Paleomagnetism and cycle stratigraphy of the Triassic Fleming Fjord and Gipsdalen Formations of East Greenland

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A 210 m section of Late Triassic Fleming Fjord Formation (the Malmros Klint Member and the lowermost 80m of the overlying Carlsberg Fjord beds of the Ørsted Dal Member) in the Tait B jerg area of the Jameson Land Basin, East Greenland, was sampled for paleomagnetic study and measured for cycle stratigraphic analysis. Paleomagnetic samples were also taken from the underlying Gipsdalen Formation in the Gipsdalen area. A high stability character-istic magnetization carried by hematite was successfully isolated in 63 sampling levels in the Fleming Fjord Formation and 9 sampling sites in the Gipsdalen Formation using pro-gressive thermal demagnetization. The mean characteristic directions for the Fleming Fjord and the Gipsdalen Formations may be be biased by sedimentary inclination error but are consistent with a northward drift of East Greenland of about 10° from the arid (ca. 25° N) to semihumid (ca. 35° N) paleoclimatic belts in the Middle to Late Triassic. Seven normal and reversed polarity intervals are clearly delineated in the Fleming Fjord Formation section. A preferred correlation of the magnetostratigraphy to a cyclostratigraphically calibrated ref-erence polarity sequence recently derived from drill cores in the Newark Basin of eastern North America suggests that the sampled interval represents about a 3.5 m.y. interval of the late Norian. The Malmros Klint Member and the overlying Carlsberg Fjord beds have com-posite sedimentary cycles that vary in thickness from 25 m to about 1 m and seem to match Milankovitch orbital climatic cyclicity with periods of ~400ky, ~100ky, ~40ky, and ~20ky. The composition and thickness ratio of the cycles suggest that the measured section of the Malmros Klint Member and the Carlsberg Fjord beds represents lacustrine accumulation over about 4 m.y., a duration consistent with the magnetostratigraphic correlations.

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Introduction

A series of continental basins that developed in the initial stages of rifting of the Pangea supercontinent are preserved along the margins of the continents bordering the North Atlantic. The basins are often fault bounded and filled with predominantly continental (fluvial and lacustrine) sediments of Triassic to early Jurassic age. A detailed lithostratigraphic, magnetostratigraphic and cyclostratigraphic record is now available for the Newark rift basin from eastern North America (Olsen et al., 1995; Kent et al., 1995; Olsen & Kent, 1995). A continuous sedimentary sequence more than 5000 m thick and encompassing about 30 m.y. of Late Triassic to earliest Jurassic history is represented in the Newark Basin and provides a reference section for chronostratigraphy in a tropical paleo-continental setting. Magnetic polarity stratigraphies for the Triassic are also being developed from Tethyan limestones keyed to standard biozonations (e.g., Gallet et al., 1993, 1994; Muttoni et al., 1995), providing the possibility of improved correlation between marine and continental realms and the development of an integrated Triassic time scale.

One of the more northerly representatives of the early Mesozoic basins is in the Jameson Land Basin of East Greenland where a thick succession of mostly continental sediments of Triassic age is exposed (Clemmensen, 1980a, b) (Fig. 1). Four formations are recognized in the Scoresby Land Group (Grasmück & Trümpy, 1969; Perch-Nielsen et al., 1974; Clemmensen, 1980a): the marine Wordie Creek Formation of Scythian age, succeeded by the predominantly continental Pingo Dal Formation of ?Early Triassic age, Gipsdalen Formation of Middle to Late Triassic age, and the Fleming Fjord Formation of Late Triassic age. The Scoresby Land Group is overlain by the Rhaetian-Sinemurian Kap Stewart Formation. The lithostrati-

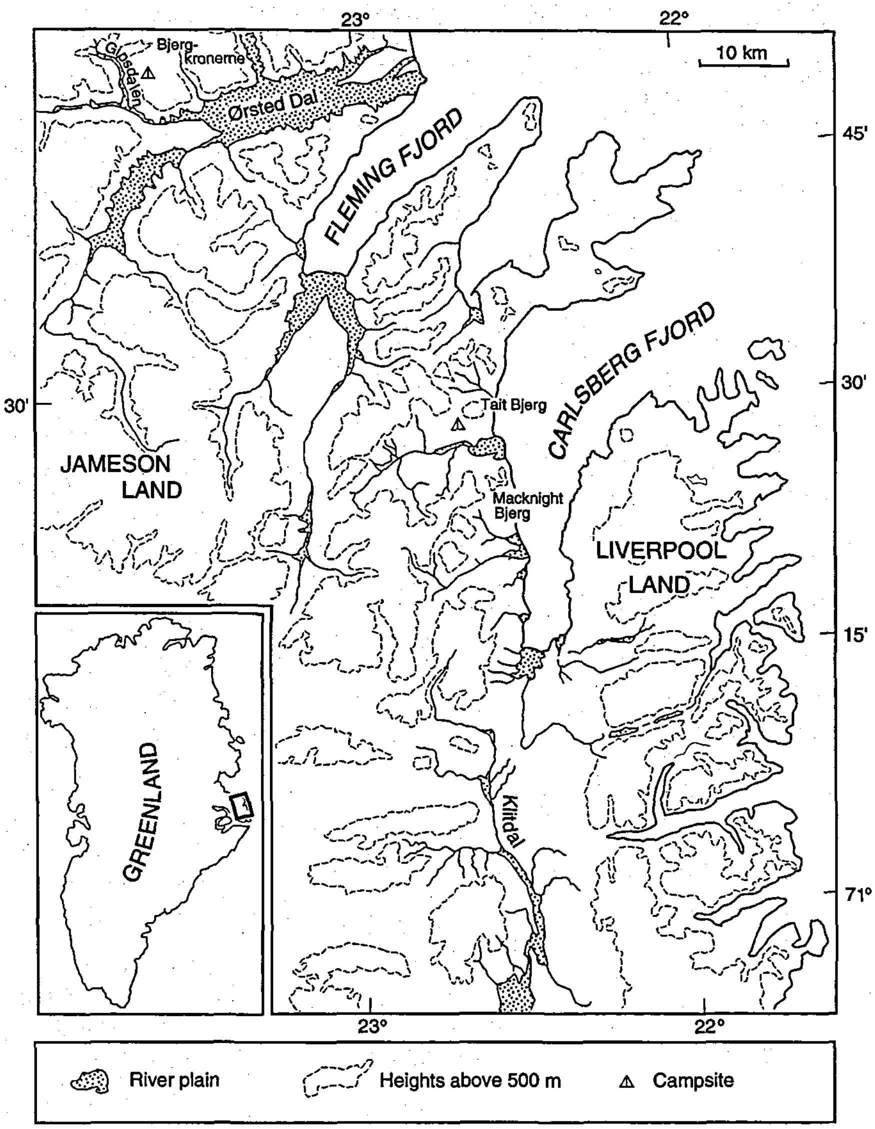


Fig. 1. Location map of sampling areas of Fleming Fjord Formation (Tait Bjerg) and Gipsdalen Formation (Gipsdalen) in East Greenland.

graphic subdivision of Middle and Upper Triassic rocks in the Jameson Land Basin is summarized in Fig. 2.

In the 1992 field season, we collected samples for paleomagnetic study and analyzed the cycle stratigraphy of a 210 m section of the Fleming Fjord Formation in the Tait Bjerg area of the Jameson Land Basin of East Greenland. A principal objective was to develop a cyclostratigraphy and magnetic polarity stratigraphy for paleoenvironmental interpretation and chronostratigraphic correlation. Paleomagnetic samples were also collected from the Gipsdalen Formation in the Gipsdalen region for the determination of a paleopole position for additional paleogeographic constraints.

Geological setting

The Jameson Land Basin is situated at the southern end of the East Greenland rift system. It is bounded to the west by a major north-south trending fault, to the east by the Liverpool Land area, and to the north by a cross-fault in Kong Oscar Fjord (Clemmensen, 1980b; Surlyk, 1990).

