

Magneto-biostratigraphy of the Middle to Upper Triassic transition, central Spitsbergen, arctic Norway

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Abstract: Palaeomagnetic and biostratigraphic data were obtained for the latest Ladinian and most of the Carnian (Botneheia Formation to basal Kapp Toscana Group), from two sections in central Spitsbergen (Svalbard archipelago). Thermal and alternating field (AF) demagnetization reveals a magnetization of both normal and reversed polarities. The mean directions pass reversal and fold tests and are similar to other European Late Triassic palaeopoles. One of the sections displays 15° of vertical-axis, clockwise, tectonic rotation on a décollement in the underlying Botneheia Formation. The magnetostratigraphy is dominated by normal polarity in the uppermost Botneheia Formation and into the Tschermakfjellet and basal De Geerdalen formations. A substantial hiatus characterizes the Ladinian–Carnian boundary in central Spitsbergen, so reverse magnetostratigraphic zones, identified in Tethyan sections near this boundary, are absent. Magnetostratigraphic correlation, along with palynostratigraphic constraints, indicates that most of the De Geerdalen Formation is Lower Carnian. The magnetostratigraphy and palynology indicates that the Isfjorden Member (upper unit of the De Geerdalen Formation) is probably mid-Carnian in age. Change in the lithological architecture in the Isfjorden Member, compared with the underlying parts of the De Geerdalen Fm, suggests a hiatus near the base of the member, which may represent a mid-Carnian unconformity, not previously recognized on Spitsbergen.

The transition from the Mid- to the Late Triassic was an interval of major progressive environmental change, the expression and causes of which are debated. During the earliest part of the Late Triassic (Carnian), turnover in terrestrial vertebrate groups led to the loss of many long-established groups, along with major diversification of many other groups (e.g. dinosaurs, crocodiles, mammals, etc.), which persisted to dominate for the remainder of the Mesozoic (Benton 1993; Lucas 1998). Extinction of many kinds of organisms in the marine realm also occurred during the Carnian (Simms *et al.* 1995; Lucas 1998). The part played by climatic change in these biotic turnovers is less than certain. However, it is clear that during the Carnian in central Europe major fluvial systems spread southwards from Baltica (German–Stuttgart Formation and equivalents; Beutler 1998). The timing of this may coincide with major progradation of coastal and fluvial systems, coincident with a major reorganization of sediment provenance, in western Boreal regions (Mørk *et al.* 1982, 1989, 1993; Mørk 1999; Egorov & Mørk 2000). In locations in the western Tethys, this may coincide with a major pulse of pollen and spores in the mid-Carnian, symptomatic of more humid and wet conditions (Fijalkowski-Mader 1998; Roghi 2004).

Understanding the timing and causes of these major environmental changes is in part hampered by poor ability to correlate in detail distant regions, and sediments deposited in marine and terrestrial environments. This study attempts to improve the situation by examination of the bio-magnetostratigraphy of the Middle and Upper Triassic transition represented by the Botneheia, Tschermakfjellet and De Geerdalen formations on Spitsbergen (the largest island in the Svalbard archipelago; Fig. 1). In addition, this study improves understanding of the age relation-

ships of the lowest part of the Upper Triassic units on Spitsbergen, better defines the nature of the Middle to Upper Triassic transition on Spitsbergen, and contributes to the debate on the definition of a Carnian magnetostratigraphy, tied to a biostratigraphic time scale. Currently, the Carnian has the least well-defined magnetic polarity time scale in the Triassic (Gallet *et al.* 1994; Muttoni *et al.* 2004).

Spitsbergen Middle and Upper Triassic and section details

The Triassic geology of Svalbard and adjacent parts of the Barents Sea continental shelf (Fig. 1) has a regional lithostratigraphy that is defined on the basis of outcrops of Triassic rocks on the Svalbard archipelago and hydrocarbon exploration wells in the surrounding seas (Mørk *et al.* 1999). In central and eastern Svalbard, the gently dipping Middle Triassic sequence is represented by the Botneheia Formation (Fm) of the Sassendalen Group, a unit of predominantly black, organic-rich marine shales (Fig. 2). This passes westwards into the Bravaisberget Fm in the fold belt of western Spitsbergen (Figs 1 and 2). In both east and west Spitsbergen, these Middle Triassic units pass up into the overlying Tschermakfjellet and De Geerdalen formations (of the Kapp Toscana Group; Fig. 2), which generally represent a west to east prograding delta-front succession. The Middle and parts of the Upper Triassic successions possess sporadic Anisian, Ladinian and Carnian ammonite faunas, which can be closely related to richer and more complete ammonite faunas in Siberia and British Columbia (Weitschat & Dagys 1989; Dagys & Weitschat 1993; Fig. 2). Also widely used in biostratigraphy in

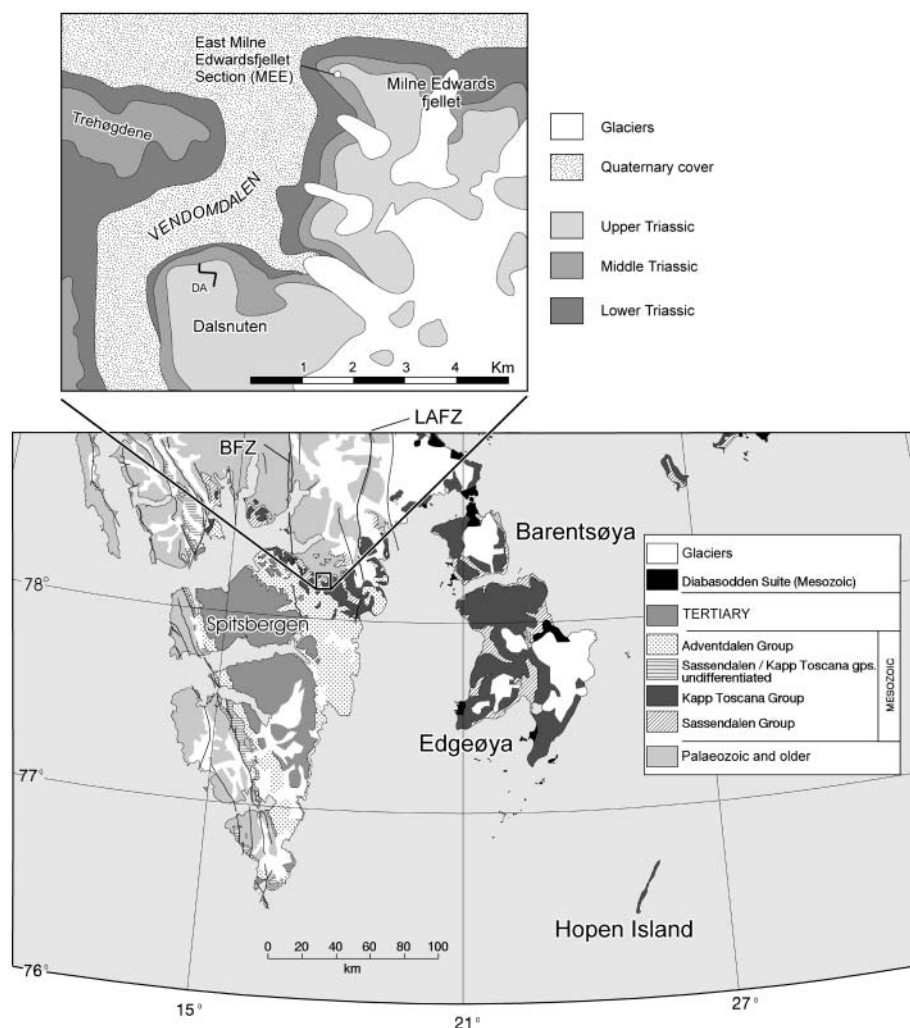


Fig. 1. Location map of Vendomdalen in central Spitsbergen, and a simplified geological map of the southern part of the Svalbard archipelago. Separate keys for main map and inset. Simplified after Dallmann (1999). BFZ, Billefjorden Fault Zone; LAFZ, Lomfjorden–Argardbukta Fault Zone. Inset shows the location of the Dalsnuten (DA) and Milne Edwardsfjellet (MEE) sections.

both Svalbard and the Barents Sea is a miospore zonation of the Lower, Middle and Upper Triassic units, based on reference outcrop and well data (Hochuli *et al.* 1989).

The bio-magnetostratigraphy of two sections in Vendomdalen, a short section (referred to as MEE, 78°13'4"N, 17°30'4"E) at Milne Edwardsfjellet, and the longer Dalsnuten section (referred to as DA, 78°11'52"N, 17°23'46"E), 5 km to the south on the southwestern side of Vendomdalen, are documented (Figs 1 and 3). These sections are *c.* 42 km east of Longyearbyen, the capital of Spitsbergen. The region around Vendomdalen is within the NE part of the Central Tertiary Basin of southern Spitsbergen and is structurally bounded to the west by the Billefjorden Fault Zone, and to the east by the Lomfjorden–Argardbukta Fault Zone, a structural unit referred to as the Ny Friesland Block (Fig. 1). This block shows evidence of inversion during the Early Tertiary, synchronous with folding in west Spitsbergen (Haremo & Andresen 1992). The Triassic within the Ny Friesland Block has a gentle regional dip to the south to SW.

The MEE section consists of the uppermost 18 m the Botneheia Fm and the lowest 2 m of the Tschermakfjellet Fm, exposed at a small plateau forming the northwesternmost point of Milne Edwardsfjellet (Fig. 1). The succession is dominated by dark grey shales that weather to a bluish colour (Fig. 3). Abundant thin (5–20 cm thick) dolomite and limestone beds, occasionally silty, are laterally persistent to impersistent. Horizons with lime-

stone nodules (up to 30 cm diameter) occur at 4.5 m and between 9 and 10 m above the base of the section (Fig. 3). A level with larger nodules occurs at 13 m and small phosphate nodules are abundant in the lower 9 m. A cliff with silty, calcite-cemented shale, which is overlain by soft grey shale, marks the top of the Botneheia Fm. Fragments of sideritic and phosphatic nodules are abundant in the debris on top of this cliff. The complete Botneheia Fm is exposed in Milne Edwardsfjellet, but the upper part of the formation is locally disturbed by a regional décollement (Andresen *et al.* 1992). The MEE section lies above this décollement, and hence is structurally separated from the lower 120 m of the Botneheia Fm.

The Dalsnuten section begins with 40 m of the Tschermakfjellet Fm resting on the Botneheia Fm in the lower part of the mountain side (Fig. 4). The section begins *c.* 1–2 m above the Botneheia Fm. The overlying *c.* 300 m of the De Geerdalen Fm has an upper division referred to the Isfjorden Member (Mbr) following the terminology of Pchelina (1980) and Mørk *et al.* (1999). The section is believed to be one of the most complete and best-exposed sections through the Upper Triassic succession in central Spitsbergen. The sedimentological profile at Dalsnuten was initially measured by Knarud (1980), and was later defined as the parastratotype for the De Geerdalen Fm by Mørk *et al.* (1982).

Most of the Botneheia Fm in the lower part of the Dalsnuten

Age			Macrofossil zones ¹			Palynology				Lithostratigraphy				Group
			British Columbia	Siberia	Svalbard	Svalbard & Barents Shelf				West	Svalbard	East	Bjørnøya	
Upper Triassic	Norian	Middle	Mesohimavatites columbianus	Otapiria ussuriensis		2	3	4	5	WILHELMØYA SUBGROUP				KAPP TOSCANA
		Drepanites rutherfordi												
		Juvavites magnus												
	Carnian	Early	Malayites dawsoni	Pterosirenites obrucevi	Pterosirenites beds	B								
		Stinkinoceras kerri												
		Klamathites macrolobatus	Sirenites yakutensis		C									
		Tropites welleri	Neosirenites pentastichus		D									
Early	Tropites dilleri			E										
	Sirenites nanseni	"Neoprotrachyceras" seimkanense		F										
	Austrotrachyceras obesum	"Protrachyceras" omkutchanicum		F	F									
Middle Triassic	Ladinian		Trachyceras desatoyense	Stolleyites tenuis	Stolleyites tenuis					TSCHERMAKFJELLET Fm				
		Frankites sutherlandi	Nathorstites lindstroemi	Daxatina canadensis	G	G							Skuld Fm	
			Nathorstites mcconnelli											
		Mclearnoceras mclearni	Nathorstites mclearni											
		Meginoceras meginiae	Indigirites krugi	Indigirites tozeri	H									
			Tsvetkovites neraensis											
			Tsvetkovites constantis	Tsvetkovites varius	I									
	Tuchodoceras poseidon													
		Early	Eoprotrachyceras matutinum	Intomites oleshkoi										

1: From Dagys & Weitschat (1993), Orchard & Tozer (1997)

2: From Hochuli et al. (1989), Svalbard & Barents Sea

3: From Mørk et al. (1990): Bjørnøya

4: From Vigran et al. (1998): Svalis Dome

5: This study: central Spitsbergen

Solid lithostratigraphic boundaries are biostratigraphically dated

Broken lithostratigraphic boundaries are inferred

Fig. 2. Outline lithostratigraphy and biostratigraphy of Svalbard, the adjacent Barents Sea, and other important Boreal ammonoid biostratigraphies.

