyke swarm is almost certainly of late Cretaceous age and there is nothing to suggest that the compositionally identical WNW–ESE and E–W sets are not of similar age. A K–Ar whole-rock age of 66 Ma has been obtained on an E–W dyke (Dawes & Soper 1971).

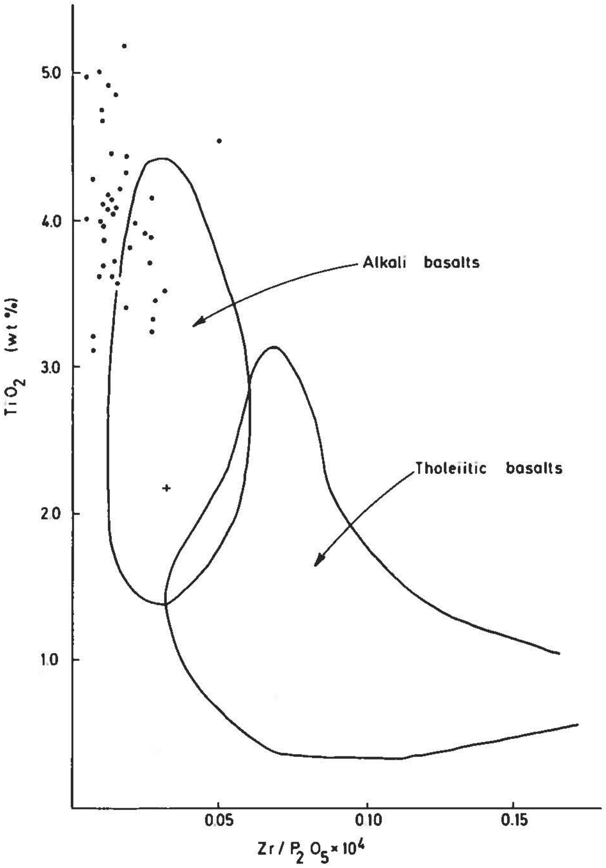

Fig. 4. A TiO_2–Zr/P_2O_5 diagram showing plots of 40 dolerite dykes and a single sample of Kap Washington basalt (cross). Fields of alkali and tholeiitic basalts after Floyd & Winchester (1975).

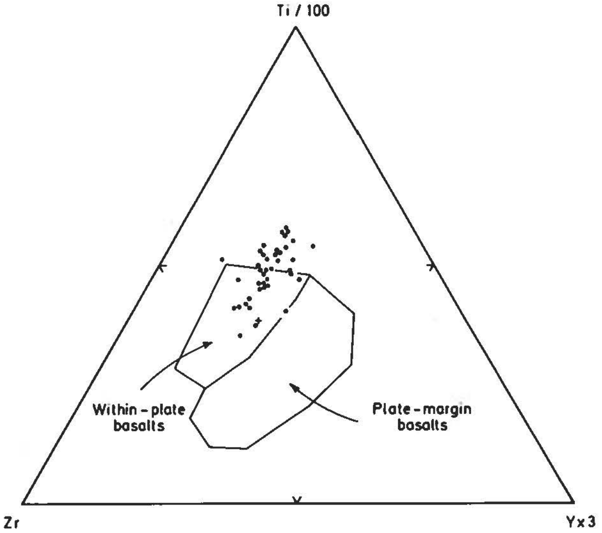

Fig. 5. A discrimination diagram using Ti, Zr and Y showing plots of 40 dolerite dykes and a single sample of Kap Washington basalt (cross). Fields of 'within-plate' and 'plate-margin' basalts after Pearce & Cann (1973).

Table 1. Chemical analyses of dolerite dykes.

	1	2	3	4	5	6	7	8	9	10	11
GGU sample No.	11351	11398	53435	53436	53473	53482	53483	53545	100606	100610	100620
SiO_2	41.98	47.71	44.45	44.49	46.17	43.01	47.18	44.11	46.38	44.50	46.08
TiO_2	4.98	3.72	4.43	5.27	3.81	4.00	3.68	4.92	4.07	3.63	3.89
Al_2O_3	11.64	12.93	13.18	12.44	14.80	13.84	13.43	12.68	14.60	14.66	14.31
Fe_2O_3	5.23	3.34	2.63	3.79	1.78	5.64	4.85	5.45	2.72	3.91	2.30
FeO	10.29	10.47	11.82	11.33	11.21	8.57	8.24	9.56	10.45	9.49	12.18
MnO	0.30	0.22	0.24	0.25	0.23	0.22	0.27	0.24	0.21	0.22	0.21
MgO	5.87	6.25	5.77	5.73	5.51	7.59	4.61	5.94	5.38	6.22	5.46
CaO	10.60	9.71	9.68	9.67	9.67	8.79	8.17	9.90	9.20	9.71	9.36
Na_2O	2.67	2.53	3.57	2.80	3.42	2.56	3.84	3.24	3.35	3.51	3.21
K_2O	0.92	0.88	1.04	1.00	0.96	1.39	1.50	0.69	1.30	0.80	1.17
P_2O_5	2.35	0.56	1.31	1.16	1.22	1.29	1.73	1.12	1.28	0.91	0.81
H_2O+	1.71	1.08	1.66	1.58	0.99	2.82	1.79	1.09	1.54	2.28	0.98
Total	98.54	99.40	99.78	99.51	99.77	99.73	99.29	99.94	99.77	99.84	99.96
CIPW weight norm											
or	5.44	5.20	6.26	5.91	5.73	8.45	8.86	4.08	7.80	4.85	6.97
ab	22.59	21.41	25.12	23.69	29.28	22.26	32.49	26.29	28.47	23.98	26.86
an	17.06	21.33	17.14	18.42	22.37	22.83	14.98	18.02	21.28	22.39	21.40
ne	–	–	3.03	–	–	–	–	0.61	0.16	3.46	0.30
di	16.72	19.15	19.04	18.18	14.94	10.98	11.86	19.65	13.74	17.20	16.73
hy	4.70	18.59	–	6.97	0.10	0.33	5.81	–	–	–	–
ol	12.22	1.55	14.70	9.06	14.70	20.91	9.76	14.14	14.83	15.98	15.54
mt	2.87	2.58	2.77	2.82	2.48	2.69	2.42	2.77	2.51	2.55	2.76
il	9.46	7.07	8.57	10.01	7.31	7.81	6.99	9.34	7.86	7.05	7.46
ap	5.44	1.30	3.08	2.69	2.85	3.08	4.01	2.59	3.01	2.15	1.90
py	–	–	0.80	–	0.80	0.90	–	–	0.80	0.50	0.10
Trace elements (ppm)											
Sr	589	471	758	699	662	574	563	678	741	695	516
Rb	26	20	14	14	16	23	24	11	21	19	25
Zn	n.d.	n.d.	149	n.d.	119	89	n.d.	n.d.	140	152	102
Cu	20	50	20	11	46	31	11	42	24	47	78
Ni	8	51	16	17	37	26	4	41	24	63	56
Ba	626	441	1071	708	1035	1961	1260	669	980	971	639
Nb	n.d.	n.d.	32	n.d.	23	29	n.d.	n.d.	28	26	29
Zr	108	148	230	209	233	113	173	130	156	120	215
Y	47	33	42	42	34	31	24	36	32	31	37