The Jameson Land Basin was the site of a complex suite of continental environments during deposition of the Gipsdalen and Fleming Fjord Formations. The Gipsdalen Formation consists of gypsum-bearing, primarily continental deposits and is subdivided into a basal Kolledalen Member (aeolian dunefields and sabkhas), the in part lateral equivalent Solfaldsdal Member (floodplains, saline lakes and sabkhas), and the Kap

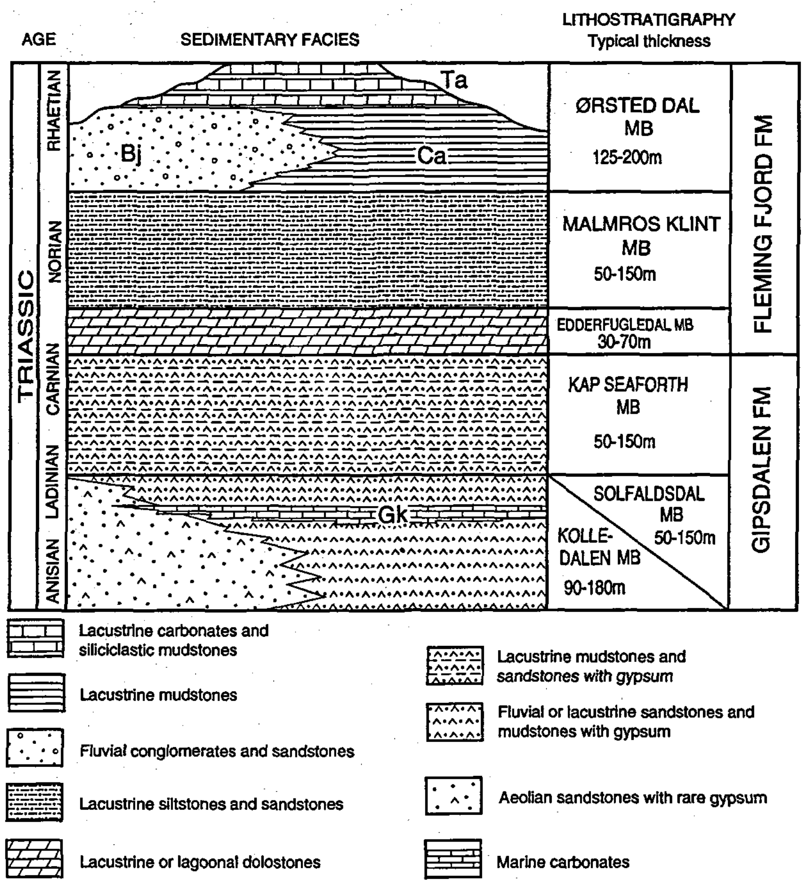


Fig. 2. Generalized stratigraphic column for Middle and Upper Triassic rock units in Jameson Land Basin, East Greenland (based on Clemmensen, 1980a, b; Jenkins et al., 1994). The Late Triassic Fleming Fjord Formation is overlain by the lacustrine Rhaetian-Sinemurian Kap Stewart Formation with an unconformity developed at the basin margins (Dam & Surlyk, 1993). Gk, Gråklint Beds; Ca, Carlsberg Fjord beds; Bj, Bjergkronerne beds; Ta, Tait Bjerg Beds.

Seaforth Member (saline lakes and sabkhas). Limestones and dark mudstones in the Solfaldsdal Member are recognized as the Gråklint Beds. The aeolian dunes in the formation formed under the influence of alternating NNE and SSE paleowinds (present coordinates). The overlying Fleming Fjord Formation is composed of a basal Edderfugledal Member (lake or lagoon), the Malmros Klint Member (mudflats and freshwater lakes), and the Ørsted Dal Member (rivers, mudflats, and freshwater lakes) (Fig. 2). The lake system evolved from saline and ephemeral (Gipsdalen Formation) to freshwater and very shallow (main part of Fleming Fjord Formation) to freshwater and relatively deep (uppermost Fleming Fjord Formation).

Age estimates of the Gipsdalen and Fleming Fjord Formations were originally based on the sparse occurrences of invertebrate fossils and palynomorphs (cf. Grasmück & Trümpy, 1969; Clemmensen, 1980a). Bivalves, gastropods and conchostracans in the Gråklint Beds of the Gipsdalen Formation indicate a Middle Triassic (Anisian-Ladinian) age. Palynomorphs in the overlying Kap Seaforth Member suggest that this part of the Gipsdalen Formation is of Late Triassic (?Carnian) age. Conchostracans in the Edderfugledal and Malmros Klint Members of the Fleming Fjord Formation suggest a Late Triassic (?Carnian to ?Norian) age, while ostracods, pelecypods, and palynomorphs in the Tait Bjerg Beds indicate that the upper-most part of the Fleming Fjord Formation is Rhaetian in age. A diverse assemblage of terrestrial vertebrates, including several species of mammals, dinosaurs, aetosaurs, turtles, amphibians and fishes has recently been described from the Malmros Klint and Ørsted Dal Members of the Fleming Fjord Formation (Jenkins et al., 1994). The assemblage is described as almost identical to well-known European faunas of Norian age, supporting also the paleogeographic proximity between Greenland and Europe in the Late Triassic.

Paleomagnetic results

Fleming Fjord Formation

Oriented hand samples or drill cores were collected from 74 stratigraphic levels distributed over an approximately 210 m section of the Fleming Fjord Formation in the Tait Bjerg area (71.47° N 22.67° W) (Fig. 1). The measured section was the same as that studied for