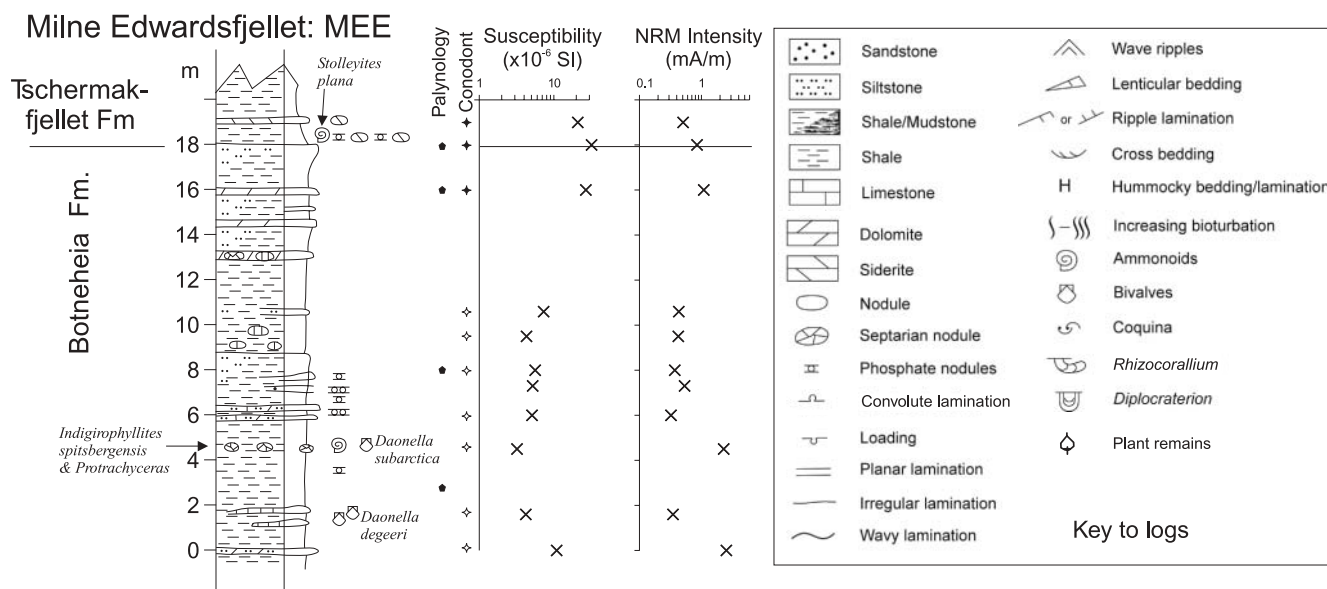


Fig. 3. Milne Edwardsfjellet (MEE) section. Magnetic susceptibility and natural remanent magnetization intensity (NRM) for the palaeomagnetic specimens, with the palynology (shown in Fig. 6) and conodont sampling locations (open symbols are barren samples).

mountain side is structurally complicated by duplex thrusting associated with the décollement, which in part imbricates the formation. This does not appear to cause disruption of the overlying Upper Triassic units, other than for one observed small monoclinical fold in the sampled section. The décollement formed during the Palaeocene or Eocene, synchronous with more severe folding in western Spitsbergen (Haremo & Andresen 1992).

The Tscherma-fjellet Fm consists of grey shales with thin, often tabular, siderite concretions and in the upper part thin, cross-laminated, sandstone beds. The unit coarsens upwards and grades into sandstones of the De Geerdalen Fm (Fig. 5).

The De Geerdalen Fm forms the upper part of Dalsnuten, where the unit consists of repeated coarsening upward sandstone bodies separated by shales (Figs 4 and 5). The sandstone bodies

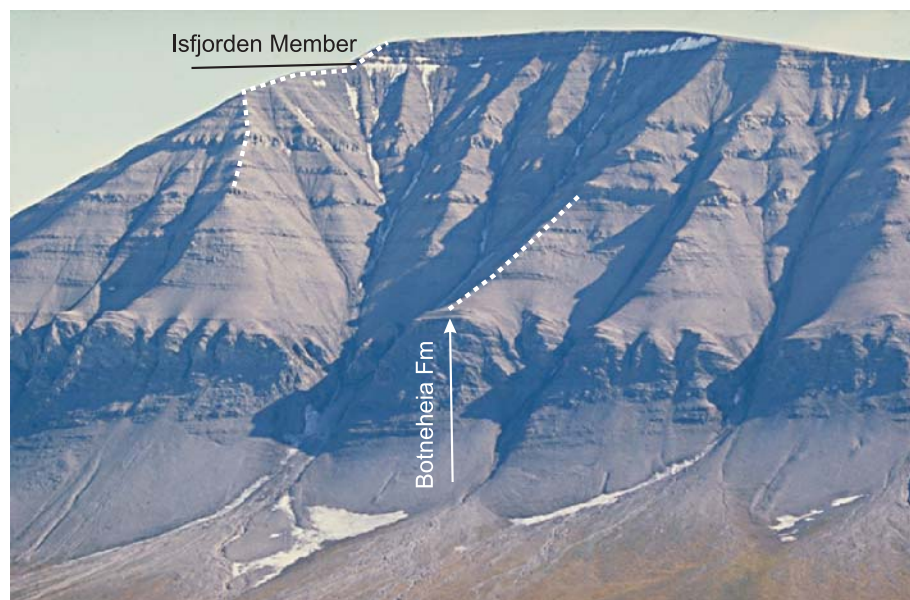


Fig. 4. The Dalsnuten section seen from the NE side of Vendomdalen. The cliff-like rib is the top of the Botneheia Fm; the Tschermaffjellet Fm (directly above the cliff) and the De Geerdalen Fm form most of the mountain side. The uppermost ledge is the Isfjorden Member. Sampled section shown as dotted line.

vary in thickness from a few metres to 30 m and represent development of deltaic and marginal marine sandstone bodies (Mørk *et al.* 1982). The sandstone bodies show great thickness variations laterally, over a distance of several hundred metres, as a result of development of the channel sandstones. Large-scale cross-bedding, hummocky cross-bedding and clay clasts are abundant in the sandstones, along with horizons rich in carbonaceous plant fragments (Mørk *et al.* 1982). Macrofossils were not seen (except in the overlying Isfjorden Mbr) but a few trace fossils (*Rhizocorallium*, *Diplocraterion* and a few unclassified tunnels) are present. Further east on Spitsbergen the deltaic influence increases and the major sandstone deposits are fluvial, whereas to the south and west a marine environment dominates.

A siderite- and calcite-rich siltstone bed crowded with bivalves and sideritic nodules occurs at level 521 m, forming the base of the Isfjorden Mbr. The beds immediately underlying this form a regressive coarsening upward unit containing *Diplocraterion* and hence, the sideritic, bivalve-bearing bed represents a transgressive base to the Isfjorden Mbr. The overlying part of the Isfjorden Mbr consists of alternating shales and evenly bedded, thin- to thick-bedded siltstones and sandstones, which display consistent thicknesses along the mountainside, as a result of the absence of major channel sandstones (Fig. 4). Above the section on the mountain plateau are fragments of siderite nodules that often contain imprints of bivalves. By comparison with other sections this horizon may represent either the base of the overlying Wilhelmøya Subgroup or a bed high in the Isfjorden Mbr.

Samples and methods

Hand samples from 72 horizons were collected for magnetostratigraphy; 10 are from the MEE section and 62 from the Dalsnuten (DA) section. In the MEE section, samples were collected from the limestones and the better-cemented shales. In the DA section the majority of the samples were collected from fine-grained, often laminated or cross-laminated sandstones when available. In some intervals sandstones were poorly cemented and too friable to be sampled. A large sampling gap in the middle part of the section is due to extensive scree cover below the thick sandstones. Within the lower part of the Tschermaffjellet Fm the samples

are dominantly from siderite concretions; the intervening shales were too friable for magnetostratigraphic sample collection. All hand samples were oriented with a special orientation staff and a magnetic compass.

In the DA section, a fold test was attempted using a localized monocline in the boundary beds of the Tschermaffjellet and De Geerdalen formations. This fold utilized samples from 21 horizons, five of which had bedding dips to the NE of 35–63°, eight of which had bedding dips to the SW of 12–19°, and the remainder had bedding dips of <5° to the SW. The deformation producing this folding was probably initiated by a splay off the duplex faulting, which locally imbricates the Botneheia Fm in the lower part of the mountain.

In the laboratory, each sample was either cut or drilled into three to five standard specimens. A total of 164 specimens were subjected to demagnetization. Both thermal demagnetization and alternating field (AF) demagnetization were utilized, using a Magnetic Measurements Ltd thermal demagnetizer and a Molspin AF demagnetizer. The specimens were measured on a CCL GM400 three-axis cryogenic magnetometer (noise level *c.* 0.002 mA m⁻¹) that makes six measurements of each specimen axis, from which the magnetization variance was calculated. Low-field magnetic susceptibility of each specimen was measured with a Bartington Ltd MS2 susceptibility meter, after each thermal treatment step. Characteristic remanent magnetization (ChRM) directions were isolated using principle component analysis as implemented in LINE-FIND (Kent *et al.* 1983). Use has been made of both linear trajectory fits and great circle data in defining the palaeomagnetic behaviour. LINE-FIND contains a statistical test to evaluate the significance of planarity in fitting great circles to the demagnetization data, and displays those that are significant (Kent *et al.* 1983). Above thermal demagnetization temperatures of 200–350 °C most samples displayed major mineralogical changes (the 'susceptibility crisis') manifested by large increases in susceptibility and acquisition of spurious viscous magnetizations. Consequently, the demagnetization strategy adopted was to thermally demagnetize to 200–350 °C (dictated by lithology), prior to the susceptibility crisis, followed by AF demagnetization. This successfully isolated the ChRM directions in most specimens. In some 5% of specimens, demagnetization failed to produce any consistent magnetization direction. In addition, in some specimens AF demagnetization alone was often as successful in isolating the ChRM. One to four specimens were measured from each hand sample.

Progressive isothermal remanent magnetization (IRM) up to 1 T was applied to a representative subset of specimens, to investigate the magnetic mineralogy, which was also studied by demagnetizing a three-component orthogonal IRM (Lowrie 1990). Magnetic fields of 1, 0.4 and 0.03 T were applied along the *x*-, *y*- and *z*-axis respectively of the

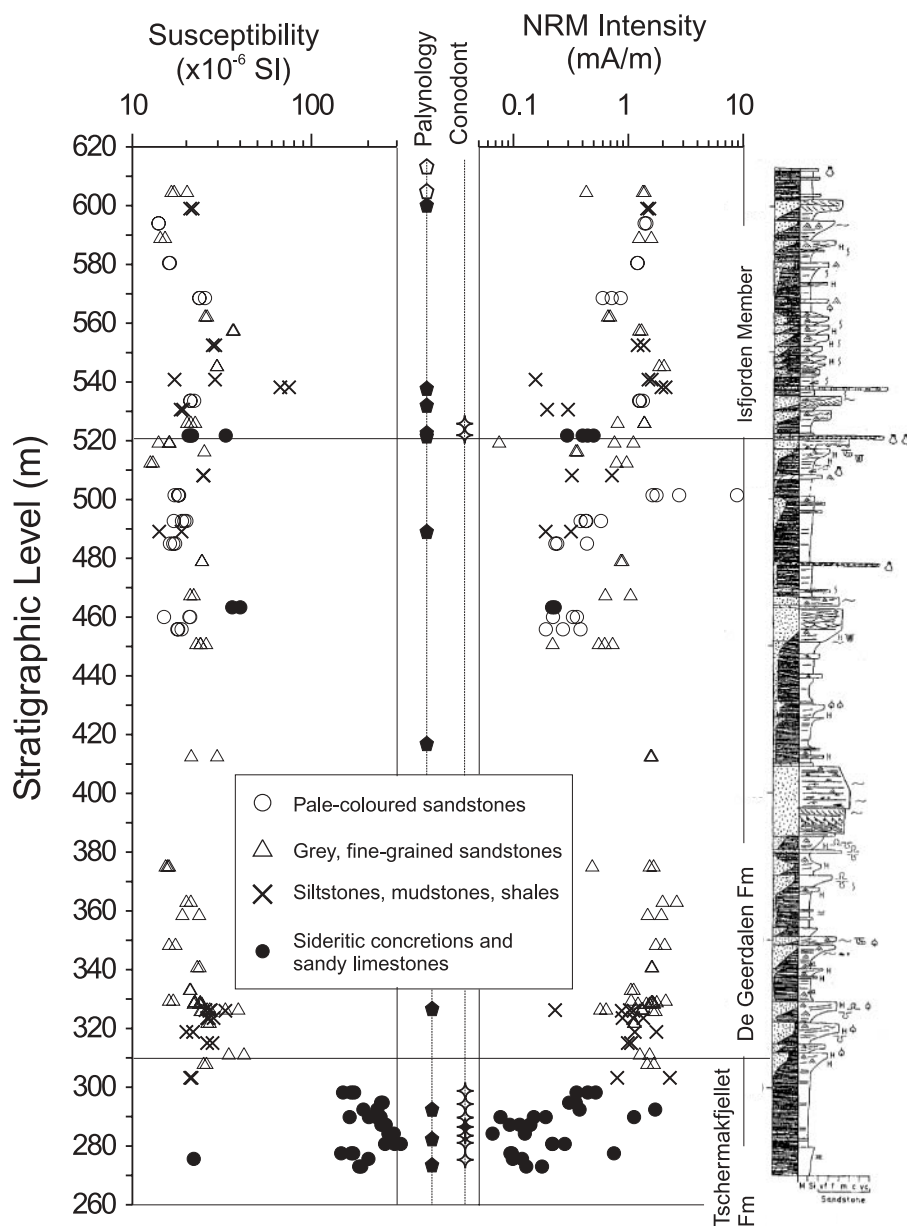


Fig. 5. Magnetic susceptibility and natural remanent magnetization intensity (NRM) for the specimens, with the palynology (shown in Fig. 6) and conodont sample levels (barren palynology and conodont levels shown with open symbol). On the right is the log of the Dalsnuten section through the De Geerdalen Formation and the Tschermakfjellet Formation (from Knarud 1980). Key to log is shown in Figure 3.

specimens. A Molspin Pulse Magnetizer (up to 0.3 T) and a Newport DC electromagnet were used. The IRM was measured with a Molspin Ltd spinner magnetometer. Some additional low-temperature remanence measurements (to 10 K) were also undertaken to identify the magnetic mineralogy, using a Princeton Measurements Ltd Magnetic Property Measurements System.