Samples 1 and 2 collected by K. Ellitsgaard-Rasmussen in 1950, 3 to 11 by P. R. Dawes and N. J. Soper in 1969.
Samples 1, 2, 4, 7 and 8: major elements by Ib Sørensen, GGU, 1978; trace elements by J. C. Bailey, University of Copenhagen, 1980. Samples 3, 5, 6, 9, 10 and 11: all major and trace elements (except H_2O^+) by J. G. Holland, University of Durham, U. K. 1972; H_2O^+ by Ib Sørensen, GGU, 1978. n.d. = not determined. Norms are calculated with Fe_2O_3 adjusted to 0.15.

The Kap Washington Volcanic Group

Limited observations made in 1969 (Dawes & Soper 1970, 1973) revealed the existence on the northern coast of Peary Land, within the North Greenland fold belt, of a varied suite of post-Palaeozoic volcanic rocks, for which the name Kap Washington Group was adopted. It was established that the volcanic rocks are tilted southwards, sheared and in places cleaved and in thrust contact with Palaeozoic metasediments of the fold belt (Fig. 6). Observations in 1980 showed that their outcrop is confined to the northern parts of the peninsulas between Kap Cannon and Lockwood Ø (Fig. 2), a more limited area than had been inferred from aerial photograph interpretation (Dawes 1976).

Field identifications of rhyolitic lavas, welded ash-flows, andesites and basic flows, water-lain tuffs, accretionary lapilli tuffs, agglomerates and breccias, with some intrusives, were made in 1969 and an explosive calc-alkaline character was inferred for the suite. Sampling was unfortunately confined largely to the felsic flows. These are typically composed of a microcrystalline matrix, with phenocrysts of perthitic K–Na feldspar and partly resorbed quartz. Flow banding characterises some rhyolitic types, although generally

Fig. 6. The Kap Washington volcanics at Kap Cannon (KC) viewed from the south-west. The volcanics are structurally overlain by metamorphosed Lower Palaeozoic and ?older strata of the North Greenland fold belt. The Kap Cannon thrust is marked by a dashed line. Photograph taken May 1969.

textures show considerable evidence of shearing and alteration. The single basic flow sampled is an olivine basalt with drusy areas which contain alkali feldspar. Detailed mapping of the volcanic rocks has now been undertaken by Brown & Parsons (1981) who report that the group has a total thickness in excess of 5 km and that considerable stratigraphic variation occurs from area to area.

Chemical analyses so far available are all, except one, of felsic flows (Table 3). These are difficult to characterise chemically, but they fall in the subalkaline field of an alkali–silica plot (Fig. 3); the single analysed sample from a basic flow is thoroughly alkaline and this is confirmed by its immobile element distribution (Fig. 4). The relatively high values of Nb, Zr and Y conform to normal values for alkaline rocks, while two samples show normative acmite; a clear indication of peralkaline character. These rocks correspond to comendites. An average composition of ten acid volcanics is given in Table 2.

Table 2. Average major element compositions of dolerite intrusions and Kap Washington Group volcanics.

	E–W dykes	Northerly dykes	Sills	Kap Washington volcanics	
No. of analyses	16	16	4	1	10
SiO_2	45.88	45.08	45.75	49.04	72.64
TiO_2	3.95	4.34	4.01	2.18	0.35
Al_2O_3	13.78	13.42	12.31	14.17	12.99
Fe_2O_3	4.95	4.25	3.94	6.23	3.01
FeO	8.83	9.88	10.99	4.59	1.10
MnO	0.21	0.23	0.23	0.20	0.11
MgO	5.67	5.54	5.60	7.09	0.14
CaO	9.18	9.40	8.44	7.21	0.38
Na_2O	3.14	3.03	2.85	3.58	4.18
K_2O	0.98	1.07	1.41	2.86	4.17
P_2O_5	0.95	1.20	1.12	0.42	0.04
H_2O+	2.18	1.89	3.09	2.43	0.89
Total	99.70	99.33	99.74	100.00	100.00

Details of the 11 analyses of Kap Washington Group volcanics are given in Table 3. E–W dyke average composed of samples GGU 11398, 53482*, 53483, 53487, 53493, 53534*, 100626, 100627*, 218978, 218994, 254647, 254737, 255781, 270831, 270832 and 270833. Northerly dyke average composed of samples GGU 11323, 11345, 11351, 11363, 53425, 53435*, 53436, 53473*, 53555, 100606*, 100610*, 100620*, 100622, 218991, 255697 and 255714. Sill average composed of samples 53454, 53455, 53459 and 255702. Analyses of the samples marked by an asterisk by J. G. Holland, University of Durham, U.K. 1972; the rest by Ib Sørensen, GGU, 1978 and 1980.

Table 3. Chemical analyses of Kap Washington Group volcanics.