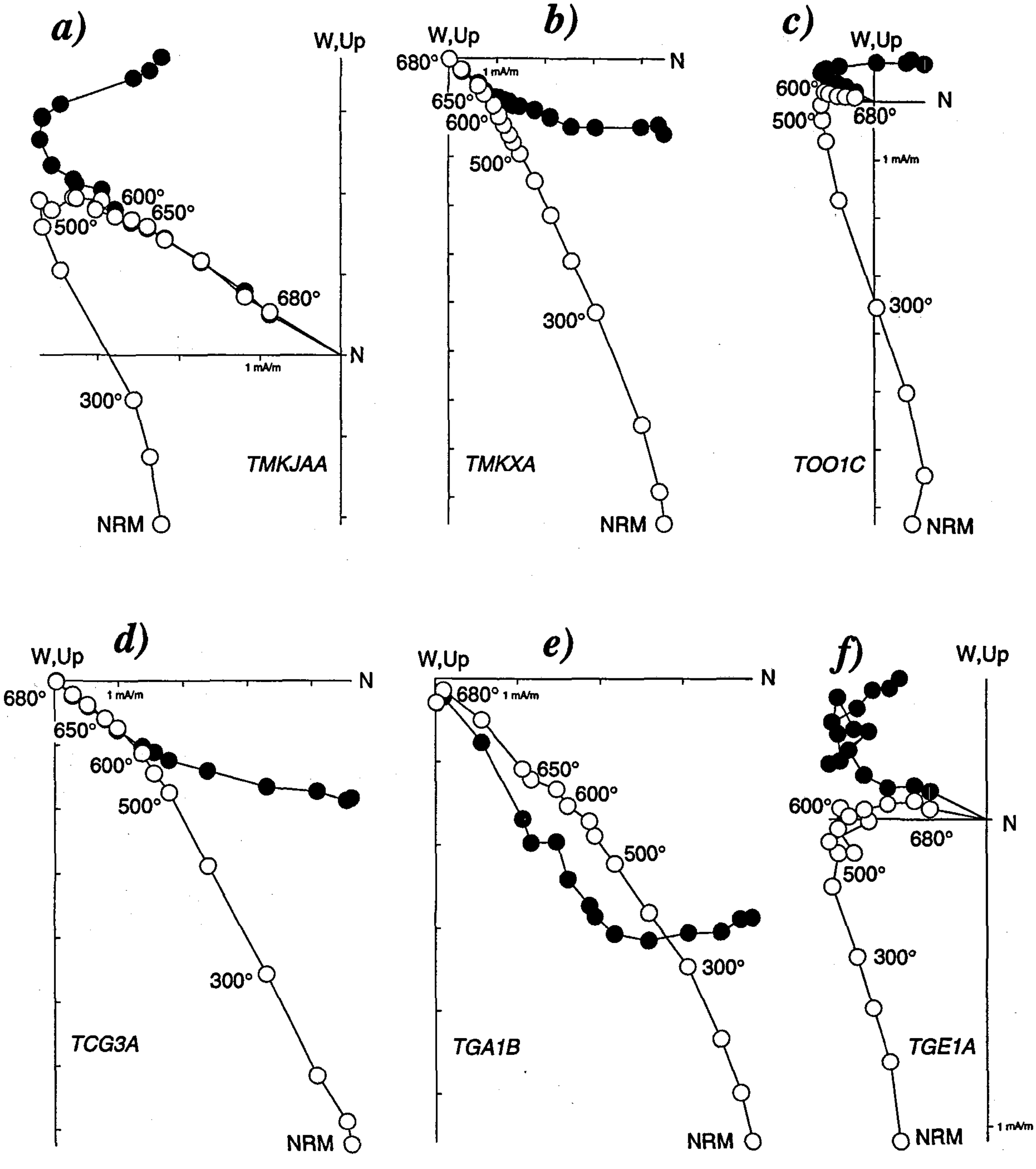
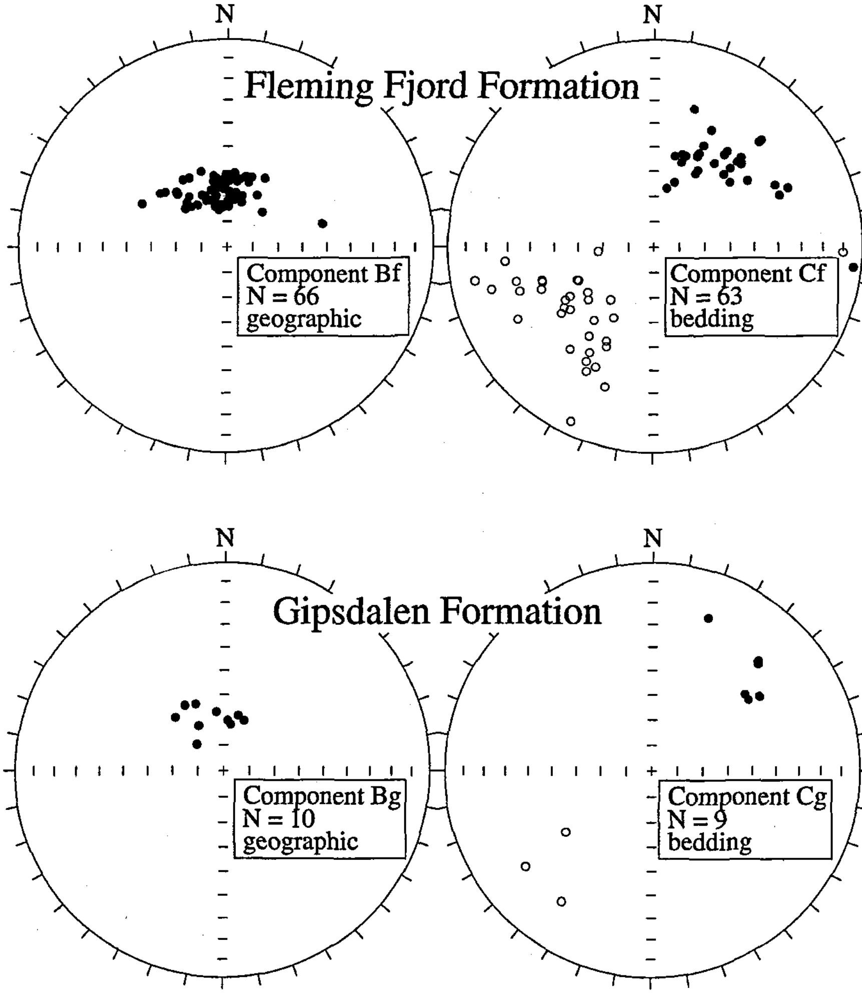


Fig. 3. Vector end-point diagrams of progressive thermal demagnetization of representative samples from the (a, b) Malmros Klint Member and (c, d) Ørsted Dal Member of the Fleming Fjord Fm., and (e, f) the Gipsdalen Formation. Open/filled symbols are projections of the magnetization vector on the vertical/horizontal planes in geographic coordinates; demagnetization temperatures in degrees Celcius.

Fig. 4. Site-mean directions of magnetization components from the Fleming Fjord and Gipsdalen Formations. Open/filled symbols are projections on upper/lower hemisphere of equal-area projections.



cyclostratigraphy and encompassed the entire Malmros Klint Member (130 m) and the lower part (80 m) of the overlying Carlsberg Fjord beds of the Ørsted Dal Member. The beds are nearly flatlying with westerly dips of less than 10°. After measurement of the natural remanent magnetization (NRM), three or more specimens from each level (total of 202 specimens) were subjected to progressive thermal demagnetization in a minimum of 10 temperature steps to 680° C. NRM directions typically are northerly and steeply down (mean of D = 347.6° I = 71.4°), suggesting severe overprinting, but thermal demagnetization was effective in resolving the underlying components of magnetization.

The NRM tended to be dominated by a component with unblocking temperatures concentrated from about 300° to 600° C and a northerly and downward direction (Fig. 3a–d). This component, labelled Bf, was calculated using principal component analysis in 66 sampling levels (169 specimens) which give a mean direction of D = 351.6° I = 67.8° $a95$ = 2.5° in geographic coordinates (Fig. 4a; Table 1). The Bf direction is significantly shallower than the present geomag-netic or dipole field (inclination ~80°); despite the relatively high unblocking temperatures, the uniform (normal) polarity nevertheless suggests that the Bf component is an overprint but of more ancient origin.

A relatively small but in most cases resolvable component of magnetization is revealed over the highest unblocking temperatures, typically from 650° to 680° C, and aligned along either a northeasterly and down (Fig. 3b, d) or southwesterly and up (Fig. 3a, c) direction in different sampling levels. This magnetization, labelled Cf, was successfully isolated in 63 sampling sites (156 specimens) and is regarded as the characteristic component. The mean directions in bedding coordinates are D = 41.7° I = 45.2° $a95$ = 7.3° (N=31 sites) for the northeasterly (normal polarity) population and D = 231.7° I = –43.9° $a95$ = 6.7° (N=32 sites) for the southwesterly (reversed polarity) population (Fig. 4b). The normal and reversed polarity directions deviate from being antipodal by only 7.2°, which is not significant, indicating that the pervasive overprint does

Tabel 1. Mean direction of magnetizations from the Fleming Fjord and Gibsdalen Formations.

		Geographic				Bedding				Pole Position		
Component	*Sites*	*k*	*a95*	*D*	*I*	*k*	*a95*	*D*	*I*	*Lat*	*Lon*	*A95*
					Fleming Fjord Formation							
NRM	74	12.1	4.9°	347.6°	71.4°	11.8	5.0°	338.7°	72.0°			
Bf	66	50.5	2.5	351.6	67.8	49.2	2.5	339.8	69.3	68.9°	172.2°	3.9°
Cf (N)	31	13.7	7.2	42.7	39.3	13.4	7.3	41.7	45.2			
Cf (R)	32	15.4	6.7	231.7	−39.6	15.6	6.7	231.7	−43.9			
Cf (N&R)	63	14.4	4.9	47.3	39.5	14.3	4.9	46.9	44.7	37.9	101.2	5.2
					Gipsdalen Formation							
NRM	10	9.9	16.1	8.1	64.5	14.6	13.1	19.5	67.1			
Bg	10	63.8	6.1	345.8	68.9	44.8	7.3	355.3	72.6	69.6	182.0	9.6
Cg (N)	6	22.8	12.7	30.9	41.8	34.9	11.5	44.2	34.0			
Cg (R)	3	11.4	38.4	221.0	−37.9	24.5	25.5	226.7	−32.8			
Cg (N&R)	9	22.0	11.2	41.6	33.1	35.3	8.8	45.1	33.6	30.6	105.2	8.1

N is normal and R is reversed polarity, *k* is estimate of precision parameter, *a95* is radius of 95% confidence circle, *D* is declination and *I* is inclination of mean directions, *Lat* and *Lon* are the north latitude and east longitude of the north paleomagnetic pole corresponding to the underlined mean directions, and *A95* is radius of 95% confidence circle about the pole position.

not systematically bias the directions. The overall mean direction after inverting the directions of the southwesterly population is $D = 46.9°$ I = 44.7° $a95 = 4.9°$ (N = 63 sites) in bedding coordinates; the dispersion and mean direction are not significantly different in geographic coordinates because of the small bedding tilts (Table 1).