Samples for palynology were collected in parallel with the magnetostratigraphy. These were preferentially collected from shale units interbedded in the sequence, and processed using standard techniques in the SINTEF laboratory. A subset of samples, at the same horizons as the palaeomagnetic samples, were processed for conodonts at the Geological Survey of Canada (Vancouver).

Ammonoid biostratigraphy

Ammonoids in the Middle and Upper Triassic units of Svalbard are sporadic and restricted to a few horizons, with the ammonoid assemblages of the Upper Ladinian providing a number of problems in their correlation to the more complete Siberian and Canadian ammonoid zones (Weitschat & Lehmann 1983; Dagys

et al. 1993). An upper 'Ptychitid layer' occurs on Spitsbergen within 2–13 m of the top of the Botneheia Fm. This level contains an ammonite fauna (*Aristoptychites kolymensis*, *Indigirites tozeri* and others, along with the bivalve *Daonella degeeri*; Weitschat & Lehmann 1983; Weitschat & Dagys 1989), which characterizes the *Indigirites tozeri* Zone throughout the Svalbard archipelago (Fig. 2). Dagys & Weitschat (1993) and Dagys & Konstantinov (1997) correlated the Svalbard *I. tozeri* Zone with the Siberian *I. krugi* Zone (Fig. 2), which is synchronous with the mid-Tethyan ammonoid *Protrachyceras* Zone. The next occurrence of ammonoids above this level in central Spitsbergen is the nathorstiid genus *Stolleyites*, within the Tschermakfjellet Fm, which is divisible into the *S. planus* and overlying *S. tenuis* Subzones (Dagys & Weitschat 1993). *Stolleyites* is also known from the *Trachyceras desatoyense* Zone of British Columbia, where it occurs with the conodonts *Neogondolella* (*Paragondolella*) *inclinata* and *Mosherella* sp. (Dagys & Weitschat 1993; Orchard & Tozer 1997). These provide evidence of the approximate equivalence of the Siberian–Svalbard *Stolleyites* Zone and

the ammonoid *Trachyceras aon* Subzone of the southern Alps (Broglia Loriga *et al.* 1999).

In the MEE section the Ptychitid layer occurs 13 m below the top of the Botneheia Fm, where *Indigirophyllites spitsbergensis* occurs with *Protrachyceras* and *Daonella subarctica* (Fig. 3). In the basal Tschermakfjellet Fm, *Stolleyites planus* (the index species of the first subzone of the *Stolleyites tenuis* Zone) is found, within siderite nodules. At Dalsnuten no ammonoids have been found in the siderite nodules of the Tschermakfjellet Fm.

In eastern Spitsbergen, east of the Lomfjorden–Arghardbukta Fault Zone (Fig. 1), the distance between the Ptychitid layer and the top of the Botneheia Fm increases, with two additional ammonite-bearing horizons. The lower one (some 3.5 m below the top of the formation) contains *Arctoptychites?* *Protrachyceras?* and *Daonella ex gr. subarctica*. The upper level (some 2 m below the top of the Botneheia Fm) contains *Protrachyceras* sp. and *Daonella?* *subarctica*. The relationship of these *Arctoptychites*-bearing beds to other Boreal zonal schemes is uncertain; they may relate to the *Meginoceras meginiae* or *Mclearnoceras mclearnii* ammonoid zones of British Columbia (Weitschat & Dagys 1989).

It is only on Bjørnøya (300 km south of Longyearbyen) that the uppermost Svalbard Ladinian ammonoid zone with *Daxatina canadensis* is known. On Bjørnøya this occurs 40 m from the top of the 140 m thick Skuld Fm (Mørk *et al.* 1989, 1990, 1999), which consists of grey shales, with siderite nodules and some sandstone beds, similar to the lithology of the Spitsbergen Tschermakfjellet Fm, implying southward diachroneity of this 'Tschermakfjellet' facies (Weitschat & Dagys 1989).

Hence, in central Spitsbergen ammonoid zones above the *Indigirites tozeri* Zone appear to be absent, but there is some evidence of an eastwards and southwards transition into younger Ladinian strata, which are not seen in central Spitsbergen. It is not clear if this is due to lack of preservation of the *Daxatina canadensis* fauna, or to a hiatus at the top of the Botneheia Fm in central Spitsbergen. The phosphate nodule-rich shale that characterizes the top of the Botneheia Fm in Spitsbergen could be indicative of condensation or hiatus. The Bjørnøya *Daxatina canadensis* Zone provides a direct link to the proposed Carnian Global Stratotype Section and Point at Stuores (Italian Dolomites), because the base of this biozone has been proposed as the base of the Carnian (Broglia Loriga *et al.* 1999).

Norian ammonoids are not known from Spitsbergen, being restricted to Hopen Island (Fig. 1), where a *Pterosirenites* and *Pteroceras* ammonoid fauna (Korchinskaya 1982) is known from the base of the Wilhelmøya Subgroup (Fig. 2).

Palynology

Palynological assemblages from the Botneheia Fm in the MEE section (Figs 3 and 6) include *Angustisulcites klausii*, *Ovalipollis pseudoalatus*, *Podosporites amicus*, *Protodiploxypinus minor*, *P. ornatus*, *Striatoabieites balmei*, *S. multistriatus*, *Triadispora verrucata* and *Voltziaceasporites heteromorpha*. In the Botneheia Fm below the décollement *Protodiploxypinus decus* and *Retisulcites perforatus* also occur. Stratigraphically all these palynomorphs distinguish the oldest Ladinian, Assemblage I (Fig. 2), of Hochuli *et al.* (1989) from the underlying Anisian in both the Botneheia and Bravaisberget formations on Svalbard. Hochuli *et al.* (1989) distinguished two younger Ladinian assemblages, H and G, based on assemblage characteristics (and the presence of an acritarch) from commercial wells drilled closer to the Norwegian mainland. These relative characteristics have been traced (but not consistently) in assemblages from the

upper part of the Botneheia and Bravaisberget formations on Svalbard, but it is not possible to confidently recognize Assemblages H and G from first or last stratigraphic appearances of miospores.

The assemblages from the Tschermakfjellet Fm of the MEE section contain abundant variants of the *Protodiploxypinus* and *Triadispora* groups. The lowest appearance of *Chasmatosporites apertus* (18 m in MEE; Fig. 6) distinguishes Assemblage F from the stratigraphically older Assemblages I–G of Hochuli *et al.* (1989). Other significant first occurrences within the Tschermakfjellet Fm are *Iraquispora (Kyrtomisporis) laevigata*, *Leschikisporis aduncus*, *Chasmatosporites major*, *Gleichenioidites senonicus* and *Aulisporites astigosus*, which, together with the continued presence of *Cordaitina minor*, *Illinites chitonoides*, *Ovalipollis pseudoalatus*, *Protodiploxypinus decus*, *Schizaeosporites worsleyi*, *Triadispora verrucata* and *Voltziaceasporites heteromorpha*, allow correlation with Assemblage F of Hochuli *et al.* (1989). Similar assemblages have been recorded in the Tschermakfjellet Fm and the lower part of the De Geerdalen Fm at Festningen (in western Spitsbergen) and on the islands east of Spitsbergen (Fig. 1). Assemblage F has been recorded only in deposits younger than the *Indigirites tozeri* ammonoid zone on Svalbard. Abundant *Micrhystridium* and other marine plankton up to the 280.7 m level at Dalsnuten confirm the marine depositional environment.

In the topmost Tschermakfjellet Fm and lower part of the De Geerdalen Fm at Dalsnuten (sample levels 292.4–489.1 m; Fig. 6) the assemblages contain abundant smooth trilete spores. The lower content of marine plankton and the presence of *Cingulizonates rhaeticus*, *Polypodiisporites ipsviciensis*, *Zebrasporites interscriptus* and *Thomsonisporites undulatus* distinguish these levels from those below. The low abundance of *O. pseudoalatus* in the sample interval 326.2–489.1 m is a feature that was used by Hochuli *et al.* (1989) to differentiate Assemblage E from the underlying Assemblage F. However, in spite of this, the miospores are more suggestive of Assemblage F. The assemblage probably derived from marshland vegetation with local coal formation and reduced marine influence.

At the 520.5 m sample level the presence of diverse plankton (abundant *Micrhystridium*) distinguishes it from the interval below (Fig. 6). However, the continued presence of *Illinites chitonoides* and the re-occurrence of *Kuglerina meieri* suggest correlation with Assemblage F of Hochuli *et al.* (1989). The abundance of *Chasmatosporites major* and the increased abundance of *O. pseudoalatus* in this sample suggest that a rising water level may have drowned parts of the coastal plain and forced a change of the previous vegetation. The reappearance of abundant *O. pseudoalatus* otherwise corresponds to features used by Hochuli *et al.* (1989) to distinguish Assemblage D from Assemblage E.

At 521.5 m, at the base of the Isfjorden Mbr, the rich and diverse plankton association also includes abundant freshwater algae; *Plaesiodyctyon mosellanum*, *Psophosphaera* and *Botryococcus*, suggesting influx from a freshwater body into a shallow marine area. The pollen and spores are more diverse than the association at the 520.5 m level, but both levels contain *Angustisulcites klausii*, *I. chitonoides*, *L. aduncus*, *Polypodiisporites ipsviciensis*, *Staurosaccites quadrifidus*, *S. worsleyi*, *Sellaspora foveorugulata* and *S. rugoverrucata* (Fig. 6). According to Hochuli *et al.* (1989), the continued presence of *I. chitonoides* (if not reworked), would indicate Assemblage F; hence, this assemblage appears to be present throughout the Tschermakfjellet Fm and lower parts of the De Geerdalen Fm, into the Isfjorden Mbr.