	1	2	3	4	5	6	7	8	9	10	11
GGU sample No.	53443	53444	53446	53447	53448	53450	53451	53452	53453	53466	53449
SiO_2	69.80	72.99	66.28	70.82	76.65	74.12	74.58	73.49	75.06	72.59	49.04
TiO_2	0.49	0.19	0.27	0.45	0.59	0.27	0.31	0.37	0.35	0.21	2.18
Al_2O_3	13.14	14.25	18.21	14.39	13.63	10.02	11.73	10.58	10.84	13.08	14.17
Fe_2O_3	4.05	1.51	1.81	3.21	0.36	6.25	3.08	4.55	4.48	0.82	6.23
FeO	0.70	1.49	2.16	1.18	0.59	0.40	1.30	1.50	0.80	0.88	4.59
MnO	0.37	0.05	0.05	0.14	0.01	0.13	0.06	0.16	0.07	0.06	0.20
MgO	0.12	0.19	0.14	0.18	0.08	0.07	0.06	0.11	0.20	0.29	7.09
CaO	1.29	0.02	0.02	0.02	0.02	0.03	0.02	0.42	0.16	1.83	7.21
Na_2O	3.56	5.23	2.50	3.82	7.17	3.51	4.03	3.71	4.00	4.27	3.58
K_2O	5.33	3.24	6.59	4.26	0.36	4.91	4.46	4.77	3.59	4.16	2.86
P_2O_5	0.11	0.02	0.03	0.02	0.03	0.03	0.02	0.03	0.04	0.05	0.42
H_2O+	1.04	0.82	1.94	1.51	0.51	0.26	0.35	0.31	0.41	1.76	2.43
CIPW weight norm											
Q	23.69	28.28	24.34	29.88	33.40	32.77	31.42	30.39	34.55	28.08	–
C	–	2.16	7.10	3.54	1.45	–	0.26	–	0.19	–	–
or	31.85	19.32	39.71	25.59	2.13	29.08	26.48	28.31	21.28	25.00	17.32
ab	30.46	44.60	21.58	32.83	61.01	24.30	34.27	27.91	34.02	36.81	24.32
an	4.16	–	–	–	–	–	–	–	0.53	4.30	14.49
ne	–	–	–	–	–	–	–	–	–	–	3.65
ac	–	–	–	–	–	2.31	–	2.15	–	–	–
ns	–	–	–	–	–	0.67	–	0.26	–	–	–
di	1.39	–	–	–	–	–	–	1.68	–	3.95	15.83
hy	6.05	4.58	5.81	6.18	0.64	9.74	5.90	8.15	7.36	0.96	–
ol	–	–	–	–	–	–	–	–	–	–	16.64
mt	0.84	0.55	0.74	0.79	0.18	–	0.78	–	0.93	0.32	2.00
il	0.95	0.36	0.53	0.87	1.12	0.51	0.59	0.70	0.66	0.40	4.25
ap	0.25	0.12	0.12	0.12	0.12	0.18	0.12	0.07	0.09	0.12	1.00
Trace elements (ppm)											
Sr	17	35	15	47	74	8	7	24	23	202	323
Rb	146	144	422	135	13	288	206	159	141	159	91
Zn	204	251	451	214	46	347	252	286	294	29	121
Cu	5	4	6	6	3	8	6	8	8	4	150
Ni	1	4	3	2	4	2	4	4	4	3	96
Ba	224	239	195	62	55	20	40	47	147	874	446
Nb	82	106	137	78	71	299	151	161	170	46	29
Zr	604	804	1092	530	444	1837	908	1023	1070	242	133
Y	55	69	105	71	37	145	86	98	103	20	28

Samples collected by P. R. Dawes and N. J. Soper in 1969.
All major and trace elements (except H_2O^+) by J. G. Holland, University of Durham, 1972. H_2O^+ by Ib Sørensen, GGU, 1978.
Major analyses recalculated to 100%. Norms are calculated with Fe_2O_3 adjusted to 0.15.

Preliminary petrological studies by Brown & Parsons (1981) confirm the thoroughly alkalic nature of the province. They report riebeckite trachytes, riebeckite comendites or pantellerites and alkalic basalts as the chief lava types; the presence of the soda amphibole riebeckite in both siliceous lavas and ash-flow tuffs demonstrates that the Kap Washington Group is of peralkaline character.

Peralkaline rocks are well known world-wide from regions associated with crustal attenuation and rift tectonics (e.g. Africa – Red Sea region, Gass 1970, Barberi et al. 1972) and are known from the North Atlantic (e.g. East Greenland, Brooks & Rucklidge 1976). We infer an extensional rift environment for the volcanicity, linking the origin of the province to that of the slightly earlier swarms of alkalic basic dykes.

The Kap Washington Group volcanic rocks overlie a sequence of shales and sandstones which have yielded angiosperm leaf fragments that are not older than mid-Cretaceous in age (B. E. Koch, pers. comm. 1981). Shales interbedded with the volcanic rocks about 3 km above the base have yielded pollen and spores of end-Cretaceous (Campanian or Maastrichtian) age (D. J. Batten, pers. comm. 1981). Thus at least the basal 3 km of the volcanic pile is of late Cretaceous age and we adopt an age range of 75–65 Ma for the volcanicity — at or just prior to the Cretaceous–Tertiary boundary (Batten et al. 1981). The upper age limit is within the experimental error of the earlier Rb–Sr whole-rock age on five rhyolite samples of 63 Ma (Larsen et al. 1978).

Kap Cannon thrust zone

The outcrop of the Kap Washington volcanic rocks is terminated to the south at a major south-dipping thrust zone along which the highly-deformed metasediments of the North Greenland fold belt are displaced relatively northwards over the volcanic rocks (Dawes & Soper 1970, Fig. 6). Mylonitisation occurs along the thrust plane, which has been traced from Kap Cannon as far west as Lockwood Ø. Widespread retrograde metamorphism affects the metasediments along the north coast, especially near the thrust, and K–Ar mineral ages of 84–42 Ma from Palaeozoic mica schists (Dawes & Soper 1971) support the idea of a weak Tertiary overprinting connected with the thrusting.

Mapping in 1980 revealed at least two further steep thrusts or reversed faults to the north of the main Kap Cannon thrust; the thrust slices contain Carboniferous–Permian sediments and Palaeozoic flysch in addition to the volcanic rocks. The volcanic rocks record only the effects of Tertiary deformation; there is little folding but locally strong cleavage development, brecciation and shearing. Displacement on the thrusts is not known. Inclination of the main thrust plane ranges from 40° at Kap Cannon in the east to near vertical on the west side of Lockwood Ø. The structures may be of upthrust type rather than far-travelled nappes.

The thrusting episode post-dates the Kap Washington Group. K–Ar whole-rock ages of 32 and 35 Ma (early Oligocene) from somewhat cleaved rhyolite samples (Dawes & Soper 1971) are perhaps best interpreted as indicating erosion and cooling during post-thrusting uplift. A late Paleocene – Eocene age for the compression event is adopted.