Gipsdalen Formation

Oriented drill core samples were collected from two different stratigraphic intervals in the Gipsdalen Formation in the Gipsdalen area (71.83° N 23.51°W) (Fig. 1). Five sites were taken from the basal part of the Kolledalen Member and distributed from about 25 to 44 m above the local base of the Gipsdalen Formation; bedding has southwesterly dips of about 16°. Another 5 sampling sites were taken in the lower part of the Solfaldsdal Member and about 180 to 193 m above the base of the Gipsdalen Formation; bedding has northeasterly dips of about 11°. The Gipsdalen rocks (mean NRM=4 mA/m, median NRM=2.5 mA/m) are more weakly magnetized than the Fleming Fjord sediments (mean NRM=10 mA/m, median NRM 9 mA/m) but they share with the Fleming Fjord a similar steeply dipping downward direction of NRM (mean D=8.1° I=64.5°) due to a strong and pervasive overprint. A total of 56 samples were subjected to progressive thermal demagnetization in 10 to 15 steps to 680° C for component analysis.

The overprint (component Bg) is typically resolved over the 200° to 500° C unblocking temperature range (Fig. 3e, f) and was recovered in 51 of the samples. The mean direction in geographic coordinates for the 10 sites is $D = 345.8°$ $I = 68.9°$ $a95 = 6.1°$ (Fig. 4c), not significantly different from the Bg overprint direction found in the Fleming Fjord (Table 1).

A dual polarity magnetization with high unblocking temperature (~600 to 680° C; Fig. 3e, f) was recovered in 41 samples from 9 sites. The magnetization, regarded as characteristic and labelled Cg, has a northeasterly and down (normal polarity) direction in 6 sites, and a southwesterly and up (reversed polarity) direction in 3 sites (Fig. 4d). The rejected site (36 m level from the Kolledalen Member) showed noisy demagnetization trajectories trending toward the southwest and up quadrant but was too heavily overprinted to adequately resolve the reversed polarity direction. The normal and reversed polarity site mean directions are within 2.4° of antipodal, providing a positive reversal test. Converted to common polarity, the overall mean direction in bedding coordinates is $D = 45.1°$ $I = 33.6°$ $a95 = 8.8°$ (N=9 sites). Dispersion is less after tilt correction although the ~50% increase in precision parameter is not significant at the 95% confidence level (Table 1). The Gipsdalen Cg direction has virtually the same declination but an inclination that is about 10° shallower compared to the Fleming Fjord Cf direction.

Paleomagnetic poles

In an early paleomagnetic study of Triassic rocks in East Greenland, Bidgood & Harland (1961) obtained a mean direction of $D = 358°$ $I = 68°$ (pole position at 69°

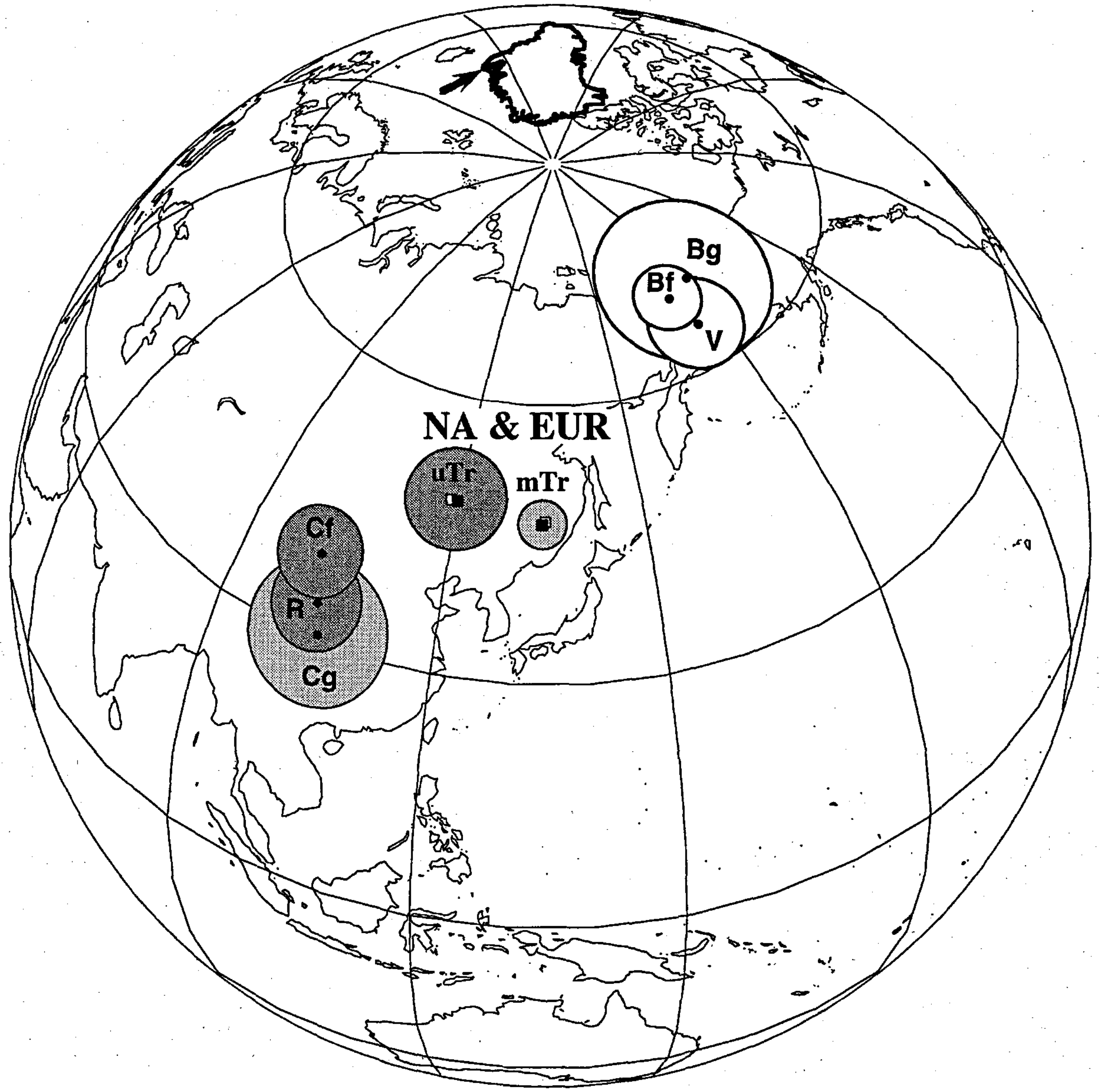


Fig. 5. Paleomagnetic poles from the Fleming Fjord and Gipsdalen formations of East Greenland compared to Triassic poles from North America and Europe. Bf, Fleming Fjord overprint direction; Bg, Gipsdalen overprint direction; V, mean pole for early Cenozoic volcanics and dikes from East Greenland (compilation in Van der Voo, 1993); Cf, Fleming Fjord characteristic direction; Cg, Gipsdalen characteristic direction; R, Fleming Fjord pole of Reeve et al. (1974). Mean Late Triassic (uTr) and Early/Middle Triassic (mTr) poles for North America (solid squares) and Europe (open squares) from Van der Voo (1990) transferred to Greenland coordinates according to rotation parameters of Bullard et al. (1965). Continents shown for reference in present positions; arrow points to sampling locality of Fleming Fjord and Gipsdalen Formations in East Greenland.

N 160° E) based on NRM measurements on 44 specimens of flat-lying red sandstones and marls from undifferentiated middle and upper Triassic units at Kap Biot. This is very similar to the overprint magnetization as well as the NRM directions obtained here for the Fleming Fjord and Gipsdalen Formations. Athavale & Sharma (1974) reported preliminary paleomagnetic results from eleven samples of fine grained red sandstones from the Pingo Dal and Gipsdalen Formations from the Jameson Land area of East Greenland. Alternating field demagnetization was employed but it was clearly inadequate to isolate a characteristic magnetization at least in the 5 Gipsdalen samples which after 100 mT treatment gave a mean direction of D = 1° I = 62° (recalculated pole position at 61.2° N 155.5° E; the quoted pole position of 49° N 158° E does not correspond to the mean magnetization direction). This is again similar to the overprint magnetization direction obtained here. Based on only 6 samples, the paleo-magnetic results reported by Athavale & Sharma (1974) for the Pingo Dal Formation, with mean direction of D = 336° I = 33°, an apparent paleolatitude of 18° and corresponding paleopole located at 35° N 184° E, are considered of low reliability and uncertain significance (e.g., excluded in pole compilation by Van der Voo, 1993).