D Dominant ○ Abundant ○ Common	Key X Present # Present below décollement	← MEE → : → DA →																	
		Botneheia Fm		Tschermakfjellet Fm					DeGeerdalen Fm										
		3.2	8.0	16.0	18.0	273.0	280.7	292.4	326.2	417.0	489.1	520.5	521.5	530.6	538.2	599.0			
Selected miospore taxa / Height (m) in section																			
Bisaccate indeterminate pollen	#	D	D	D	D			D	x	x	x	x	O	O	x	x	x	x	
<i>Deltoidospora australis</i> (Couper 1953) Pocock 1970	#	x	x					x	o	O	O	x	x	x	x	x	x	x	
<i>Deltoidospora minor</i> (Couper 1953) Pocock 1970	#	x	x					x	O	D		x	o	x	o	o	O	x	x
<i>Schizaeosporites worsleyi</i> Bjaerke & Manum 1977	#				x							x		x	x			x	
<i>Annulispora folliculosa</i> (Rogalska 1954) de Jersey 1959	#	x			x								x					x	
<i>Illinites chitonoides</i> Klaus 1964	#	x						x	x			x	x	x	x	x	x		
<i>Ovalipollis pseudoalatus</i> (Thiergart 1949) Schuurmann 1976	#	x	x		o	o		O	o	o		x		o	D	x			
<i>Protodiploxypinus ornatus</i> (Pautsch 1973) Bjaerke & Manum 1977	#	x			x							x			x	o			
<i>Camarozonosporites rudis</i> (Leschik 1955) Klaus 1960	#								x			x	x	x		x	O		
<i>Accinctisporites circumdatus</i> (Leschik in Krausel & Lescchik 1955) Jain 1968	#	x														x			
<i>Angustisulcites klausii</i> Freudenthal 1964	#	x	x								x					x			
<i>Araucariacites australis</i> Cookson 1947	#										x			O		x			
<i>Camarozonosporites laevigatus</i> Schulz 1967	#											x			x	x			
<i>Cordaitina minor</i> Pautsch (1971) Pautsch 1973	#										x			x		x			
<i>Lunatisporites noviaulensis</i> (Leschik 1956) Scheuring 1970	#	x	x		x			x	x						x	x			
<i>Podosporites amicus</i> Scheuring 1970	#	x	x		x	x		x	x							x			
<i>Protodiploxypinus macroverrucosus</i> Bjaerke & Manum 1977	#				x				x	x						x	o		
<i>Protodiploxypinus minor</i> Bjaerke & Manum 1977	#	x			x											x	o		
<i>Protodiploxypinus</i> spp.	#	x			x			x	x	O		x	x	x	x	o			
<i>Staurosaccites quadrifidus</i> Dolby in Dolby & Balme 1976	#	x			x					x					x	x			
<i>Striatoabieites multistriatus</i> (Balme & Hennelly 1955) Hart 1964	#	o	x		x	o									x	x			
<i>Triadispora verrucata</i> (Schulz 1966) Scheuring 1970	#	x	x		x				o	x						x			
<i>Voltziaesporites heteromorpha</i> Klaus 1964	#	x	x		o			x	x	x		x	x	o		O			
<i>Kuglerina meieri</i> Scheuring 1978	#														x				
<i>Protodiploxypinus decus</i>	#								x	x									
<i>Duplicisporites granulatus</i> (Leschik 1955) Scheuring 1970	#				x														
<i>Eucommiidites major</i> Schulz 1967	#				x	x													
<i>Striatoabieites balmei</i> (Klaus 1964) Scheuring 1978	#	x	x																
<i>Staplinisporites caminus</i> (Balme 1957) Pocock 1962	#				x	o													
<i>Chasmatosporites apertus</i> (Rogalska 1954) Nilsson 1958	#				x				x	x									
<i>Iraquispora laevigata</i> (Mädler 1964) Lund 1977	#								x	x	x					x	x		
<i>Leschikisporis aduncus</i> (Leschik 1955) Potonié 1958	#								x	O	O	x	o	x	o	x	D		
<i>Chasmatosporites major</i> (Nilsson 1958) Pocock & Jansonius 1969	#								x				x	O	x	D			
<i>Dictyophyllidites mortonii</i> (deJersey) Playford & Dettmann 1965	#								O	x		x	o	o	o	O	O	x	D
<i>Gleicheniidites senonicus</i> Ross 1949	#								x	x			x				x		
<i>Aulisporites astigmaticus</i> (Leschik 1955) Klaus 1960	#								D	x		x	x	O					
<i>Polypodiisporites ipsviensis</i> (de Jersey 1962) Playford & Dettmann 1965	#								o			x	x	x			o	O	x
<i>Cingulizonates rhaeticus</i> (Reinhardt 1962) Schulz 1967	#								x						x			x	
<i>Kraeuselisporites dentatus</i> , <i>K. cooksonae</i> , <i>K. spinosus</i>	#								x			x	x	x				x	
<i>Uvaeisporites argenteaformis</i> (Bolkhovitina 1953) Schulz 1967	#								x	x		x					x		
<i>Eucommiidites microgranulatus</i> Scheuring 1970	#								x							O			
<i>Protodiploxypinus gracilis</i> Scheuring 1970	#								O			x	x		x				
<i>Concentrisporites hallei</i> (Nilsson 1958) Wall 1965	#											x	x		o	x			
<i>Zebrasporites interscriptus</i> Klaus 1960	#											x					x		
<i>Thomsonisporites undulatus</i> Leschik 1955	#											x							
<i>Infernopollenites</i> spp.	#														x	x			
<i>Retisulcites perforatus</i> (Mädler 1964) Scheuring 1970	#														x	x			
<i>Sellaspora foveorugulata</i> van der Eem 1983	#														x	x			
<i>Sellaspora rugoverrucata</i> van der Eem 1983	#														x	x			
<i>Vallatisporites</i> spp.	#															x			
Plankton taxa																			
<i>Micrhystridium</i> spp.	#	O	O		x			O	D						O	O			x
Tasmanitids	#	O	O		O	o													
<i>Veryhachium</i> spp.	#				x	x		O	x		x	x			x	x			
<i>Cymatiosphaera</i> sp.	#							o							x	D			
<i>Dictyotidium</i> spp.	#									x	x	x			x	O			
<i>Tasmanites</i> spp.	#									x					x				
<i>Botryococcus</i> spp.	#													x	x	O	x	x	x
<i>Plaesiodyctyon moesellianum</i> Wille 1972	#															D	x		
<i>Psophosphaera</i> spp.	#														x	x	o		
Equivalent assemblages of Hochuli et al. (1989)																			

Fig. 6. Summary palynology of the Milne Edwardsfjellet (MEE) and Dalsnuten (DA) sections. (See Figs 3 and 5 for location of samples.) Miospores in bold are those of most stratigraphic significance. Unpublished data from the Botneheia Formation below the décollement at Milne Edwardsfjellet are also shown.

A number of taxa sporadically present through the Botneheia and Tschermakfjellet formations are apparently absent in the Isfjorden Mbr at the 530.6 m level (Fig 6). However, these same taxa are also intermittently present in the lower parts of the De

Geerdalen Fm, and their absence at this level can probably be ascribed to local environmental control of the parent flora.

At the highest non-barren sample level (599.0 m; Figs 5 and 6) there are no age-restricting palynomorphs. The low-diversity

assemblage is dominated by *Dictyophyllidites mortoni* and other smooth spores (from ferns). The abundance of ferns suggests a humid environment, or a change to a more humid climate. Small *Botryococcus* colonies (4–8 cells), fungal hyphae and spores are also dominant at this level, and may represent remains of lichens from the modern soil.

Conodonts

Of the 10 samples processed from the MEE section, only three samples between the 16 and 19 m levels yielded conodonts (Fig. 3). These are predominantly *Neogondolella* specimens assignable to *N. inclinata*. This species generally ranges through the Ladinian–Carnian boundary (under most definitions) but is generally dominant in the Late Ladinian (Orchard & Tozer 1997). The topmost sample at 19 m (in the Tschermaffjell Fm; Fig. 3) contains very early forms of *Metapolygnathus* ex gr. *polygnathiformis*, which appears close to or at the base of the Carnian (depending on its final definition; Orchard & Tozer 1997; Balini *et al.* 2001). Nine samples from the Dalsnuten section were processed; only one of these (at 297 m) provided a single (non-useful) conodont element. This sample and three others provided a variety of foraminifers, ostracodes and ichthyoliths.

Magnetic mineralogy

For the DA section, magnetic susceptibility and natural remanent magnetization (NRM) display some relationship to lithology (Fig. 5). Siderite-bearing lithologies show most deviation from the norm, with high magnetic susceptibility, as a result of the FeCO_3 , and often low NRM intensity, caused by siderite diluting the clastic fraction, especially in the concretions (Fig. 5). In the clastic lithologies, siltstones have marginally larger magnetic

susceptibilities and lower NRM intensities than sandstones. The largest effect on the NRM intensities appears to be related to stratigraphic range, with values generally $<1 \text{ mA m}^{-1}$ in the 60 m below the Isfjorden Mbr, and $>1 \text{ mA m}^{-1}$ in the Isfjorden Mbr and base of the De Geerdalen Fm (Fig. 5). For the MEE section, NRM intensity was between 0.3 and 2 mA m^{-1} , and displayed some stratigraphic variation (Fig. 3).

For the DA section, the IRM on selected specimens mostly shows near saturation at about 200–300 mT (Fig. 7a), with only one specimen (DA2, 599 m) showing significant magnetization acquisition after 300 mT. These data are compatible with the ability to AF demagnetize the NRM, indicating a low-coercivity mineral such as magnetite being responsible for most of the natural remanence. The presence of magnetite is confirmed by the low-temperature measurements, which display the Verwey transition at 120 K, on both warming and cooling of the applied IRM (Fig. 8). If the magnetite was of a restricted particle size, the relatively moderate to small loss in IRM of DA57 and DA27 at *c.* 120 K could be interpreted as the result of either a population of small ($<0.2 \mu\text{m}$) magnetite particles (King & Williams 2000) or larger magnetite particles that are in part substantially oxidized (Özdemir *et al.* 1993; Smirnov & Tarduno 2002). There is no transition at 34 K, indicating the absence of pyrrhotite. The rapid loss of remanence at 18 K in DA27 on warming is attributed to the siderite in this sample (Fig. 8a). The substantial difference of this value from the Neel temperature of 38 K for pure siderite (Frederichs *et al.* 2003) probably indicates substantial cation (Ca?) substitution in the siderite.

Demagnetization of the three-component IRM for the DA samples shows that the specimens are characterized by distributed unblocking temperatures, up to 550–600 °C (Fig. 7c and d). This is probably indicative of a wide range of magnetite grain sizes. Significantly, there is no rapid decay of intensity at temperatures below 300 °C, like that displayed by the NRM (see

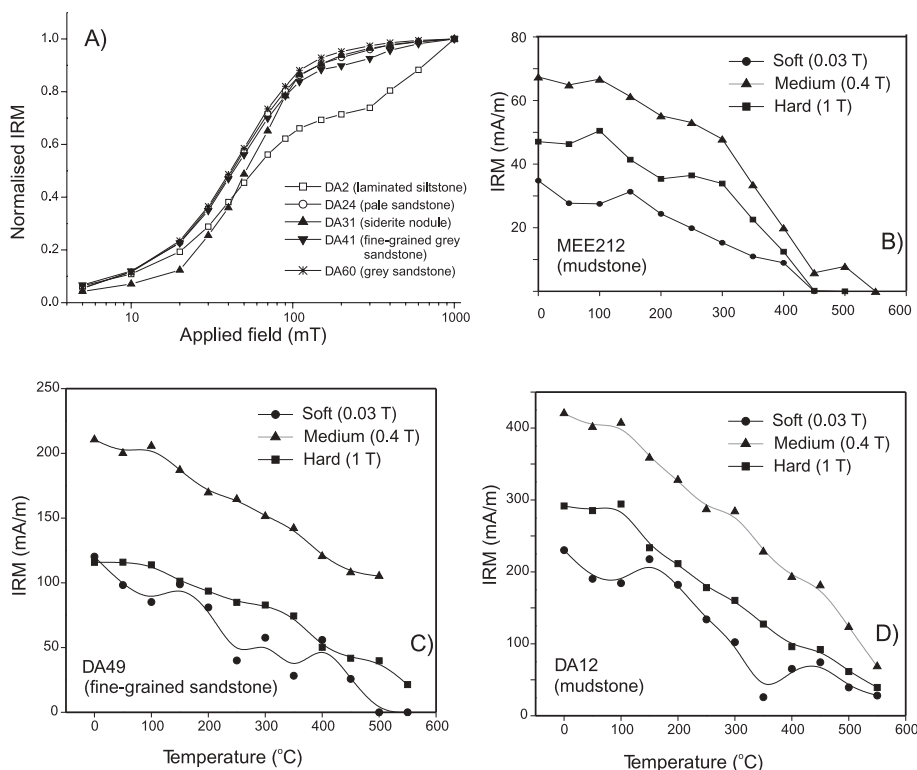


Fig. 7. (a) Isothermal remanent magnetizations for five representative specimens from Dalsnuten. (b–d) Thermal demagnetization of a three-component IRM for samples MEE212 (b), DA49 (c) and DA12 (d). Specimen stratigraphic location is shown in Figures 11 and 12. Some data points $>500^\circ\text{C}$ in (c) and (d) have been removed because of large acquisition of spurious viscous remanence caused by mineralogical alteration.

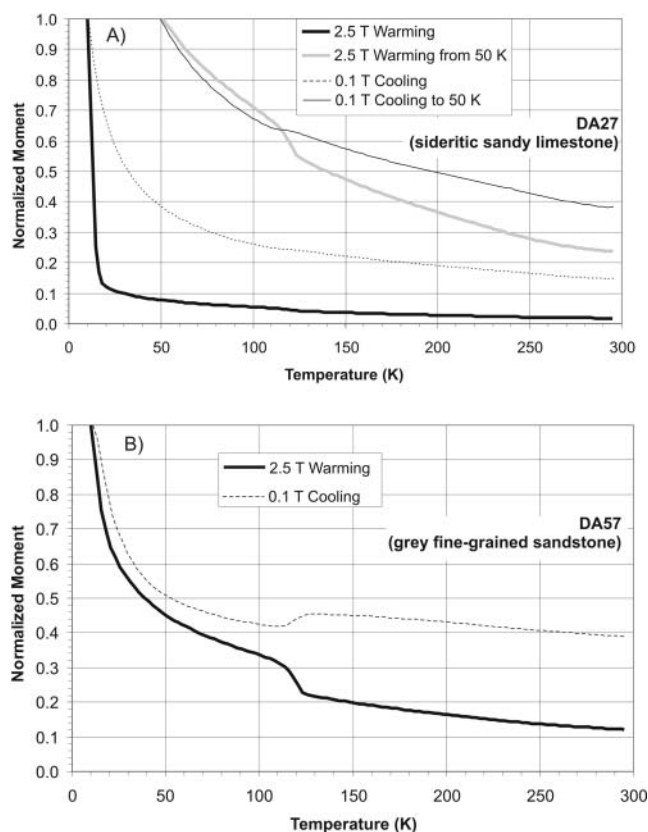


Fig. 8. Low-temperature (zero field cooling) remanence measurements on two samples from Dalsnuten. (a) and (b) each shows a 0.1 T IRM cooled from room temperature to 10 K (thin dotted line). The thick black lines are a 2.5 T IRM applied at 10 K then warmed to room temperature. Those in (a) have also been replotted using the IRM at 50 K to normalize the moment, to see more clearly the Verwey transition in this specimen. Specimen stratigraphic location is shown in Figure 12.

below). These data suggest that, for the DA section, the magnetic signal is dominantly carried by (detrital) magnetite, probably with a wide range of grain sizes, some of which are in part oxidized to maghemite. The specimens from the MEE section display different behaviour showing loss of the IRM (of all coercivities) beginning at about 150 °C and showing total IRM loss by *c.* 450–500 °C (Fig. 7b). This lower temperature loss of the IRM may indicate the finer grain size and/or more extensive oxidation of the magnetite. Alternatively, it may be that an unidentified magnetic sulphide contributes to the magnetization, which is rapidly oxidized on heating.