Wandel Sea Basin structure

South-east-trending faults and open folds affect the Wandel Sea Basin cover sequences of eastern Peary Land and Kronprins Christian Land (Haller 1970, Dawes 1976, Håkansson 1979, Håkansson et al. 1981). These structures are parallel to the Wandel Sea continental margin and Spitzbergen fracture zone. Faults of similar trend are present in adjacent parts of the Palaeozoic platform and several major examples have been mapped recently in the fold belt north of the Harder Fjord fault zone (Fig. 2). Håkansson (1979) has shown that the main block faulting affected eastern Peary Land after deposition of the youngest recorded cover sediments (?Paleocene) and that the sense of displacement across the faults implied by the geometry of drag folds is opposite to that determined from matching the stratigraphy. He suggests that faults of pre-Tertiary (perhaps Jurassic) age were reactivated in post-Paleocene time by tension followed by compression, or vice versa. As discussed below the latter is more consistent with displacement models.

Harder Fjord fault zone

This major E–W-trending fault zone has been traced for over 250 km across the North Greenland fold belt from the Wandel Sea coast to Nansen Land (Fig. 2). Its eastern portion takes the form of a narrow graben in which representatives of the post-Franklinian cover sequence are preserved. These comprise non-marine upper Permian, marine Cretaceous, and non-marine early Tertiary (Paleocene) sediments (Croxton et al. 1980, Wagner et al. in press), as well as greenstones, subaqueous extrusives in part, of unknown age. Doleritic dykes are associated with the central part of the fault zone and, also in this area, a few kilometres to the south of the fault several breccia pipes and necks occur, which contain basement gneiss xenoliths and volcanic lithologies comparable to some of those in the Kap Washington Group (Soper et al. 1980, Parsons 1981).

The origin of the fault zone is uncertain. Surlyk et al. (1980) postulate that the fault had a considerable influence on the early development of the Franklinian trough sedimentation, although the direct field evidence of this has been questioned (Higgins et al. 1981). The fault zone probably influenced the pattern of emplacement of the late Cretaceous basic dyke swarms, the breccia pipes and necks and the greenstones could be directly linked to the existence of the fault, and there were substantial movements subsequent to the deposition of the early Tertiary sediments. Some basic dykes aligned in the fault zone in the Frederick E. Hyde Fjord area show evidence of post-dyke crushing (Dawes & Soper 1979). Recent solfatara activity is recorded from its eastern end and earthquake activity along the fault zone indicates continuing movement (Dawes & Peel 1981).

Dawes (1973) noted the parallelism of this and other large faults in Peary Land to the Spitzbergen fracture zone, although recent work has not revealed clear evidence of wrench displacement. For example, 25 km south of Lockwood Ø the plane of the northern branch of the Harder Fjord fault zone is well exposed, inclined 60° to the south. Downthrow is to the south and equivalent to the whole thickness of the Polkorridoren Group — as much as 2 km (Higgins et al. 1981). It is thought to be associated with a period of stress relaxation after the Kap Cannon thrusting episode and probably marks the main uplift of the north Peary Land mountain chain in mid- to late Tertiary time.

Spreading history of the ocean basins around Greenland

The chronology of sea-floor spreading in the oceans around Greenland is now established in outline, although the origin of Baffin Bay and the whole Amerasia Basin remains unclear. There are problems concerning the timing of the onset of spreading in many areas and the degree to which thinning and extension of the continental crust were involved. We briefly review the spreading history of the North-East Atlantic, Eurasia Basin and Labrador Sea – Baffin Bay.

North-East Atlantic

In Mesozoic time zones of crustal extension were already developing on the sites of Greenland's future rifted margins. A shallow seaway was initiated between Greenland and Scandinavia in late Palaeozoic time and a marine connection was maintained through it between the Boreal Ocean and western Europe for much of the Jurassic and Cretaceous, associated on the Greenland side with coast-parallel extensional faulting (Birkelund & Perch-Nielsen 1976, Surlyk 1977).

Anomaly 24 (52 Ma) is the earliest recognised with certainty in the Norwegian Sea (Vogt et al. 1970) and between South-East Greenland and Rockall (Vogt & Avery 1974) and the general view has been that spreading commenced between Greenland and Europe shortly before anomaly 24 time (early Eocene), preceded by an abortive spreading phase in the Cretaceous which opened Rockall Bight. However, anomaly 24 lies oblique to the shelf edge in some areas and as much as 100 km to seaward of it. Phillips et al. (1978) have interpreted the intervening areas as oceanic crust generated by pre-anomaly 24 spreading. These areas could equally represent thinned and subsided continental crust as found, for example, off South-East Greenland by Featherstone et al. (1977). Two lines of evidence confirm the initiation of spreading at anomaly 24 time. The eruption of very large volumes of tholeiitic basalt onto the East Greenland continental margin can realistically only be associated with major rifting at the onset of spreading. The main basalt pile in this area has been dated palynologically as late Paleocene – early Eocene (55–53 Ma, Soper et al. 1976) and this is consistent with a fission-track age of 54.6±1.7 Ma on the Skaergaard intrusion (Brooks & Gleadow 1977). The reversed magnetic polarity of the basalt sequence confirms its eruption in the 24–25 reversed epoch (Faller 1975). Secondly, a pronounced shift in spreading geometry occurred in the Labrador Sea at anomaly 24–25 time and this must reflect a major change in plate configuration. The evidence thus confirms that the initiation of the main phase of spreading between Greenland and Europe took place at about the Paleocene–Eocene boundary (53 Ma) (Talwani & Eldholm 1977).

Eurasia Basin and Spitzbergen fracture zone

The ocean floor of the Eurasia Basin was generated by spreading on the Nansen axis which connects to the North Atlantic via the Spitzbergen fracture zone and the Knipovich Ridge (De Geer line). The outermost ridge flank anomaly recognised in the Eurasia Basin is 24 (52 Ma), but there is room for anomalies 25 to 27 between it and the continental slope (Vogt et al. 1979). The intervening zone may be floored by oceanic crust as old as 62 Ma or by thinned and subsided continental crust. Limited geological evidence bearing on this early extensional phase is discussed subsequently. Notable features in the present context are the Morris Jesup Rise and Yermak Plateau, northern extensions of the North Greenland and Svalbard continental shelves. These are interpreted by Feden et al. (1979) as aseismic ridges generated by a Yermak hot spot in the period between anomalies 18 and 13 (40–35 Ma) and subsequently rafted apart by post-anomaly 13 spreading. Johnson et al. (in press) interpret that part of the Yermak plateau south of 82°N as continental crust. These authors, and Phillips et al. (1978), Feden et al. (1979) and Kovacs (this volume), all associate the Yermak hot spot with a triple rift junction north of Greenland, active until anomaly 13 time, whose third arm was represented by the Nares Strait lineament. These models require substantially greater displacement on the Nares Strait line than is permitted by the geological constraints (Christie et al. 1981, Dawes & Kerr, this volume).