The paleomagnetic study by Reeve et al. (1974) of a 12 m drill core from the lower middle part of the Malmros Klint Member of the Fleming Fjord Formation in Jameson Land (71.33° N 23.70° W) employed thermal demagnetization and provides more useful information. The removal of a large normal polarity component overprint revealed a high unblocking temperature magnetization of normal and reversed polarity. A magnetic polarity stratigraphy was developed from measurements on 134 samples from the core; 30 samples with a reversed polarity high unblocking temperature component were deemed suitable for calculation of a paleomagnetic pole. The paleopole, located at 34° N 103.2° E, is confirmed by the present study of the Fleming Fjord Formation (Fig. 5).

With regard to the origin of the overprint magnetizations, Reeve et al. (1974) already suggested that the NRM directions obtained by Bidgood & Harland (1961) are likely to be dominated by a Cenozoic overprint, perhaps associated with widespread igneous activity in East Greenland during the early Cenozoic. Van der Voo (1993, Table A3) lists 8 paleomagnetic results from early Cenozoic dikes and lavas in East Greenland; the mean pole position is 65.2° N 174.9° E (K = 110, $A95$ = 5.3°). The overprint magnetization paleopoles from the Fleming Fjord and Gipsdalen Formations are virtually identical to the mean early Cenozoic pole from Greenland (Fig. 5), suggesting that the overprints were acquired at about the time of the igneous activity as thermochemical magneti-zations. Although both normal and reversed polarity directions are observed in the Early Cenozoic igneous rocks, the presumably related overprint in the Triassic sediments predominantly reflects only the normal polarity state.

Prior to the opening of the Labrador and Norwegian Seas in the later Mesozoic and Cenozoic, Greenland was adjacent to North America and Europe as part of the northern supercontinent of Euramerica. Recent analyses by Van der Voo (1990, 1993) show that Late Triassic and Early/Middle Triassic mean paleomagnetic poles for North America and Europe agree well in a reconstructed framework, especially in the fit of Bullard et al. (1965), hence there is the expectation that Triassic poles from the intervening Greenland continent should agree with those from North America and Europe. However, the paleopoles from the Fleming Fjord Formation (Reeve et al., 1974; this study) and the Gipsdalen Formation (this study) are systematically more southerly with respect to the mean Triassic poles from North America and Europe in a reconstructed framework (Fig. 5).

Late Triassic poles from Europe are not well grouped ($A95$=14°; Van der Voo [1990]) but the mean Late Triassic pole position for North America (52° N 96° E $A95$ = 5°; Van der Voo [1990]) is well defined and supported by more recent results and analyses (Kent & Witte, 1993; Kent et al., 1995). The mean Late Triassic pole for North America transferred to Greenland coordinates according to the reconstruction parameters of Bullard et al. (1965) (calculated rotation of 17.99° about a Euler pole at 70.53° N 94.34° W) predicts a paleolatitude of 34.1° N for the sampling locality in East Greenland; this compares with 26.3° N determined in our study of the Fleming Fjord Formation, giving a significant paleolatitudinal difference of 7.8° ± 5.7° (estimation of errors using method of Demarest [1983]). The observed mean declination (46.9°) is also more clockwise than predicted (28.4°), with a difference of 18.5° ± 11.6°. Using the more recently published reconstruction parameters of Roest & Srivastava [1989] (rotation of 13.78° about a Euler pole at 67.5° N 118.48° W) only slightly reduces the paleolatitudinal discordance to 7.6° ± 5.7° although the difference of 13.2° ± 11.6° between observed and predicted declinations becomes appreciably smaller.

Similarly, the mean Early/Middle Triassic paleopole for North America (52° N 110° E $A95$ = 3°; Van der Voo, 1990) predicts a more northerly paleolatitude of 31.5° N (31.3° N with rotation parameters of Roest & Srivastava [1989]) for the sampling locality in East Greenland, compared to 18.4° N determined in our study of the Gipsdalen Formation. The paleolatitudinal discordance of 13.1° ± 7° (12.9° ± 7° with Roest & Srivastava parameters) is statistically significant. The mean declination for the Gipsdalen Formation (45.1°) is also more clockwise than predicted (18.2° for Bullard, 23.5° for Roest & Srivastava parameters), with differences of 26.9° ± 22° and 21.6° ± 22° using the respective reconstruction parameters. At least the upper part of the Gipsdalen Formation may be Late Triassic rather

Fig. 6. Pangea in the Late Triassic. The northern landmass of Euromerica (including Greenland) is reconstructed according to Bullard et al. (1965) and positioned according to a combined mean Late Triassic pole from North America and Europe (51° N 96° E in North American coordinates; Van der Voo, 1993); the southern landmass of Gondwana is reconstructed according to Smith & Hallam (1970) and is in a Pangea A2 relationship with respect to Euromerica (Van der Voo et al., 1984). The *paleolatitude of the Fleming Fjord Cf direction (26° N) at sampling locality in East Greenland (star) is shown for comparison.*

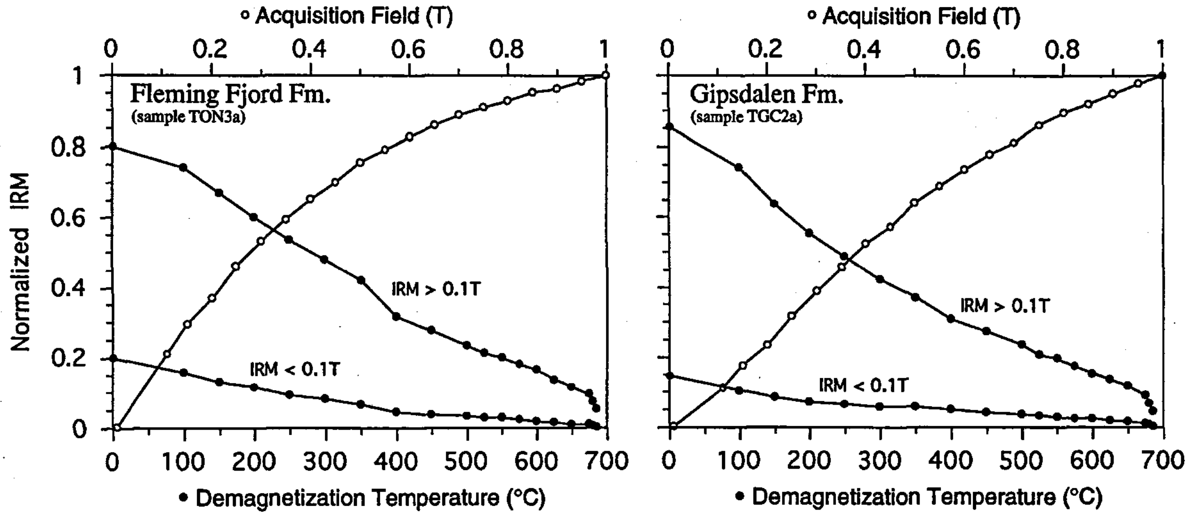


Fig. 7. Acquisition of isothermal remanent magnetization and subsequent thermal demagnetization of low coercivity (<0.1T) and high coercivity (>0.1T) orthogonal fractions according to method of Lowrie (1990), for a representative sample from the Fleming Fjord Formation (left) and from the Gipsdalen Formation (right).

than Middle Triassic in age and thus a comparison with predicted directions from the mean Late Triassic pole position for North America may be more appropriate. Using the Roest & Srivastava reconstruction parameters, this results in a somewhat larger paleolatitudinal difference of 15.9° ± 7.7° but a much reduced declination difference of 11.9° ± 22.6° that is not significant.