Palaeomagnetic analysis and results

In many specimens from the MEE section, the NRM intensity loss on thermal demagnetization up to 250 °C was <30%. In contrast, specimens from the DA section often showed intensity loss >85% on thermal demagnetization to *c.* 250 °C (Fig. 9). In both cases most of the remaining NRM intensity was reduced on AF demagnetization up to *c.* 80 mT. Specimen demagnetization produced two magnetization directions: first the A-component, interpreted as an overprint, and then the ChRM component, interpreted as a Triassic dual polarity magnetization.

The A-component, a generally steep, northerly, downward-

directed component, was sometimes removed at lower demagnetization stages (<200 °C), but often persisted to the early stages of AF demagnetization (Fig. 9). Alternating field demagnetization only, up to *c.* 30 mT, was also often effective in removing this component in the DA specimens (Fig. 9a). In 5% of specimens the A-component persisted until complete demagnetization. In all the measured specimens at 589 m and 541 m in the Dalsnuten section, only the A-component was extracted. In the MEE section the A-component has a mean *in situ* direction of 036°, 82° ($\alpha_{95} = 6.0$, $k = 66$, $n = 10$). In the DA section the A-component is notably more dispersed in those specimens with interpreted Triassic reverse magnetizations (mean direction of 306°, 84°; $\alpha_{95} = 22.6$, $k = 4.2$, $n = 34$) than those with interpreted Triassic normal polarity (mean direction of 008°, 85°; $\alpha_{95} = 5.5$, $k = 12$, $n = 62$).

The ChRM component comprised both ENE downward-directed magnetizations (interpreted as Triassic normal polarity) and WSW upward-directed reverse magnetizations (Fig. 10). In the DA section, 60% of specimens display a linear segment (mostly origin fitted) on the Zijderveld plots interpreted as the ChRM. For those specimens subjected to initial thermal demagnetization this component was often isolated from the last thermal demagnetization stage, or early AF demagnetization steps. For those specimens subjected to AF demagnetization alone this was often isolated at high AF fields (*i.e.* >50 mT). The isolation of the ChRM was often made difficult by the ChRM intensity often forming less than 10–15% of the NRM intensity. In the MEE section the ChRM was mostly isolated at AF fields greater than *c.* 50–60 mT (Fig. 9c). These linear-fit specimen ChRM directions produced a mean direction for the DA section of 069°, 69° ($k = 28$, $\alpha_{95} = 2.9$) and for the MEE section of 050°, 68° ($k = 125$, $\alpha_{95} = 4.3$; Table 1; Fig. 10). The reverse and normal populations pass the McFadden & McElhinney (1990) reversal test with class C (observed $\gamma = 11.6$, critical $\gamma = 12.3$). These linear principle component determinations were classed as S-type behaviour and visually classified into three S-type ‘quality’ groups (S_1 , S_2 and S_3), based on the length and precision of the line fit, with S_1 having the highest precision and S_3 the lowest (Fig. 9). The average excess standard deviation parameter (ρ , Kent *et al.* 1983) was 2.0 (average $\alpha_{95} = 11^\circ$), indicating that the average variance for the fitted lines is slightly more than the magnetometer-measured variance. The stratigraphic distribution of this type of demagnetization behaviour shows that specimens in the Tschermakfjellet Fm and the De Geerdalen Fm below *c.* 530 m display the better quality line fits, whereas specimens from the MEE section are dominated by this type of demagnetization behaviour (Figs 11 and 12).

The remaining 30% of specimens failed to produce reliable line fits, but did display fitted great circle paths as a result of incomplete separation of the A-component and ChRM. This type of demagnetization behaviour is termed T-type, and is most clearly evident in reversely magnetized specimens, but is also common in normally magnetized specimens, in which case the length of the great circle path is shorter. Specimens with T-type behaviour were classified qualitatively into best quality (T_1) and inferior quality (T_2 , T_3 ; Fig. 9) great circle trends, based on the directional scatter and the degree of approach to either WSW negative, or ENE positive direction. It is possible to define a polarity for many T-type specimens by the directional trend of the demagnetization path. This type of demagnetization behaviour is most evident in the Isfjorden Mbr (Fig. 12), perhaps because samples from this unit were the most weathered, being nearest to the present-day tundra surface on the mountain top. The average ρ was 2.3 (average $\alpha_{95} = 9.4$), indicating that the

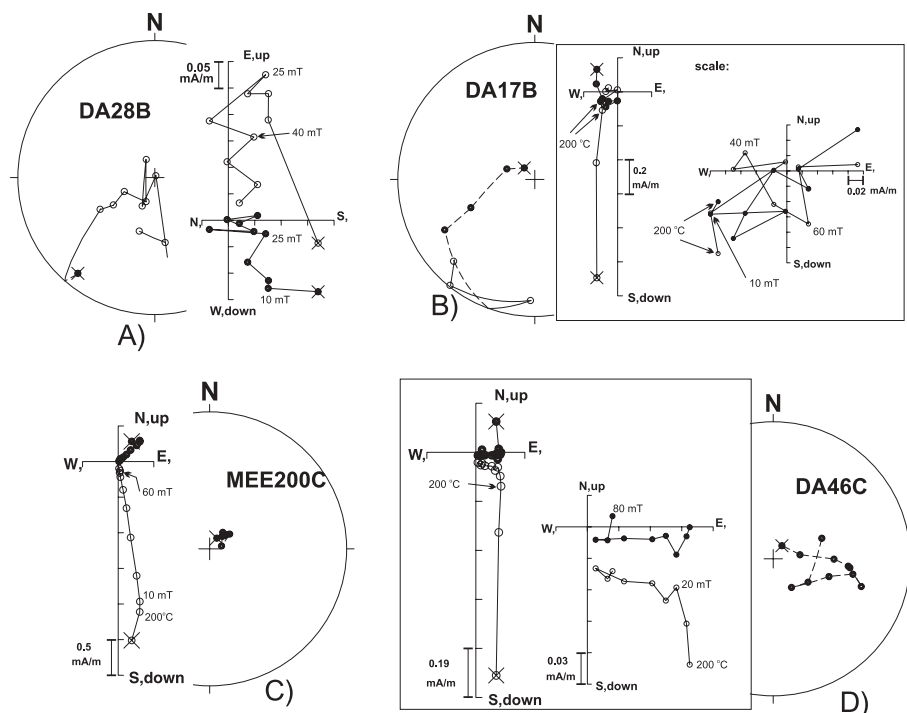


Fig. 9. Representative demagnetization data for specimens in orthogonal vector and stereographic projections (stratigraphic coordinates). (a) DA28B (pale medium-grained sandstone, 460 m) shows the removal of a steep SW downward-directed magnetization (A-component), and the recovery of ChRM of 188° , -72° , from 40 mT to the origin (S1-type, specimen polarity quality R). (b) DA17B (fine-grained sandstone, 520 m) shows the removal of the A-component up to 10 mT, followed by recovery of a ChRM of 242° , -47° between 40 and 60 mT. Central part of Zijdeveld plot shown in close-up (S3-type, polarity quality R?). (c) MEE200C (calcareous shale, 0 m) shows a perhaps contaminated ChRM, which is removed at 60–80 mT of 028° , 72° , which is less steep than the A-component overprint between 10 and 60 mT (S3-type, N?). (d) DA46C (laminated siltstone, 326 m; bedding strike/dip $139^\circ/17^\circ$), shows removal of the A-component (262° , 76°) up to 20 mT. Central part of Zijdeveld plot shown in close-up. No suitable line fit was possible and a plane with pole at 018° , 6° was fitted to the data (T1-type, polarity quality N).

variance required for fitting the great circle planes is on average slightly greater than the magnetometer measured variance, and slightly larger than the average ρ for the line fits.

Using the line-fitted ChRM component and the origin-intersection great circle paths in a combined mean direction (McFadden & McElhinney 1988) for the Tschermakfjellet and De Geerdalen formations produces the mean directions in Table 1. That for the De Geerdalen Fm and the aggregated data for the DA section pass the reversal test (Table 1). An incremental unfolding fold test (Watson & Enkin 1993), using the data from 21 horizons (Fig. 13), produces the largest Fisher k at 90% unfolding, with 95% confidence limits of 73% and 105% unfolding (using a Fisher parametric simulation). The ‘means’ fold test of McFadden (1990), using two bedding clusters (NW and SE groups), gives a value of the f -statistic of 34.2 and 2.7 for 0% and 100% unfolding, respectively, indicating magnetization being acquired at 100% unfolding as a probable scenario, compatible with the confidence limits on the incremental fold

test. The DC fold test of Enkin (2003) produced best unfolding at 90% (95% confidence limit $\pm 9\%$) giving an apparent indeterminate fold test, although at 97.5% confidence ($\pm 10\%$ confidence limits) the fold test is just barely positive. Hence, overall the ChRM appears to predate the formation of the décollement and its associated deformation in the Palaeocene–Eocene.

Magnetostratigraphy

The line-fit ChRM directions from Vendomdalen (i.e. S-type) were converted to virtual geomagnetic pole (VGP) latitude (Opdyke & Channell 1996, p. 91) using the mean direction from each section (great circle combined mean for Dalsnuten; Table 1, Figs 11 and 12). For those specimens from the DA sections that had no line fit, the point on their great circle trend nearest the combined mean (i.e. that used in the combined mean) was used for calculating the VGP latitude (Fig. 12c). All specimens were also assigned a ‘polarity quality’ based on the quality of

Table 1. Fisher means, reversal tests and virtual geomagnetic pole

	Dec.	Inc.	k	α_{95}	n_1/n_p	Reversal test	G_o/G_c	P_{lat}	P_{long}	D_p/D_m	E
Milne Edwardsfjellet section	50.1	67.9	125	4.3	10/0	—	—	57	134	6.0/7.2	5.7
Dalsnuten section											
De Geerdalen Fm	66.2	71.1	24.4	3.1	63/27	Rb	8.3/11.2	59	113	4.7/5.4	1.4
Tschermakfjellet Fm	65.7	67.6	46.3	4.1	27/0	—	—	54	117	5.7/6.8	2.5
All data (fitted lines only)	68.8	69.2	27.8	2.9	90/0	Rc	11.6/12.3	56	115	4.2/4.9	1.6
All data	66.9	70.3	27.3	2.6	90/25	Rb	9.6/10	57	114	3.9/4.5	1.6

n_1 , number of specimens with fitted lines; n_p , number of specimens with great circle planes used in the determining the mean direction; α_{95} , Fisher 95% cone of confidence; k , Fisher precision parameter; G_o , angular separation between the inverted reverse and normal directions; G_c , the critical value for the reversal test. In all cases simulation was used; different k values were applied to reverse and normal specimen populations for the reversal test. A reversal test was not possible for the Tschermakfjellet Fm or the MEE section. P_{lat} , P_{long} , latitude and longitude of the mean virtual geomagnetic pole, respectively; E , eccentricity of distribution measured by the ratio of intermediate to minimum eigenvalues. Rb, passed reversal test with G_o of 5 to 10° ; Rc, passed reversal test with $G_o > 10^\circ$ (McFadden & Elhinny 1990).

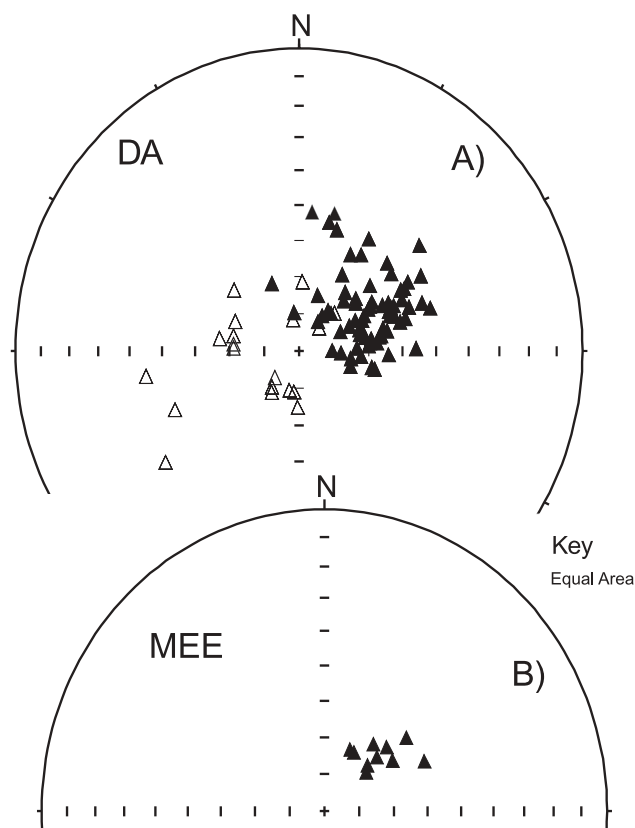


Fig. 10. Stereographic projection of the ChRM directions (S-type) from (a) the Dalsnuten (DA) section and (b) the Milne Edwardsfjellet (MEE) section. Filled symbols, lower hemisphere; open symbols, upper hemisphere.

demagnetization behaviour and, if from T-type specimens, the length and end-point position of the great circle trend. This was used in evaluating the horizon polarity shown in Figures 11 and 12. One specimen of good-quality polarity (i.e. S-type) was sufficient to define the horizon polarity, whereas for specimens of poorer quality at least two were needed.