Geometrical problems are also associated with the De Geer line along which the Barents Shelf has been displaced dextrally with respect to the Wandel Sea margin of North-East Greenland. Early reconstructions which juxtaposed the coastlines or continental margins (Bullard et al. 1965, Bott & Watts 1971, Pitman & Talwani 1972) involved unacceptably large overlaps in this region. Reconstructions by means of magnetic anomaly matching (Talwani & Eldholm 1977, Phillips et al. 1978) produced small overlaps between the Wandel Sea and Barents shelves until about anomaly 13 time, implying oblique compression until the early Oligocene followed by oblique extension.

Labrador Sea and Baffin Bay

The origin of the Labrador Sea – Baffin Bay seaway is still poorly understood. The nature of the crust is uncertain and so therefore is the position of the continent–ocean boundary; many reconstructions show coastline mismatches too large to be readily accommodated by within-plate deformations.

Linear magnetic anomalies in the southern Labrador Sea are contiguous with ocean floor anomalies 20–24 in the North Atlantic with which they define an ancient triple junction (Vogt et al. 1969, Laughton 1971, Vogt & Avery 1974). It is widely accepted that the Labrador Sea was opened by a phase of sea-floor spreading which terminated after anomaly 13 time — early Oligocene (Kristoffersen & Talwani 1977). However, the extent of this spreading and its time of onset are in doubt. A number of authors, most recently Srivastava (1978), recognise anomalies as old as 32 (75 Ma) and Srivastava suggests a northward propagation of the spreading axis, with anomalies 31 and 28 the oldest present in the central and northern sectors of the Labrador Sea. Conversely van der Linden (1975) showed, mainly on the basis of seismic refraction data, that the north-western margin of the Labrador Sea could be floored by thinned continental crust and Hinz et al. (1979) confirmed that the continent–ocean boundary cannot be defined in this region by means of conventional geophysical characteristics. On the basis of borehole and seismic reflection data, Grant (1980, this volume) established that an early Cretaceous unconformity, presumably overlying continental basement, occurs in the same area as anomaly 28 in Srivastava's model, suggesting that the outer magnetic anomalies in the Labrador Sea have been incorrectly identified. Spreading may well have commenced significantly later than 75 Ma, but preceded in Cretaceous time by extensional thinning and subsidence of the continental crust.

The origin of Baffin Bay is even more problematical. Two schools of thought have emerged. One regards the thin (c. 5 km) crust of the deeper parts of the bay as oceanic, generated by spreading from a northward extension of the Labrador Sea axis (Keen & Barrett 1972, Keen et al. 1974, Kristoffersen & Talwani 1977, Srivastava 1978). This is supported by Jackson et al. (1979), who report linear magnetic anomalies in central Baffin Bay which they correlate with ocean floor anomalies 13–21 and a central linear gravity low which they associate with the extinct spreading axis. The other school interprets the bay as a thinned and subsided continental region (Kerr 1967, 1980, van der Linden 1975), thus avoiding some of the geometrical problems associated with a 'pure spreading' hypothesis.

Two lines of geological evidence must be considered in relation to the problem. One concerns the small sinistral wrench displacement — as little as 25 km — which appears permissible on the Nares Strait line (Christie et al. 1981). This effectively rules out a 'pure spreading' origin for Baffin Bay. It is impossible to accommodate the required displacement of Greenland plus the attached part of Ellesmere Island with respect to the remainder of the Canadian Arctic Archipelago without invoking within-plate deformations greatly in excess of those associated with the Eurekan structures. Much of Baffin Bay must be floored by continental crust. Smaller displacements can, however, be accommodated and are required by the conventional ideas of opening of the Labrador Sea.

The second line of evidence concerns the similarity between the basalt provinces of East and West Greenland, not only as regards the composition and volume of the picritic-tholeiitic extrusives, but also the pre-eruptive sedimentary history (Higgins & Soper 1981). By analogy with the geotectonic setting of the East Greenland province, we support the idea that the West Greenland basalts mark the major rifting episode in the Davis Strait (Clarke & Upton 1971), an idea adopted by Srivastava (1978) in his model. The Nûgssuaq basalts have been precisely dated palynologically as Danian (Jürgensen & Mikkelsen 1974), a date supported by Ar 40–39 ages in the range 62–57 Ma (Parrott & Reynolds 1975). The dominantly reversed polarity of the West Greenland basalts (Athavale & Sharma 1975) thus indicates eruption in the 3 Ma reversed epoch between anomalies 26 and 27 (61–58 Ma). This is the most probable date at which rifting propagated through the Davis Strait into Baffin Bay — somewhat earlier than envisaged by Kristoffersen & Talwani (1977) or Srivastava (1978). It should be noted that a coast-parallel basic dyke swarm of supposed Jurassic age in South-West Greenland has been associated by Watt (1969) with earlier tensional rifting in the region.

A complex history for the Labrador Sea and Baffin Bay must therefore be envisaged, with the attenuation and 'oceanisation' of continental crust as particularly important, plus some phases of spreading. Assuming spreading occurred in Baffin Bay it must have been extremely slow (close to the rotation pole) until anomaly 21 time, with a change in geometry at anomaly 24 (Srivastava 1978) and a final phase of slow, oblique spreading in the period of anomalies 21–13. The displacement vector for this last phase adopted by Srivastava (1978) is parallel to Nares Strait and the wrench motion required there to satisfy the Jackson et al. (1979) model for central Baffin Bay is in excess of 100 km, which can be ruled out. The vector derived by Kristoffersen & Talwani (1977) is about 35° oblique to Nares Strait and can in principle be resolved into sinistral slip parallel to the Strait and compression normal to the northern Ellesmere – North Greenland sector of the Eurekan fold belt. However, the slip component is still geologically unacceptably large and the displacement vector inclined at such an acute angle to the assumed Baffin Bay spreading axis as to render the model geometrically improbable. The geological constraints on the total wrench offset on Nares Strait thus raise serious problems for any hypothesis for the origin of Baffin Bay which involves significant sea-floor spreading and it questions the interpretations of Jackson et al. (1979) who identify sea-floor magnetic anomalies and a central spreading axis in the Bay.