The differences between observed and predicted declinations, where significant, conceivably could be attributed to local clockwise tectonic rotations or uncertainties in the reconstruction of Greenland to North America. However, the shallow characteristic inclinations in the Fleming Fjord and Gipsdalen Formations which give paleolatitudes that are consistently about 10° lower than predicted are more difficult to attribute to tectonic processes or uncer-tainties in reconstruction. A strict interpretation of the paleomag-netic data would suggest that in the Triassic, Greenland was something like 1000 km farther south with respect to North America and Europe in a Pangea fit (Fig. 6). This seems improbable in the context of geologic and paleomagnetic constraints on the position of the other continents even in alternative Pangea reconstructions (e.g., Morel & Irving,1981). We therefore consider the possibility that the characteristic magnetization directions of the Greenland Triassic sediments are systematically biased by the recording process.

Rock magnetic properties

The anomalously shallow characteristic magnetization directions in the Fleming Fjord and Gipsdalen Formations may be due to a sedimentary inclination error produced during or shortly after deposition. Inclination error is the difference between expected (Ie) and observed (Io) inclinations which are related as: $f = \tan(Io)/\tan(Ie)$ for detrital remanent magnetizations (King, 1955). The flattening factor, f, can range from 0 (complete shallowing) to 1 (no shallowing). The presence of a systematic inclination error in ancient red beds has long been disputed but generally thought to be unimportant (e.g., Irving, 1967; Creer, 1967; Van der Voo et al., 1995). However, detrital remanent magnetization carried by hematite in laboratory redeposition experiments as well as in some modern natural deposits has been associated with an inclination error characterized by f values of about 0.5 for ambient field inclinations up to 60° (Tauxe & Kent, 1984; Lovlie & Torsvik, 1984). For the Fleming Fjord and Gipsdalen Formations, the flattening factor would need to be only about 0.6 if inclination error was responsible for the anomalously shallow directions.

The reddish color of the sampled Fleming Fjord and Gipsdalen sediments and the high maximum unblocking temperatures of their NRM point to hematite as the principal magnetic mineral. This is confirmed by acquisition and thermal demagnetization of isothermal remament magnetization experiments (Fig. 7) which show that a saturation remanence is not achieved by 1 T, and both the small low coercivity fraction as well as the comparatively large high coercivity fraction have maximum unblocking temperatures of about 685° C, properties that are again indicative of the predominance of hematite as the magnetic carrier in these rocks.

A primary magnetic fabric, characterized by an oblate magnetic susceptibility ellipsoid with near-vertical minimum axis, is associated with a detrital remanence carried by hematite (e.g., Lovlie & Torsvik, 1984); conversely, highly disturbed fabrics are more likely to

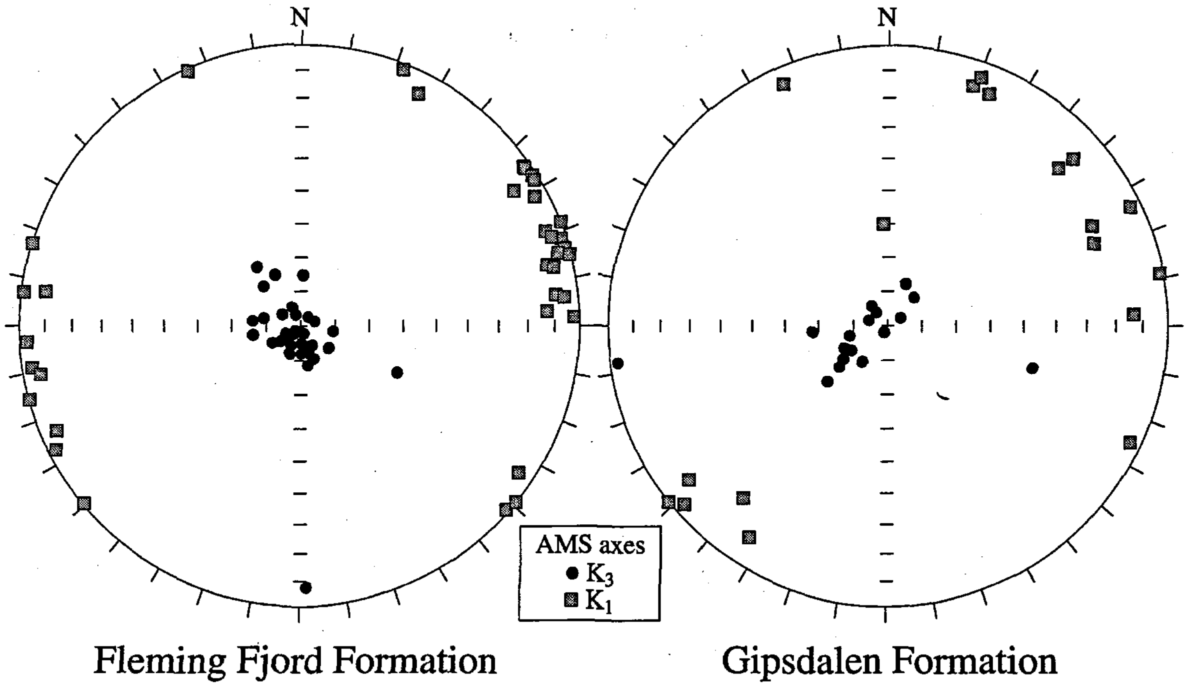


Fig. 8. Minimum (K_3) and maximum (K_1) magnetic susceptibility axes for samples from the Fleming Fjord and Gipsdalen Formations plotted in bedding coordinates on lower hemisphere of equal-area projections.

be associated with a post-depositional magnetization without inclination error (Tauxe et al., 1990). Magnetic fabric can thus provide a basis to assess whether the characteristic magnetizations of the Fleming Fjord and Gipsdalen Formations are of detrital origin and therefore likely to have been affected by sedimentary inclination error.

To characterize the magnetic fabrics (Ellwood et al., 1988), a Kappabridge model KLY-2 was used to measure the anisotropy of magnetic susceptibility (AMS) on 35 specimens from the Fleming Fjord Formation and 18 from the Gipsdalen Formation that were not thermally demagnetized. Bulk susceptibilities average about $75 * 10^{-6}$ SI, with the degree of anisotropy (K_1/K_3-1) averaging about 0.035 in the Fleming Fjord and 0.050 in the Gipsdalen. Most of the specimens show evidence of a primary magnetic fabric: the susceptibility ellipsoids are typically oblate (foliation [K_2K_3-1] exceeds lineation [K_1/K_2-1] on average by a factor of 7 in the Fleming Fjord and 11 in the Gipsdalen), and the minimum susceptibility (K_3) axes are preferentially grouped near to vertical whereas the maximum (K_1) and intermediate (K_2) susceptibility axes tend to lie in the bedding plane (Fig. 8). If the AMS measurements reflect the hematite carrier of the characteristic magnetizations, the observed magnetic fabrics are consistent with a detrital origin and allow the possibility of inclination error as an explanation for the shallowed characteristic directions.

Magnetostratigraphy and cycle stratigraphy

Although sedimentary inclination error complicates paleogeographic interpretations of the paleomagnetic data, it provides supportive evidence that acquisition of the characteristic magnetizations was closely linked to the depositional history. In the sampled part of the Fleming Fjord Formation, seven polarity intervals are delineated by two or more consecutive sampling levels (Fig. 9). A normal-reversed-normal polarity magnetozone sequence (labeled as F1n, F1r, F2n) occurs in the lower half of the Malmros Klint Member, followed by a thick (~85m) reversed polarity magnetozone (F2r) that extends into the lower part of the Ørsted Dal Member. The remaining part of the Ørsted Dal Member that was sampled has another normal-reversed-normal polarity sequence (F3n, F3r, F4n) with a single-sampling level reversed polarity interval within magnetozone F3n. The magnetostratigraphic section in the Malmros Klint Member reported by Reeve et al. (1974), whose distinctive feature is a ~6.5 m reversed polarity interval followed by a ~4 m normal polarity interval, most likely correlates to magneto-zones F1r and F2n because there are no other reversed to normal polarity transitions in our magnetostra-tigraphic section of the entire Malmros Klint Member.