Major magnetozone reverse and normal couplets in the DA section have been numbered (Fig. 12d) from the base of the section, in the manner proposed by Kent *et al.* (1995) but using the prefix DL. This hierarchical scheme is similar to that used in numbering ocean-floor anomalies, but the oldest part of a magnetozone couplet is of normal polarity (Opdyke & Channell 1996). Five magnetozones (DL1n.1r, DL4n, DL4r, DL5n.1n, DL6n) are defined by multiple specimens at a single horizon and two are defined by a single specimen (DL5n.1r and DL5r.1n). All but one of these are within the Isfjorden Mbr, which indicates the larger rate of polarity change in this unit. This could be because either the Isfjorden Mbr had a lower sedimentation rate than the underlying De Geerdalen Fm, or the frequency of field polarity changes was greater during its deposition.

Discussion

Magnetic overprints

A Tertiary-age, normal polarity, partial remagnetization, associated with the western Spitsbergen orogen, has been widely reported from central, western and southern Spitsbergen (Nawrocki 1999; Michalski & Lewandowski 2004). The 95% confidence cones of the mean overprint directions, in the clearly documented cases (Halvorsen 1972; Sandal & Halvorsen 1973), enclose the Holocene geocentric axial dipole field direction and the European apparent polar wander (APW) path of Besse & Courtillot (2002) from the present to *c.* 50 Ma. This is also the

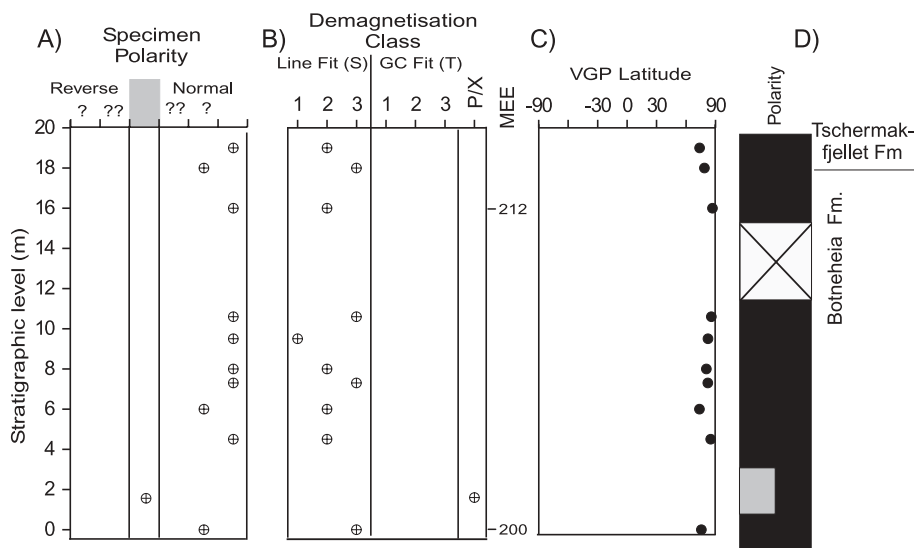


Fig. 11. Milne Edwardsfjellet section magnetostratigraphic data for all specimens. (a) Specimen polarity classification, with reversely magnetized specimens on the left and normal polarity specimens on the right. The ? and ?? columns indicate specimen polarity quality, with ?? the least reliable, ? intermediate quality and unflagged column, excellent quality. The grey flagged column indicates those specimens not assigned a polarity. (b) Demagnetization behaviour, with S-type behaviour (line fits) on the left and great circle (GC) (T-type) behaviour shown. The P/X column indicates those specimens either totally overprinted by the A-component, or mineralogical alteration too severe to interpret any CRM component. The right-hand side shows the location of specimens (MEE code) discussed in the text. (c) Virtual geomagnetic pole (VGP) latitude. Filled symbols, VGP latitude for those samples possessing an S-type ChRM; open symbols, VGP latitude for specimens with T-type behaviour. (d) Interpreted magnetic polarity column. White, reversed polarity; black, normal polarity; grey, uncertain; X, sample gap.

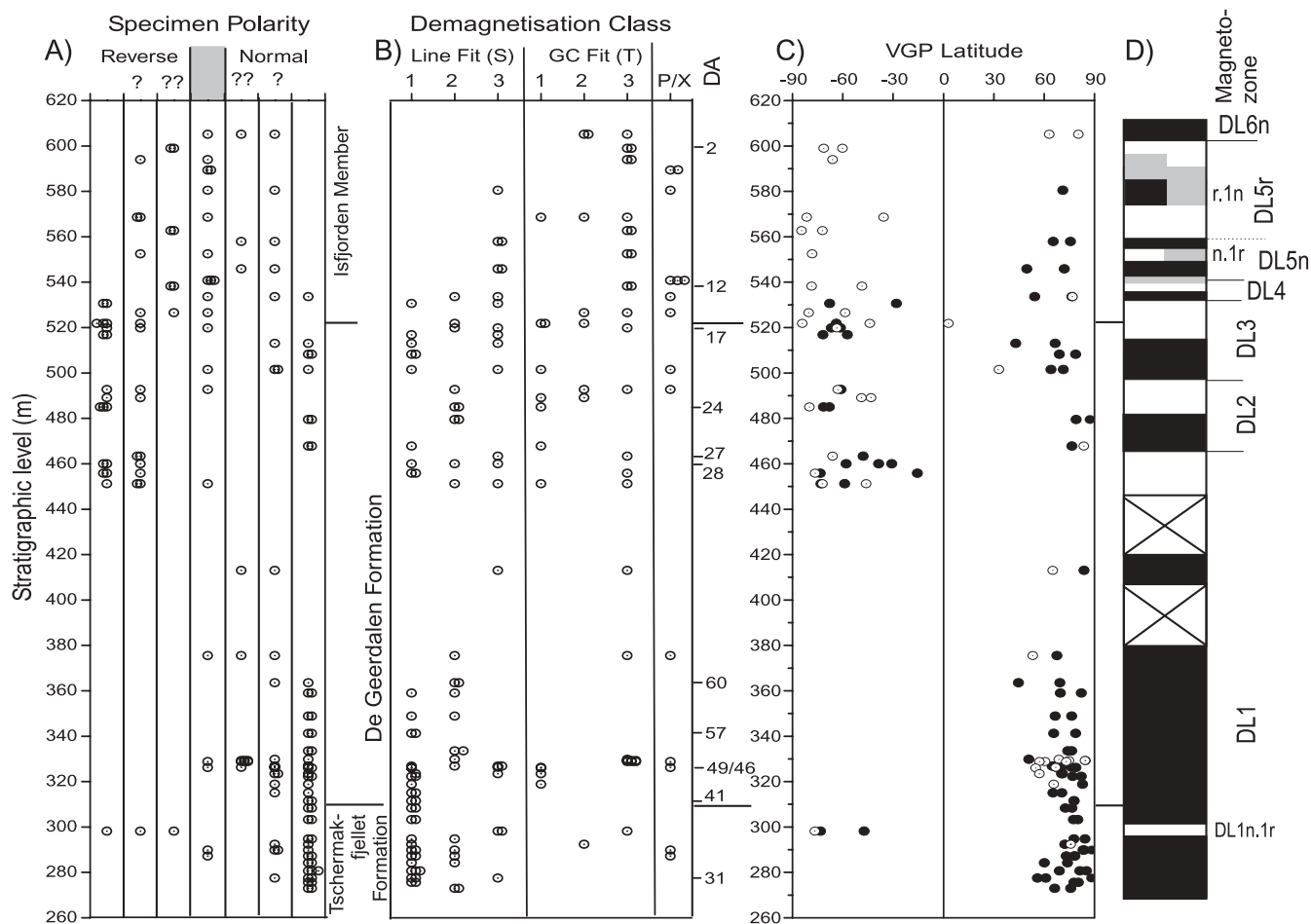


Fig. 12. Dalsnuten section magnetostratigraphic data for all specimens. Details as in Figure 11.

case for the A-component identified here, indicating that it is not possible to statistically distinguish these overprints from a Holocene-age magnetization. Only the C2 overprint of Michalski & Lewandowski (2004) appears distinct from the Holocene field direction. The mean directions of the documented overprints are also statistically different from the primary magnetizations determined from Cretaceous dolerite intrusions common in Svalbard (e.g. Sandal & Halvorsen 1973; Halvorsen 1989), indicating that they do not represent a Cretaceous heating event. We interpret the A-component as being probably a composite (Holocene or Tertiary) magnetization, combined with a small amount of Triassic magnetization. The composite nature of this magnetization explains the much greater directional dispersion of the A-components in those specimens with reverse polarity ChRM directions. We have no data to determine if the overprint is a viscous remanence, a weathering-related (i.e. Holocene) chemical remanence, or a Tertiary thermo-viscous or chemical remanence associated with orogen-related high heat flow (Mørk & Bjørøy 1984); although Spall (1968) documented the partly viscous nature of these overprints.

Palaeomagnetic poles

Most previous studies have suggested that the Svalbard Mesozoic APW path was essentially similar to that of continental Europe (Torsvik *et al.* 1985; Nawrocki 1999). The mean VGPs for the

Tschermakfjellet and the De Geerdalen formations both fall near the western end of the Permo-Triassic segment of the European APW, before it trends to the NE in the Jurassic and Cretaceous (Fig. 14). The Vendomdalen poles fall over the 220–200 Ma segment of the European path defined by Torsvik *et al.* (2001), which utilizes the age model of the palaeomagnetic database (in which the base of the Carnian is 228 Ma). A current best estimate of the radiometric age of the base Carnian is c. 235–228 Ma (Muttoni *et al.* 2004).

The Dalsnuten VGP mean differs from that of the MEE section, which falls closer to the Carnian segment of the European APW path (Fig. 14). We interpret this to indicate longitudinal displacement of the Dalsnuten palaeopoles from the Carnian segment of the European APW. Such displacement is compatible with the high quality Ladinian and Carnian age European palaeopole data of Theveniaut *et al.* (1992) and Edel & Düringer (1997), which fall close to the average European APW (Fig. 14). Our preferred explanation of this discrepancy is a clockwise, vertical-axis rotation of the succession overlying the regional décollement in the underlying Botneheia Fm. This rotation does not appear to have taken place at Milne Edwardsfjellet, perhaps because the deformation of the underlying Botneheia Fm is very minor compared with that at Dalsnuten.

However, this alone cannot account for the displacement of the Dalsnuten results from the European APW track, as restoration of any clockwise vertical-axis rotation would also have the effect of

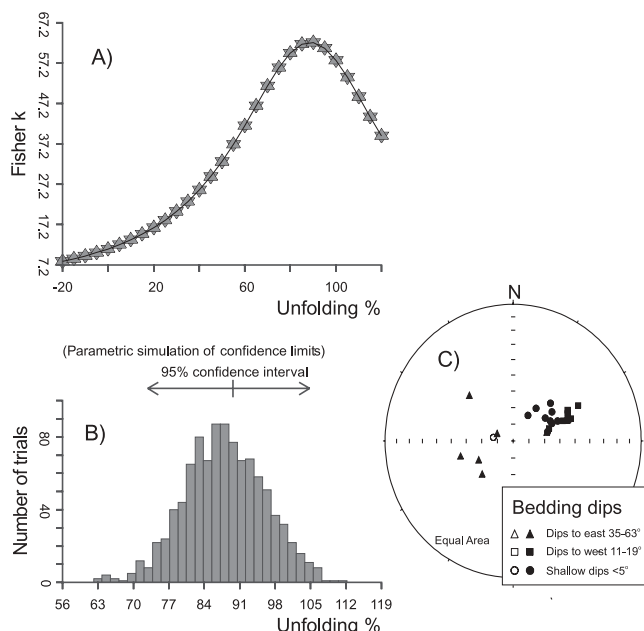


Fig. 13. Fold test data. (a) Fisher concentration parameter (k) v. amount of simultaneous unfolding. (b) Simulation using 1000 trials, with a parametric resampling from a Fisher distribution, to determine the 95% confidence limit on the maximum unfolding percent (Watson & Enkin 1993). (c) Fold test data in *in situ* coordinates, with different symbols for the three groups of bedding dips. Filled symbols (20 horizon means), lower hemisphere; open symbols (1 horizon mean), upper hemisphere.

steepening the palaeopoles from Dalsnuten (Fig. 14). Only with a shallowing of the inclination by some 7° ($\pm 3^\circ$), along with a 15° ($\pm 5^\circ$) anti-clockwise restoration of the directions, do the directions fall close to the Carnian segment of the European APW

path (Fig. 14). This explanation requires some apparent oversteepening of the Vendomdalen inclinations, which could be explained by the contamination of a proportion of the specimen ChRM by the Holocene (or Tertiary) age overprint. This contamination may be most prevalent in the De Geerdalen Fm, as the palaeopole for this formation is displaced further to the north than the palaeopole for the Tschermakfjellet Fm (Fig. 14).