Makarov Basin

Almost every conceivable origin has at some time been proposed for the Alpha Ridge (continental crust, spreading centre, subduction zone, incipient island arc) and for the Makarov Basin which lies between that and the Lomonosov Ridge (see Sweeney 1978). Recent interpretations of aeromagnetic data obtained in 1977–78

(Taylor et al. 1981, Kovacs, this volume) have tentatively identified sea-floor spreading type anomalies in the Makarov Basin and suggest that spreading began at anomaly 34 time (late Cretaceous) and continued to anomaly 19 (42 Ma, late Eocene). The implications of a model which involves simultaneous spreading on the Makarov and Nansen axes in the Eocene, but which also accepts strictly limited displacement on the Nares Strait line have yet to be explored.

Discussion

The extent to which the Cretaceous–Palaeogene history of North Greenland can be related to the changing geotectonic pattern outlined above, is discussed in the following.

Late Cretaceous events (pre-anomaly 29, 65 Ma)

A reconstruction at 65 Ma, (the Cretaceous–Tertiary boundary) shows North Greenland, the Barents Shelf and Lomonosov continental fragment in juxtaposition, prior to the onset of spreading on the Nansen axis but, if the proposals of Taylor et al. (1981) and Kovacs (this volume) are correct, with spreading already taking place in the Makarov Basin (Fig. 7A). The North Greenland dyke swarms were emplaced into this continental region. Their alkalic composition, 'within-plate' chemical character and spatial relationship to the later explosive alkalic volcanicity of the Kap Washington Group all point to their association with a continental rift system. The main concentration of coast-normal dykes aligns rather better with the Makarov than the Nansen axis. If Johnson et al. (in press) are correct in interpreting part of the Yermak Plateau as continental, and if the Nansen spreading axis did not propagate across it until about anomaly 13 time, then the dykes are unlikely to represent the southern termination of the Nansen axis. The only credible alternative, on present evidence, would then be that they represent the termination of the Makarov axis. However, the model of Johnson et al. requires that between the times of anomalies 24 and 13 the spreading axes of the Eurasia and North-East Atlantic basins were linked by a transform fault system through Svalbard. The improbability of this, together with the existence of pre-anomaly 24 crust of unknown origin in the Eurasia Basin leads us to favour a link between the coast-normal dyke swarm and early extensional rifting along the Nansen axis as implied in Fig. 7.

We note the contrast in composition between the North Greenland dykes and the tholeiitic basalts exposed on the Barents margin of the Eurasia Basin in Svalbard and Franz Josef Land (Tyrell & Sandford 1933, Gusev 1979), which are of comparable but poorly defined age (Dibner 1957, Gayer et al. 1966, Burov et al. 1976), and the fact that the alkaline olivine plateau basalt lavas of the Woodfjorden area in northern Spitzbergen, are now dated as late Tertiary (Prestvik 1978). On the grounds of both their age and alkalic composition, we do not relate either the basic dykes or the Kap Washington volcanics directly to the Yermak hot spot, as suggested by Feden et al. (1979). The geographical coincidence of the alkalic rift magmatism and later hot spot activity may well have petrogenetic significance, but exploration of this problem must await the completion of petrological studies of the North Greenland rocks and also sampling of the hot spot products from the submarine plateaux. Furthermore, in view of the late Cretaceous age assignment of the Kap Washington volcanics, it hardly seems likely that the Paleocene tuffs of Svalbard (Dypvik & Nagy 1978) were derived from the Kap Washington volcanism as has been suggested (Feden et al. 1979).

We regard the alkaline–peralkaline volcanicity of the Kap Washington Group as a further manifestation of the crustal tension and rift magmatism represented by the dykes. The timing of this final phase of alkalic magmatic activity has geotectonic implications of regional importance. The dyke swarms and rift volcanics show that the late Cretaceous tectonic regime in the region was extensional — as indeed it was throughout the whole crustal segment from the Sverdrup Basin through North Greenland to the northern Barents Shelf. The volcanic activity is therefore likely to pre-date the onset of spreading in the Labrador Sea, since the anti-clockwise rotation of Greenland with respect to North America while still pinned to the Eurasia plate clearly induced compression in the region, as evidenced by the later episode of Kap Cannon thrusting. The age of the Kap Washington Group may thus be regarded as a maximum for the onset of true spreading in the Labrador Sea, and places this event no earlier than end-Cretaceous.

This represents a significant revision of the time at which major anti-clockwise displacement of Greenland commenced. Supporting evidence comes from the Canadian Arctic Archipelago where a fundamental but hitherto unexplained change in tectonic regime also took place at about the end of the Cretaceous. The Mesozoic history of continuous subsidence of the Sverdrup Basin, with a sediment fill derived from extra-basinal sources, changed in latest Cretaceous time to one of differential vertical movements with uplifted blocks supplying mainly locally derived non-marine sediments of the Eureka Sound Formation (Trettin et al. 1972, Balkwill 1978). Basic intrusive and effusive activity, centred on Axel Heiberg Island, ceased at the same time. These changes are consistent with our proposed timing for the onset of spreading in the Labrador Sea, and also with Srivastava's location of the rotation pole for the early opening phase (prior to anomaly 25, 56 Ma) immediately south of the Sverdrup Basin.