The average duration of Late Triassic polarity intervals as documented in the Newark reference section is about 0.5 m.y. In the Fleming Fjord magnetostratigraphy, the presence of seven polarity intervals (or per-

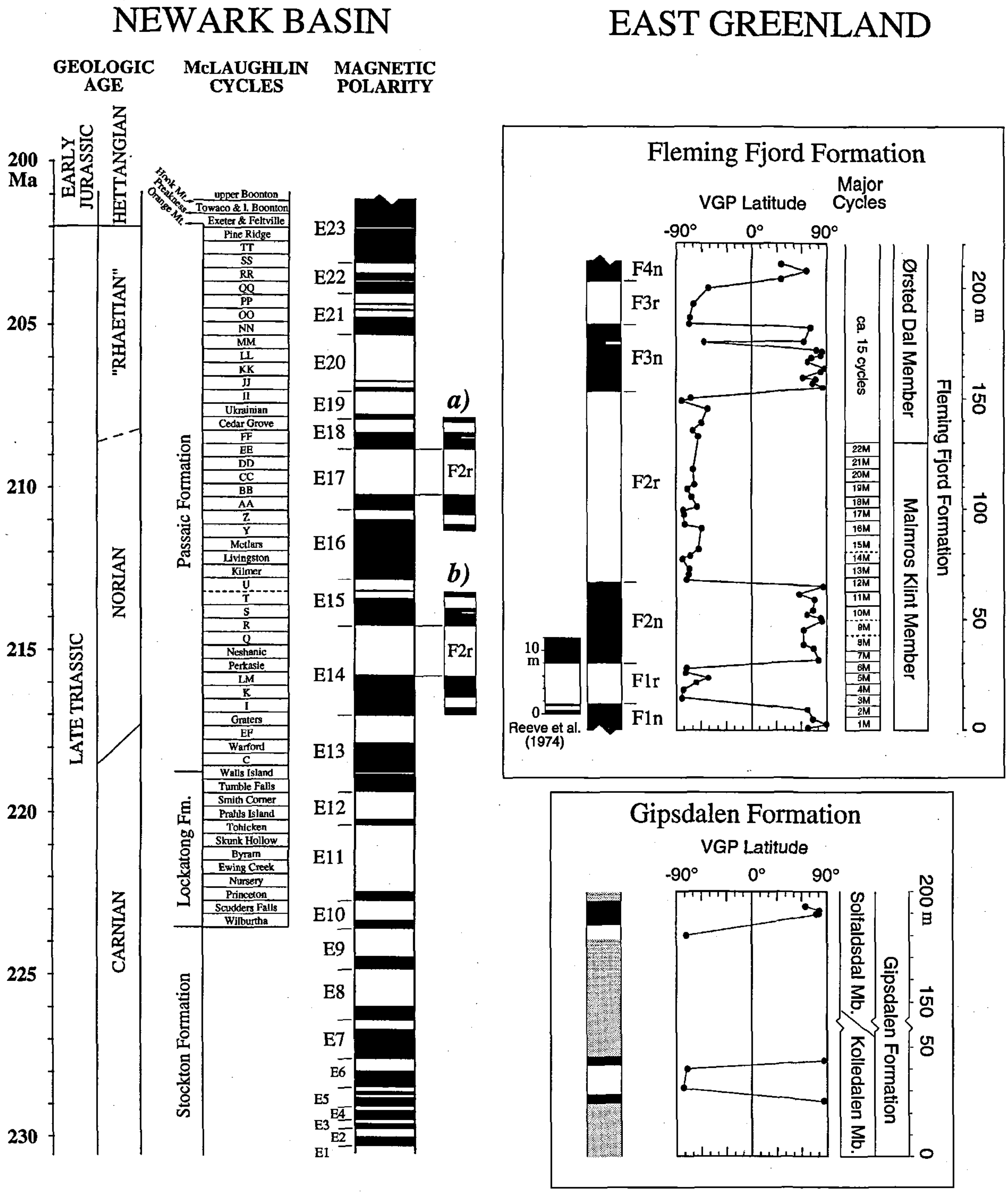


Fig. 9. Magnetic polarity stratigraphy and major climate cycles of the Tait Bjerg section of the Fleming Fjord Formation from East Greenland (upper right). A correlation to the Tait Bjerg section of a magnetostratigraphy reported by Reeve et al. (1974) from a 12 m core hole in the Malmros Klint Member is also shown. Magnetic polarity intervals (filled/open are normal/reversed polarity) are interpreted from the latitude of the virtual geomagnetic pole (VGP) calculated from the characteristic magnetization at each sampling level relative to the mean north paleomagnetic pole. Note that the McLaughlin climate cycles in the Newark Basin reference section (left; Kent et al., 1995) represent the 413 k.y. Milankovitch periodicity (Olsen & Kent, 1995), whereas the major climate cycles shown for the Fleming Fjord Formation are interpreted to represent the ~100 k.y. Milankovitch periodicity. Two options are shown for the correlation of the polarity sequence of the Tait Bjerg section to the Newark geomagnetic polarity time scale, based on scaling magnetozone F2r to either a) E17r, or b) E14r; the first option is preferred. Lower right: Preliminary magnetic polarity stratigraphy for the Gipsdalen Formation from the Gipsdalen area.

haps nine if the single level reversal within magnetozone F3n is included) would thus suggest that the sampled section represents roughly 4 m.y. and accumulated at an average rate of about 50m/m.y. This estimate can be compared with the timing deduced from analysis of the cyclostratigraphy.

The Malmros Klint Member and the basal part of the overlying Ørsted Dal Member were the subject of a cyclostratigraphic study in the 1992 field season. The results of the analysis will be described in more detail elsewhere but the main findings are summarized here. The Malmros Klint Member is composed of cyclically bedded red-brown lacustrine mudstones and fine-grained sandstones, and palaeosols. These deposits form vertical cliffs 100-130 m high along Carlsberg Fjord in the eastern part of the basin. The member displays a composite cyclicity; based on current data the most obvious cycles have mean thicknesses of (25m), 5.3 to 6.5m, 2.1m, and 1.3 to 1.6m. The cycles with a thickness of 5.3 to 6.5 m are clearly visible in the landscape as major step-like ledges (cf. Jenkins et al., 1994); in the Malmros Klint Member, 22 of these cycles occur in superposition (Fig. 9). The composite cyclicity is thought to record variations in sediment yield to the basin and lake environment due to orbital forcing of climate (precipitation). Small-scale cycles between 0.25 and 0.6 m in thickness may record short-term climatic variation, or autocyclic processes. The composite cyclicity seems to match Triassic orbital cycles with the following periods: (413 ky), 95–123 ky, 36 ky, 18–21.5 ky (Berger et al., 1992). If this interpretation is correct, the Malmros Klint Member accumulated in a little more than 2 m.y.

The overlying Ørsted Dal Member is composed of two units in the eastern part of the basin. The lowermost Carlsberg Fjord beds are up to 115 m thick and composed of cyclically bedded variegated lacustrine claystones, and calcareous siltstones. This unit is also characterized by a composite cyclicity with cycles varying in thickness between 0.9 and 6.3 m. As in the Malmros Klint Member, cycles with thicknesses around 5 to 6 m are prominent in the field; the Carlsberg Fjord beds are composed of ca. 15 of these cycles. Thus it appears that depositional conditions in the lake system continued to be controlled by Milankovitch-type fluctuations in climate. Although data are not very complete at present, the composition and thickness ratio of the cycles suggest that the Carlsberg Fjord Beds represent a little less than 2 m.y. of accumulation. The uppermost Tait Bjerg Beds in the Ørsted Dal Member is composed of cyclically bedded lacu-strine limestones and clastic mudstones but have not yet been studied in detail.