A second possible explanation is that the European APW path may be skewed to lower latitudes by the large number of VGPs in the Triassic APW path derived from sedimentary units at low to mid-palaeolatitude, in which sedimentary inclination shallowing may be common (Kent & Tauxe 2005). Any inclination shallowing of the Dalsnuten data seems unlikely, as the eccentricity in the ChRM component distributions (Table 1) is oriented more north-south (216°), inconsistent with that expected from inclination shallowing, but more consistent with the TK03.GAD statistical field model (Tauxe 2005). An 'elongation-inclination' analysis using the TK03.GAD field model (Tauxe & Kent 2004; Tauxe 2005) on all the data from Dalsnuten gives a shallower than observed inclination solution of 61° (95% confidence limits 50° , 71° , $n = 115$) indicating that it is inappropriate to apply the inclination shallowing model to the Dalsnuten data. Hence, inclination shallowing in the Vendomdalen data does not appear to be significant.

Magnetostratigraphy: comparisons with Tethyan sections

There are currently no other published magnetostratigraphic studies on the Boreal Middle and Upper Triassic, so to obtain parallels with other Triassic successions it is necessary to establish base correlations with better-studied Tethyan sections and the Newark Supergroup in the USA.

The uppermost parts of the Carnian have been extensively studied in a number of sections in Turkey, Slovakia and Italy, where the correlation between sections, and their linking,

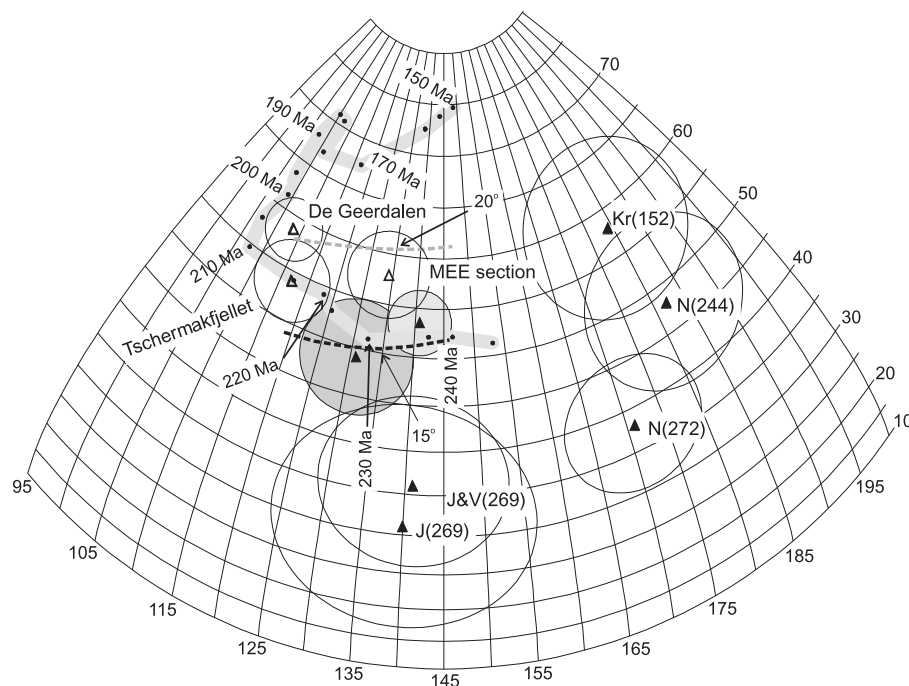


Fig. 14. Comparison of the mean VGP directions for the De Geerdalen and the Tschermakfjellet formations and the MEE section (Δ with 95% confidence cones); with other relevant VGP data from Svalbard and continental Europe. The thick grey line is the mean European APW path from Torsvik *et al.* (2001), with age of poles indicated in 5 Ma increments from 245 Ma. $\blacktriangle$ with unfilled 95% confidence cones are other data from the Late Palaeozoic and Mesozoic (<150 Ma) from Svalbard, labelled with source and age in million years. Kr(152), Krumsiek *et al.* (1968); N(244) and N(272), Nawrocki (1999); J&V(269), Jeleńska & Vincenz (1987); J(269), Jeleńska (1987). $\blacktriangle$ with filled confidence cones are central European Ladinian (light grey) and Carnian (dark grey) palaeopoles from Theveniaut *et al.* (1992) and Edel & Düringer (1997). Dotted grey line is the 'VGP track' if the mean VGP for the whole Dalsnuten section is rotated anti-clockwise. The intersection with 135° longitude is marked with the amount of vertical-axis rotation. The black dotted line is the vertical axis rotated 'VGP track' if the inclination is 7° shallower.

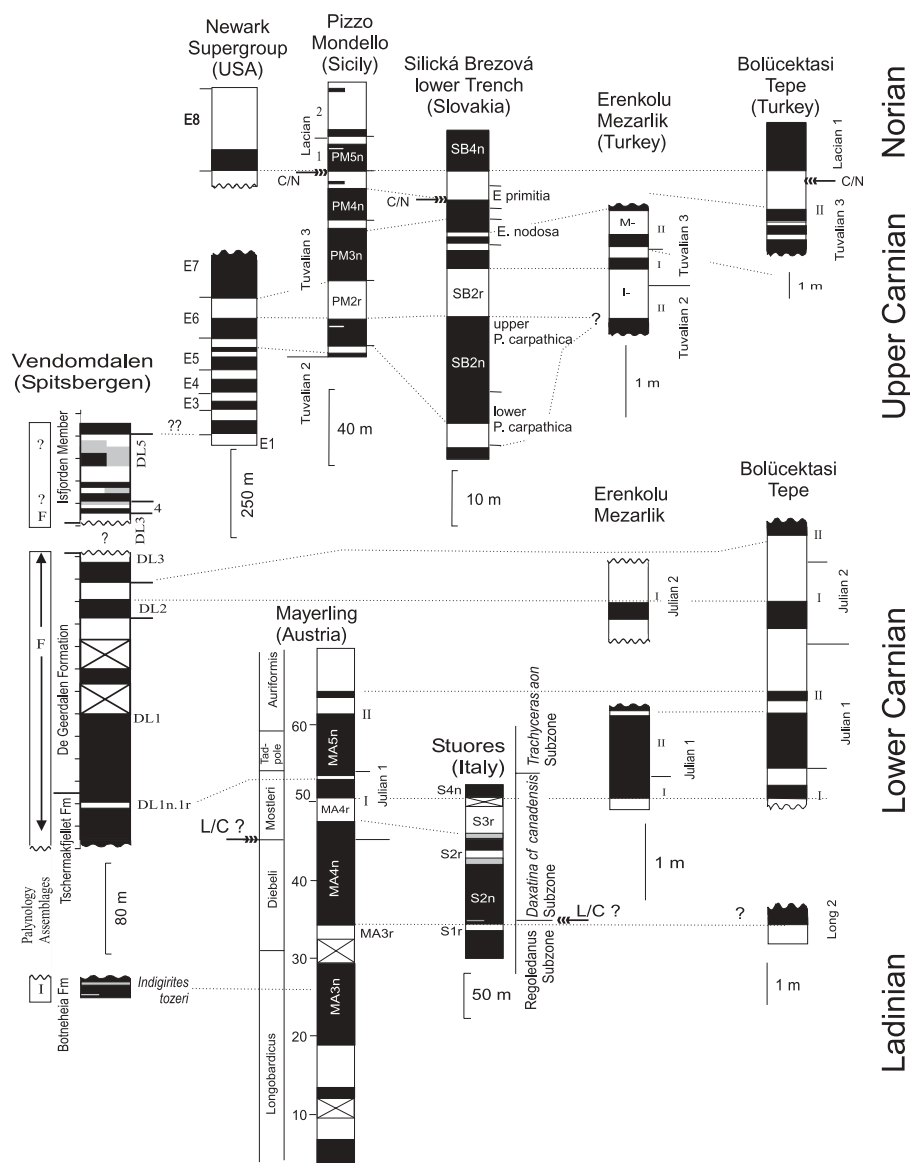


Fig. 15. Other Late Ladinian and Carnian bio-magnetostratigraphies and probable correlations to the Central Spitsbergen sections. Newark Supergroup from Kent *et al.* (1995) and Le Tourneau (1999). Pizzo Mondello from Muttoni *et al.* (2001, 2004). Silická Brezová from Channell *et al.* (2003). Erenkolu Mezarlik, Bolücektasi and Mayerling from Gallet *et al.* (1992, 1994, 1998). Stuoeres from Broglio Loriga *et al.* (1999). All sections in metres thickness scale. C/N, proposed Carnian–Norian boundaries; L/C, proposed Ladinian–Carnian boundaries. Julian–Tuvalian biostratigraphic subdivisions after Gallet *et al.* (1994).

predominantly conodont biostratigraphy, makes magnetostratigraphic intersection correlation clear (Fig. 15; Krystyn *et al.* 2002; Channell *et al.* 2003; Muttoni *et al.* 2004). Extending the correlations downwards into the older parts of the Upper Carnian (Tuvalian) sequences has proved problematic, in that magnetostratigraphy through a section dated to the older parts of the Tuvalian has not been published. The current best estimate of the magnetostratigraphy through the lower Tuvalian is provided by magnetostratigraphic correlation of the Tethyan sections to the lowest parts of the Newark Supergroup. This appears to extend into an interval that is older than any published magnetostratigraphy from Tethyan sections (Muttoni *et al.* 2004; Fig. 15). This suggests that the early part of the Tuvalian was characterized by approximately equal duration reverse and normal magnetozones. Other non-marine Carnian sections do not show such frequent polarity changes as the Newark Supergroup; this feature may suggest that they are incomplete (Molina-Garza *et al.* 1996).

Magneto-biostratigraphies for the Lower Carnian are based around sections in Turkey, Italy and Austria (Fig. 15). The Mayerling and Stuoeres sections cover the transition from the

Ladinian to the Carnian, and apparently provide a good degree of correlation between the magnetostratigraphies, albeit with different proposed conodont and ammonite definitions for the base of the Carnian (Gallet *et al.* 1998; Broglio Loriga *et al.* 1999). The polarity pattern of these sections seems to be confirmed by the condensed (and stratigraphically segmented) pelagic limestone sections at Erenkolu Mezarlik and Bolücektasi Tepe in Turkey (Fig. 15).

Within the Mayerling section in Austria *Neogondolella inclinata* ranges from the base of the section to the top of the Diebeli conodont zone (c. 45 m level), and *Metapolygnathus polygnathiformis* ranges from near the base (46 m level) of the *Mostleri* Zone to the top of the section (Gallet *et al.* 1998). These conodonts provide direct ties to the uppermost part of the MEE section, where the same forms occur, suggesting that the upper part of the Mayerling section (MA5n) correlates in part to the Tchernmakfjellet Fm in Spitsbergen. In addition, *Stolleyites* is known from the *Trachyceras desatoyense* Zone of British Columbia (Fig. 2), which has allowed Broglio Loriga *et al.* (1999) to suggest the equivalence of the *Stolleyites tenuis* Zone

and the Tethyan ammonoid *Trachyceras aon* Subzone of the southern Alps. This suggests that the normal magnetozone DL1n corresponds to the uppermost part of the section at Stuores (S4n) and the predominantly normal polarity interval that ranges from the Julian-1(I) to Julian-1(II) interval at Mayerling (Fig. 15). The reverse magnetozone DL1n.1r may equate to the reverse magnetozone in the uppermost part of Julian-1(I) at Mayerling and Bolücektasi Tepe (Fig. 15). No substantive reverse polarity interval, like that at Mayerling (MA4r) and Stuores (S3r), has been detected at the base of the Tschermafjellet Fm or top of the Botneheia Fm, indicating that this interval is not represented at Vendomdalen, probably because of a disconformity at the base of the Tschermafjellet Fm. These tie points would appear to suggest that DL2n and DL3n correspond to the equivalent normal magnetozones in the Julian-2 parts of Erenkolu Mezerlik and Bolücektasi Tepe.

The palynological assemblages from the *Daxatina canadensis* Subzone at Stuores are most similar to the G and H assemblages of Hochuli *et al.* (1989), in that they contain *Patinasporites summus* and *Camerosporites secatus* (Broglio Lorigo *et al.* 1999). We cannot confidently identify these same assemblage zones in the uppermost Botneheia Fm at Milne Edwardsfjellet. This, together with the fact that reverse polarity intervals have not been detected in the uppermost Botneheia Fm, and that the uppermost Ladinian ammonoid zones are undetected in central Spitsbergen, suggests a substantial hiatus at the Botneheia–Tschermafjellet boundary. Dagys & Konstantinov (1995) suggested that the *Indigirites tozeri* Zone equates to the upper part of the Tethyan *Archelaus* Zone, which suggests that the normal polarity magnetozone in the uppermost Botneheia Fm is the equivalent of the MA3n magnetozone in the Mayerling section (Fig. 15). This is compatible with the age range of the conodont *N. inclinata* (Gallet *et al.* 1994).