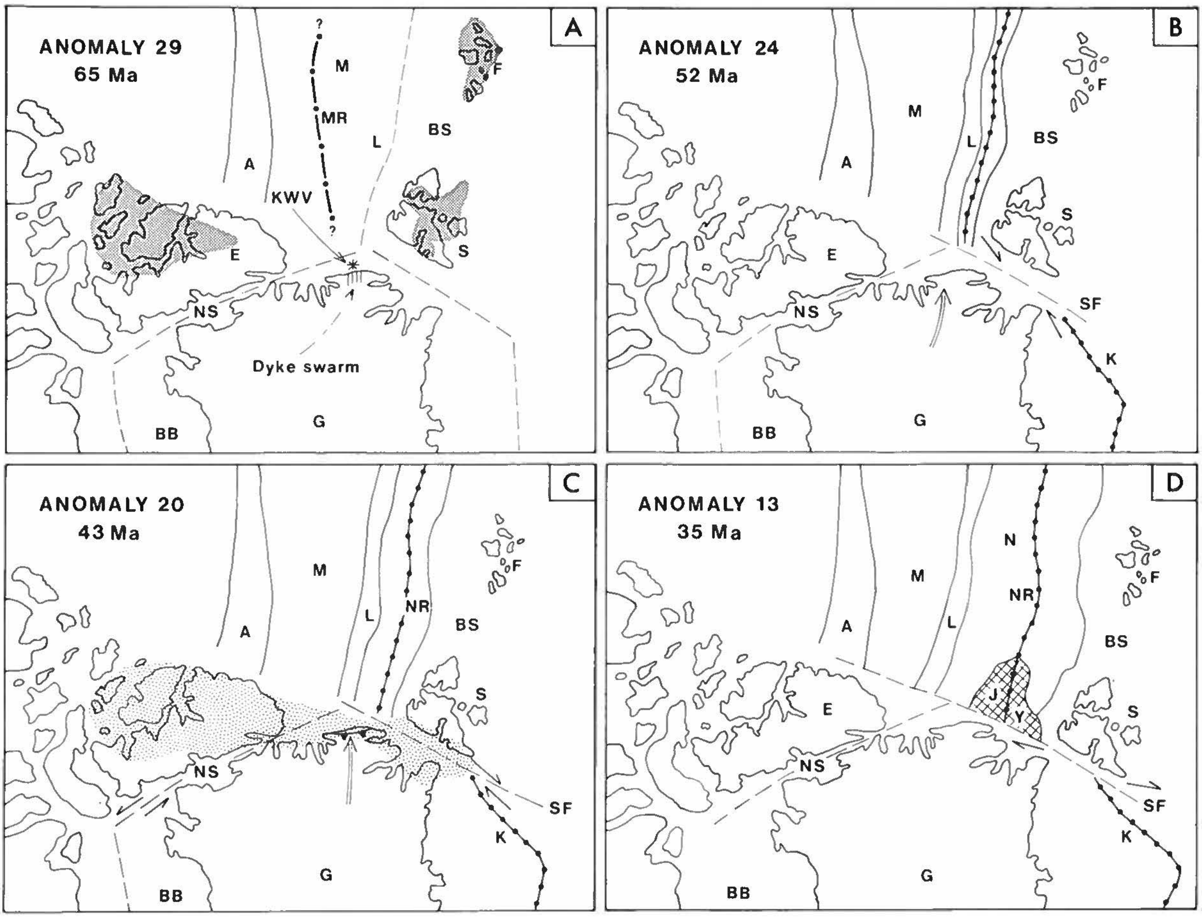

Fig. 7. Cartoon reconstructions of northern Greenland, eastern Canadian Arctic and the Barents Shelf. Fracture zones are shown dashed, active spreading axes are dotted, dot-dash line represents a possible extinct spreading axis, and plate movement direction is shown by single arrows. The fit of northern Greenland and the Barents Shelf approximately follows Phillips et al. (1978); fit of Greenland and Ellesmere Island approximately after Le Pichon et al. (1977).

G – Greenland, E – Ellesmere Island, S – Svalbard, F – Franz Josef Land, BS – Barents Shelf, A – Alpha Ridge, MR – Makarov Ridge, L – Lomonosov Ridge, NR – Nansen (Gakkel) Ridge, K – Knipovich Ridge, BB – Baffin Bay, M – Makarov Basin, N – Nansen Basin, SF – Spitzbergen Fracture Zone, NS – Nares Strait fracture.

A. Reconstruction at anomaly 29 time (Cretaceous–Tertiary boundary) showing main Mesozoic magmatic events. KWV – Kap Washington volcanics (75–65 Ma); shaded areas represent mainly basaltic magmatism (Jurassic–Cretaceous).

B. Reconstruction at anomaly 24 time (early Eocene). The double arrow represents the inferred compression direction in North Greenland.

C. Reconstruction at anomaly 20 time (middle Eocene) showing minor inferred strike-slip displacement along Nares Strait, the Kap Cannon thrust and the areas affected by the Eurekan orogeny in Canada and Greenland and the West Spitzbergen – Wandel Sea orogenic zone (shaded).

D. Reconstruction at anomaly 13 time (early Oligocene). Sea-floor spreading ceases between Greenland and North America. Arrows on Spitzbergen fracture zone indicate the inferred extensional displacement. J – Morris Jesup Rise, Y – Yermak Plateau.

Anomaly 29 to anomaly 13 (65 Ma to 35 Ma)

This represents the period of sea-floor spreading in the Labrador Sea. Spreading commenced on the Nansen axis perhaps as early as anomaly 27 time (61 Ma) and in the North-East Atlantic as late as 52 Ma. Thus for a period in the Eocene and early Oligocene (52–35 Ma) Greenland underwent displacements with respect to all adjacent continental masses. A geometrical consequence of this was the eastward migration of the southern end of the Nansen axis past the North Greenland continental margin, presumably along a major transform fault associated with the Spitzbergen fracture zone (Fig. 7B). A second consequence was the compression of North Greenland and Ellesmere Island against the Lomonosov continental fragment, producing the E–W portion of the Eurekan fold belt.

The main associated structure in North Greenland is the Kap Cannon thrust complex whose Eocene age is rather poorly defined geologically but entirely consistent with the spreading chronology adopted here. Trettin & Balkwill (1979) have emphasised the apparent discrepancy in timing between the opening of the Labrador Sea and Baffin Bay (previously thought to be late Cretaceous – Paleocene) and the supposedly associated compressional phase of the Eurekan orogeny in the eastern Canadian Arctic Islands (middle Eocene to early Miocene). Bearing in mind that the early Miocene date refers to the post-Eurekan cover sequence, and thus represents a minimum age for the closure of deformation, and the younger (early Oligocene) age for the cessation of spreading in the Labrador Sea inherent in the revised polarity time scale, our interpretation largely resolves the discrepancy. If, as we propose, spreading did not commence in the Labrador Sea until about the Cretaceous–Tertiary boundary and Greenland remained 'pinned' against the Barents Shelf until Eocene time, Eurekan compression could have been associated with the anti-clockwise rotation of Greenland and Ellesmere Island with respect to the North American plate and have been restricted to Eocene – early Oligocene time. Any sinistral displacement on the Nares Strait lineament is likely to have taken place during this period (Fig. 7C). The proposal that spreading was taking place simultaneously on the Makarov axis does not materially change this scenario, but does prove difficult to reconcile with limited displacement on the Nares Strait line.

The Yermak and Morris Jesup aseismic ridges were generated in the period 41–35 Ma (Fig. 7D) and a speculative connection can be made between compression and perhaps over-riding of the southern flank of the Yermak hot spot by the North Greenland continental margin, and excess basalt discharge from the adjacent section of the Nansen spreading axis. Termination of this process synchronously with the end of spreading in the Labrador Sea must otherwise be seen as unrelated coincidence.