In attempting to correlate the Fleming Fjord section more precisely with the Newark, we suggest that two most likely alternatives are that the long reversed magnetozone F2r of the Fleming Fjord section matches either magnetochron E17r or E14r of the Newark sequence (Fig. 9). The first option provides a good overall correspondence of the polarity patterns and would imply that the Fleming Fjord section extends from the just within magnetochron E16n (ca. 207.5 Ma) to magnetochron E19n (ca. 211 Ma), representing a duration of about 3.5 m.y. in the late Norian. The magnetostratigraphies do not correspond as well in the second option, for example, magnetozone F2n is disproportionately short compared to magnetochron E14n. This correlation would imply that the Fleming Fjord section extends from just within magnetochron E13n (ca. 218 Ma) to just within magnetochron E16n (ca. 212.5 Ma), representing a duration of about 5.5 m.y. in the early Norian. Based on the available biostratigraphic age constraints and cyclostratigraphic duration estimates, which suggest that the Malmros Klint and Ørsted Dal members are at least in part late Norian in age (Jenkins et al., 1994) and the sampled portion represents about 4 m.y. or less, the first correlation option is preferred. This option would place the Malmros Klint and lower Ørsted Dal Members as time equivalent to the Sevatian, according to a suggested magnetostratigraphic correlation between Tethyan substages (Gallet et al., 1993) and the Newark sequence (Kent et al., 1995).

A strong cyclicity also characterizes much of the Gipsdalen Formation and especially the Solfaldsdal and Kap Seaforth Members (cf. Clemmensen, 1978; Clemmensen, 1980b). In the sampled section of the Solfaldsdal Member, cycles vary in thickness between ca. 1 and 5 m. The accessible exposures of the Gipsdalen Formation, however, only allowed a fragmen-tary magnetostratigraphic representation (Fig. 9). Normal and reversed polarity intervals are evident in both of the short sampled intervals of the Kolledalen and Solfaldsdal Members but it is not possible to correlate the very discontinuous polarity sequence to other Triassic magnetostratigraphies.

Conclusions

The characteristic magnetizations of the Fleming Fjord and Gipsdalen Formations give pole positions that are systematically displaced with respect to reference Triassic poles transferred from North America. There is better agreement between expected and observed declinations with the reconstruction parameters of Roest & Srivastava (1989) compared to those of Bullard et al. (1965), but the observed inclinations and inferred paleolatitudes are consistently shallower by ~10° compared with expected values with either of these or other continental reconstructions (e.g., Rowley & Lottes, 1988). The observation of primary magnetic fabrics in the Fleming Fjord and Gipsdalen sediments suggests that a probable cause for the discrepancy is sedimentary inclination error, whereby the detrital hematite carrier of the characteristic magnetizations has a tendency for the basal plane and the remanence aligned within it to rotate into the bedding plane. We note that sedimentary inclination error has been observed in some late Cenozoic hematite-bearing sediments (Tauxe & Kent, 1984; Garces et al., 1995) and may be more prevalent than usually supposed.

Comparison between the characteristic directions of the Gipsdalen and Fleming Fjord Formations suggests that East Greenland migrated northward by 7.9° ± 7.8°; correction for a systematic sedimentary inclination error corresponding to a flattening factor of ~0.6 on average in both formations would result in a northward shift a few degrees larger, to about 10°. The northward paleolatitudinal shift implied by the Gipsdalen and Fleming Fjord data for East Greenland is compatible with the sense of paleolatitudinal change that would be predicted from North American (or European) reference poles. For example, the more age-specific Carnian pole from the lower Newark in eastern North America (53.5° N 101.6° E A95=4.8°; Witte & Kent, 1989) and the mean Norian paleopole for North America (57.4° N 91.0° E A95=3.8°; Kent & Witte, 1993), predict a 5.7° ± 4.9° northward paleolatitudinal shift for East Greenland (using the Roest & Srivastana [1989] rotation parameters) over the Late Triassic. We cannot exclude the possibility that inclination shallowing has also affected the Traissic reference poles from North America which are mostly from sedimentary rocks. However, the relevant rock units from North America are from a lower latitudinal setting compared to East Greenland so that any inclination error for a given flattening factor is expected to have a smaller biasing effect on paleopole determination.

Lithofacies of the Gipsdalen and Fleming Fjord Formations have been previously interpreted to signify northward continental drift across paleoclimatic belts (Clemmensen, 1980b). In particular, arid conditions indicated by the presence of eolian deposits and evaporites in the Gipsdalen Formation are succeeded by wetter conditions indicated by freshwater lake and fluvial deposits that characterize the Fleming Fjord Formation. The coal-bearing rocks of the Rhaetian-Sinemurian Kap Stewart Formation extend this trend to more humid conditions. The stratigraphic succession of climate-sensitive rock facies may have resulted from the northward drift of East Greenland from the arid tropical into the more humid subtropical or temperate climate belt over the Late Triassic. However, the present data suggesting that the Jameson Land area was aligned more clockwise with respect to present coordinates will necessitate a reconsideration of the paleowind pattern during deposition of the Gipsdalen Formation.

On a more detailed level, the cyclostratigraphy of the Tait Bjerg section of the Fleming Fjord Formation shows a hierarchy of lithologic cycles which are interpreted to reflect Milankovitch control of depo-sitional environment and sediment supply. The cycles appear to correspond to climatic forcing at the full spectrum of orbital periodicities of about 400ky, 100ky, 40ky, and 20ky. The obliquity (~40ky) cycle was virtually absent in the tropical Newark Basin setting (within 10° of paleoequator; Kent et al., 1995; Olsen & Kent, 1995) but appears to more obviously present in the subtropical or temperate paleolatitudes of deposition (~25-35° N) in East Greenland. On the other hand, whereas the 400ky cycle is the most obvious in the Newark section (i.e., the McLaughlin cycle), the 100ky eccentricity cycle appears to be the most recognizable and mappable climate cycle within the Fleming Fjord Formation.

Finally, although the stratigraphic correlations and the pattern of climatic control of lithological variation for the Fleming Fjord Formation need to be confirmed, the present study already provides another example that rock sequences of continental, and especially lacustrine, facies can be successfully placed in a precise and testable chronostratigraphic framework using magnetostratigraphic and cyclostratigraphic analysis.

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Dansk sammendrag

Kontinentale sedimenter fra Trias er velblottede i Jameson Land Bassinet i Østgrønland. I sommeren 1992 blev der indsamlet sedimentprøver til palæomagnetisk analyse fra en 210m mægtig lakustrin lagserie i mellemste del af Fleming Fjord Formationen (Øvre Trias). Der blev også indsamlet sedimentprøver fra to niveauer med fluviale og associerede red-beds i den underliggende Gipsdalen Formation (Mellem Trias).

De palæomagnetiske analyser viser, at Jameson Land Bassinet var placeret på ca 18 grader N i Mellem Trias og ca 26 grader N i Sen Trias. Sammenligninger med data fra palæomagnetiske studier af Triassisk materiale fra Nordamerika og Europa viser imidlertid, at disse værdier er ca 10 grader for sydlige. En mulig forklaring er, at hæmatitmineralernes orientering i de østgrønlandske sedimenter er blevet påvirket ved kompaktion.

De palæomagnetiske data viser også, at den undersøgte del af Fleming Fjord Formationen kan inddeles i syv magnetozoner. Disse magnetozoner kan tilsyneladende korreleres med tilsvarende magnetozoner i den Sen Triassiske Newark Supergroup i Nordamerika. Korrelationen viser, at Fleming Fjord Formationen er af Norian alder, og at den undersøgte del af formationen blev aflejret på ca 3.5 millioner år.

De lakustrine sedimenter i Fleming Fjord Formationen er cyklisk aflejrede. I Malmros Klint Member, som på nuværende tidspunkt er bedst undersøgt, ses sedimentære cykler med mægtigheder på (25 m), 5.3–6.5 m, 2.1m og 1.3–1.6m Denne sammensatte cyklisitet tolkes som et resultat af astronomisk styret klimavariation i aflejringsbassinet. De sedimentære cykler kan derfor repræsentere følgende Triassiske orbitale svingninger: (413.000 år), 95–123.000 år, 36.000 år og 18–21.500 år. De overliggende lakustrine sedimenter i Ørsted Dal Member er ligeledes cyklisk aflejrede, og udfra cyklisiteten beregnes det, at den undersøgte del af Fleming Fjord Formationen blev aflejret på ca 4 millioner år.

De palæomagnetiske data viser, at Jameson Land Bassinet langsomt migrerede nordpå (ca 8 grader) under aflejringen af den Mellem-Øvre Triassiske lagserie. Denne ændring i palæobreddegrad afspejles i det kontinentale aflejringsbassins sedimentære udvikling fra aride facies i Gipsdalen Formationen til mere humide facies i Fleming Fjord Formationen.

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