Compressional folding of the Wandel Sea Basin sediments of eastern Peary Land on ESE axes in post-Paleocene time can be associated with oblique compression of North-East Greenland against the Barents Shelf continental margin prior to anomaly 13, as required by most spreading models. These folds are presumably counterparts of the better dated (Eocene – early Oligocene) main phase structures of the West Spitzbergen orogen (Harland & Horsfield 1974). Deformation was more intense in Spitzbergen, with easterly directed thrusting, and has been similarly interpreted as the effect of dextral transpression across the Spitzbergen fracture zone (Harland 1965, 1969, Lowell 1972). One might predict the intensification of the Peary Land deformation offshore towards the Wandel Sea shelf edge, with SW-directed thrusting.

Post-anomaly 13

After spreading had ceased in the Labrador Sea, North-East Greenland ceased to be pinned against the Barents Shelf and oblique extension commenced along the Spitzbergen fracture zone, as evidenced by the overlaps of a few tens of kilometres shown in this region on recent anomaly 13 reconstructions (Talwani & Eldholm 1977, Phillips et al. 1978). Onshore structures are compatible with limited extension, rather than a hundred or more kilometres demanded by earlier models. The principal extensional structure in Svalbard is the Forlandsundet graben, whose sediment fill Harland (1969) considered to post-date the West Spitzbergen orogeny. We associate block faulting of the Wandel Sea succession with this phase; of the two alternative structural interpretations proposed by Håkansson (1979) reactivation of earlier structures by compression followed by extension is the preferred model. Again, more intense extensional faulting might be expected offshore.

In northernmost Greenland, major uplift took place in post-anomaly 13 time, with significant displacement on E–W fractures such as the Harder Fjord fault zone. Feden et al. (1979) have discussed the elevation of this region as a response to the development of the Yermak hot spot. Isostatic response to crustal thickening due to the Kap Cannon compressive phase is also a possibility. Current resurgence of the hot spot suggested by Feden et al. (1979) is perhaps evidenced by the solfataric activity (see Troelsen 1949) along E–W fault lines in the region.

Conclusions and the Nares Strait problem

The presence on the northern coast of Greenland in Peary Land of late Cretaceous intrusive and extrusive rocks cannot be used specifically (as can other geological features described in this symposium) to infer the magnitude of possible offset along Nares Strait. They do, however, provide some restraint on reconstructed relative positions of northern Greenland and Ellesmere Island at the time of their genesis. In many pre-drift reconstructions involving substantial strike-slip motion along Nares Strait, Peary Land is placed adjacent to north-eastern Ellesmere Island, a region devoid of comparable magmatic rocks (e.g. Feden et al. 1979). While an argument based on the absence of rocks can never be conclusive, the geotectonic interpretations of this paper suggest that the lack of magmatic products in this part of Ellesmere Island is not coincidental.

The Greenland alkaline magmatism was associated with continental rifting that pre-dated the opening of the Eurasia Basin. Of particular interest is the conspicuous coast-normal basic dyke swarm that intensifies northwards towards the outer coast and that is regarded as the southernmost effect of the suturing that separated the Lomonosov Ridge from the Barents Shelf. Dykes occur in Greenland as far west as about long. 48° and thus in reconstructions of the type mentioned above, it is surprising that no corresponding magmatic products are known from north-eastern Ellesmere Island. We suggest that the late Cretaceous magmatism of the region, manifested onshore in north coastal Peary Land, occurred in a specific geological location far removed from north-eastern Ellesmere Island.

This paper shows that the magmatic and tectonic history of North Greenland in Cretaceous and Tertiary time relates well to the timing and pattern of plate displacements when viewed in the context of the Hailwood et al. (1979) polarity time scale. Broadly, three tectonic periods can be recognised on Greenland's northern margin: a late Cretaceous phase of extensional tectonism associated with continental rift magmatism, a Paleocene–Eocene phase of compression and northward thrusting, and a later Tertiary period of vertical displacements in a mildly extensional tectonic environment. The Palaeogene compressional phase is seen as a consequence of anti-clockwise rotation of Greenland due to the opening of the Labrador Sea. We have shown, by two independent lines of evidence, that this compressive phase commenced at about the Cretaceous–Tertiary boundary, suggesting that major spreading in the Labrador Sea also commenced then, some 10 Ma later than previously supposed. At that time Greenland was still part of the Eurasia plate and the compressional buildup was not manifested in thrust displacement of the northern margin against the Lomonosov continental fragment until the north-eastern (Wandel Sea) margin began to become 'unpinned' from the Barents Shelf in Eocene time.

Recognition of a coherent relationship between events on the continental margins of Greenland and spreading patterns does not in itself go far towards 'solving' the Nares Strait conflict. But in combination with the constraints on the magnitude of displacement along Nares Strait recognised by Christie et al. (1981) it does provide an essential framework within which the problem should be viewed. The geological evidence shows that Greenland did undergo significant anti-clockwise rotation with respect to the North American plate in early Tertiary time, and this precludes an origin for the Labrador Sea and Baffin Bay in which sea-floor spreading plays no part. The problem is to define the extent of spreading, and how the resulting displacements have been accommodated. Clearly, within-plate strains are important, not only in this context but generally: the failure of rigid-plate theory to produce mutually compatible reconstructions of the North-East Atlantic, Labrador Sea and Eurasia Basin illustrates the point.

Geologists would welcome a magnetic anomaly superposition study of these three oceans directed towards producing a series of finite difference displacements which not only satisfy the constraints provided by the geological evidence, but serve specifically to quantify the within-plate deformations at each stage. Some recent studies have made a start in this direction, but better integration of the onshore geological evidence with the interpretation of offshore geophysical data is required.

Acknowledgements

Two of us, N. J. S. and P. R. D., thank John D. C. Peacock and Bruce K. Reid, leaders of the Joint Services Expedition to Peary Land for support and coordination of the geological work into the 1969 expedition programme. P. R. D. thanks C. K. Brooks and J. C. Bailey for discussions on the chemical composition of the Kap Washington Group volcanics. C. K. Brooks, G. L. Johnson, L. M. Larsen and H. P. Trettin critically read the manuscript and offered helpful suggestions for which we are grateful. The paper is published with the permission of the Director of the Geological Survey of Greenland.

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