The first occurrence of *Aulisporites astigosus* at 281 m and the last occurrence of *Lunatisporites noviaulensis*, *Kuglerina meieri*, *Infernopollenites* spp. and *Sellasporea rugoverrucata* at c. 521 m (Fig. 6) are useful indicators of Lower Carnian (Julian) strata in Alpine sections (Visscher & Brugman 1981; Roghi 2004). Hence, these characteristics of the Dalsnuten palynological assemblages support the magnetostratigraphic correlations, suggesting the De Geerdalen Fm below about 530 m at Dalsnuten is equivalent to the Lower Carnian of the Alps. Roghi (2004) has also suggested that some of these palynological features define an *Aulisporites–Aratisporites* Acme Zone, which characterizes the boundary interval between the Lower and Upper Carnian in Europe. The significant hydromorphic composition of this acme zone across Europe indicates a humid palaeoclimate (Simms *et al.* 1995; Fijalkowska-Mader 1998), which is consistent with the prograding De Geerdalen Fm delta systems, and the carbonaceous-rich character of the formation.

In the palynology from Dalsnuten there are no indications of Norian or younger palynomorphs, such as *Iraquispora* (*Kyrtomisporeis*), which dominate the Early Norian (*Pterosirenites* beds; Fig. 2) assemblages from the base of the Wilhelmøya Subgroup. *Illinites chitonoides*, present in the miospore assemblage at 530 m, is a consistent indicator of Early Carnian strata in the Barents Sea and Svalbard, indicating no clear evidence of Late Carnian strata in the Isfjorden Mbr. The magnetostratigraphy of the Isfjorden Mbr is characterized by an alternation of thin magnetozones (DL4 to DL5n), with, in the upper part, the more regular presence of reverse polarity (DL5r; Fig. 15). If the lowest part of the Newark Supergroup is a reasonable representation of the Late Carnian magnetostratigraphy (Channell *et al.* 2003; Muttoni *et al.* 2004; Fig. 15), then the Isfjorden Mbr does not

seem to be its equivalent. Hence, it seems most probable that the Isfjorden Mbr represents an interval in the mid-Carnian, which is not represented by any published magnetostratigraphies. Clearly, more detailed stratigraphic investigations at this level are necessary to confirm this speculation. The dramatic change in lithological architecture in the Isfjorden Mbr, compared with the underlying parts of the De Geerdalen Fm, is indicative of a major change in depositional style, suggesting a possible hiatus at or slightly below the base of the Isfjorden Mbr (Fig. 15). This hiatus may be equivalent to the mid-Carnian unconformity in the Barents Sea (Van Veen *et al.* 1992), which is there coincident with a major shift in the drainage regime.

Conclusions

This study confirms the close association between what has been traditionally considered the Ladinian–Carnian boundary interval in both the Boreal and Tethyan realms. However, in the marine succession in central Spitsbergen, the latest Ladinian is largely absent because of a hiatus prior to deposition of the Tschermafjellet Fm. Further south in the Barents Sea, on Bjørnøya, a more complete succession across this boundary may be found. The earliest part of the Late Triassic successions on Spitsbergen provides a magnetic polarity record that shows continuous and high deposition rates during the Early Carnian. These magnetostratigraphic correlations are confirmed by ammonoid, palynology and conodont data. Some 200 m of Lower Carnian sediments were deposited in central Spitsbergen, compared with some 120 m for the entire Middle Triassic, indicating that this rapid fill of accommodation space is a major event in the Boreal ocean. It would seem likely that this event is closely linked to a transition from a dry to a wet climate, which reached its most humid phase during the later part of the Early Carnian. This event also seems to be recognized throughout the northwestern Tethys. The upper unit of the De Geerdalen Fm, the Isfjorden Mbr, is tentatively considered to represent an interval spanning the mid-Carnian, with a probable hiatus at its base, which represents a regional mid-Carnian unconformity that is also seen in the Barents Sea.

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References

- ANDRESEN, A., HAREMO, P., SWENSSON, E. & BERGH, S.G. 1992. Structural geology around the southern termination of the Lomfjorden Fault Complex, Agardhdalen, east Spitsbergen. *Norsk Geologisk Tidsskrift*, **72**, 83–91.
- BALINI, M., KRZYSTYN, L., NICORA, A. & TORTI, V. 2001. The Ladinian–Carnian boundary succession in Spiti (Tethys Himalaya) and its bearing to the definition of the GSSP for the Carnian stage (Upper Triassic). *Journal of Asian Earth Sciences*, **19**, 3–4.
- BENTON, M.J. 1993. Late Triassic extinctions and the origin of the dinosaurs. *Science*, **260**, 769–770.
- BESSE, J. & COURTILLOT, V. 2002. Apparent and true polar wander and the geometry of the geomagnetic field over the last 200 Myr. *Journal of Geophysical Research*, doi:10.129/2000JB000050.
- BEUTLER, G. 1998. Keuper. In: BACHMANN, G.H., BEUTLER, G. & LERCH, I. (eds) *Excursions of the International Symposium on the Epicontinental Triassic, Halle (Saale), September 1998*. Hallesches Jahrbuch für Geowissenschaften, Reihe B, Beiheft 6, 45–68.
- BROGLIO LORIGA, C., GIANOLLA, P. & FRANCO LAGHI, G. *ET AL.* 1999. The Prati di Stuores/Stuores Wiesen section (Dolomites, Italy): a candidate Global Stratotype Section and Point for the base of the Carnian Stage. *Rivista Italiana di Paleontologia e Stratigrafia*, **105**, 37–78.

- CHANNELL, J.E.T., KOZUR, H.W., SIEVERS, T., MOCK, R., AUBRECHT, R. & SYKORA, M. 2003. Carnian–Norian biomagnetostratigraphy at Silická Brezová (Slovakia): correlation to other Tethyan sections and to the Newark Basin. *Palaeogeography, Palaeoclimatology, Palaeoecology*, **191**, 65–109.
- DAGYS, A.S. & KONSTANTINOV, A.G. 1995. New zonation scheme of the Ladinian stage for northeast Asia. *Stratigraphy and Geological Correlation*, **3**, 314–320.
- DAGYS, A.S. & KONSTANTINOV, A.G. 1997. A revision of the Nathorstiidae (Ammonoidea) from Northeastern Asia. *Palaeontological Journal*, **31**, 478–488.
- DAGYS, A.S. & WEITSCHAT, W. 1993. Correlation of the Boreal Triassic. *Mitteilungen aus dem Geologisch-Paläontologisches Institut der Universität Hamburg*, **75**, 249–256.
- DAGYS, A.S., WEITSCHAT, W., KONSTANTINOV, A.G. & SOBOLEV, E.S. 1993. Evolution of the boreal marine biota and stratigraphy at the Middle/Upper Triassic boundary. *Mitteilungen aus dem Geologisch-Paläontologisches Institut der Universität Hamburg*, **75**, 192–209.
- DALLMANN, W.K. (ED.) 1999. *Lithostratigraphic Lexicon of Svalbard. Review and Recommendations for Nomenclature Use. Upper Palaeozoic to Quaternary Bedrock*. Norsk Polarinstittutt, Tromsø.
- EDEL, J.B. & DURINGER, P.H. 1997. The apparent polar wander path of the European plate in Upper Triassic–Lower Jurassic times and the Liassic intraplate fracturing of Pangea: new palaeomagnetic constraints from NE France and SW Germany. *Geophysical Journal International*, **128**, 331–344.
- EGOROV, A.Y. & MØRK, A. 2000. The East Siberian and Svalbard Triassic successions and their sequence stratigraphical relationships. *Zentralblatt für Geologie und Paläontologie, Teil, 1*, 1377–1430.
- ENKIN, R.J. 2003. The direction-correction tilt test: an all purpose tilt/fold test for palaeomagnetic studies. *Earth and Planetary Science Letters*, **212**, 151–166.
- FIJALKOWSKA-MADER, A. 1998. Palynostratigraphy, palaeoecology and palaeoclimatology of the Triassic of south-east Poland. In: BACHMANN, G.H. & LERCHE, I. (eds) *Epicontinental Triassic, Vol. 1. Zentralblatt für Geologie und Paläontologie*, **7-8**, 601–628.
- FREDERICH, T., VON DOBENECK, T., BLEIL, U. & DEKKERS, M.J. 2003. Towards the identification of siderite, rhodochrosite and vivianite in sediments by their low temperature magnetic properties. *Physics and Chemistry of the Earth*, **28**, 669–679.
- GALLET, Y., BESSE, J., KRISTYN, L. & MARCOUX, J. 1992. Magnetostratigraphy of the late Triassic Boucekasi Tepe section (south western Turkey): implications for changes in magnetic reversal frequency. *Physics of the Earth and Planetary Interiors*, **93**, 273–282.
- GALLET, Y., BESSE, J., KRISTYN, L., THEVENIAIT, H. & MARCOUX, J. 1994. Magnetostratigraphy of the Mayerling section (Austria) and Erebkolu Merzarlik (Turkey) section: improvement of the Carnian (late Triassic) magnetic polarity time scale. *Earth and Planetary Science Letters*, **125**, 173–191.
- GALLET, Y., KRISTYN, L. & BESSE, J. 1998. Upper Anisian and Lower Carnian magnetostratigraphy from the northern Calcareous Alps (Austria). *Journal of Geophysical Research*, **103**, 605–622.
- HALVORSEN, E. 1972. A paleomagnetic study of two volcanic formations from northern Spitsbergen. *Norsk Polarinstittutt, Arbok*, **1970**, 70–75.
- HALVORSEN, E. 1989. A paleomagnetic pole position of Late Jurassic/Early Cretaceous dolerites from Hinlopenstretet, Svalbard, and its tectonic implications. *Earth and Planetary Science Letters*, **94**, 398–408.
- HAREMO, P. & ANDRESEN, A. 1992. Tertiary décollement thrusting and inversion structures along Billefjorden and Lomfjorden Fault Zones, east central Spitsbergen. In: LARSEN, R.M., BREKKE, H., LARSEN, B.T. & TALLERAAS, E. (eds) *Structural and Tectonic Modelling and its Application to Petroleum Geology*. Norwegian Petroleum Society, Special Publications, **1**, 481–494.
- HOCHULI, P.A., COLIN, J.P. & VIGRAN, J.O. 1989. Triassic biostratigraphy of the Barents Sea area. In: COLLINSON, J.D. (ed.) *Correlation in Hydrocarbon Exploration, Norwegian Petroleum Society*. Graham & Trotman, London, 131–153.
- JELEŃSKA, M. 1987. Aspects of pre-Tertiary paleomagnetism of Spitsbergen and their tectonic implications. *Tectonophysics*, **139**, 99–106.
- JELEŃSKA, M. & VINCENZ, S.A. 1987. Origin of the magnetization of Permian–Carboniferous sediments of Spitsbergen, Svalbard Archipelago. *Earth and Planetary Science Letters*, **85**, 173–182.
- KENT, D.V. & TAUXE, L. 2005. Corrected Late Triassic latitudes for continents adjacent to the North Atlantic. *Science*, **307**, 240–244.
- KENT, D.V., OLSEN, P.E. & WITTE, W.K. 1995. Late Triassic–earliest Jurassic geomagnetic polarity sequence and paleolatitudes from drill cores in the Newark rift basin, eastern North America. *Journal of Geophysical Research*, **100**, 14965–14998.
- KENT, J.T., BRIDEN, J.C. & MARDIA, K.V. 1983. Linear and planar structure in ordered multivariate data as applied to progressive demagnetization of palaeomagnetic remanence. *Geophysical Journal of the Royal Astronomical Society*, **81**, 75–87.
- KING, J.G. & WILLIAMS, W. 2000. Low temperature magnetic properties of magnetite. *Journal of Geophysical Research*, **105**, 16427–16436.
- KNARUD, R. 1980. *En sedimentologisk og diagenetisk undersøkelse av Kapp Toscanaformasjonens sedimenter på Svalbard*. Candidate real. thesis, University of Oslo.
- KORCHINSKAYA, M.V. 1982. *An explanatory note to the stratigraphic scheme of the Mesozoic (Trias) of Svalbard*. Ministry of Geology, USSR, NPO, 'Sevmorgeo', Leningrad, 40–99 [in Russian].
- KRUMSIEK, K., NAGEL, J. & NAIRN, A.E.M. 1968. Record of palaeomagnetic measurements on some igneous rocks from Isfjorden region, Spitsbergen. *Norsk Polarinstittutt, Arbok*, **1966**, 76–83.
- KRISTYN, L., GALLET, Y., BESSE, J. & MARCOUX, J. 2002. Integrated Upper Carnian to Lower Norian biochronology and implications for the Upper Triassic magnetic polarity timescale. *Earth and Planetary Science Letters*, **203**, 343–351.
- LE TOURNEAU, P.M. 1999. *Depositional history and tectonic evolution of the late Triassic age rifts of the U.S. central Atlantic margin: results of an integrated palaeomagnetic analysis of the Taylorville and Richmond basins*. PhD thesis, Columbia University, New York.
- LOWRIE, W. 1990. Identification of ferromagnetic minerals in a rock by coercivity and unblocking temperature profiles. *Geophysical Research Letters*, **17**, 159–162.
- LUCAS, S.G. 1998. The epicontinental Triassic, an overview. In: BACHMANN, G.H. & LERCHE, I. (eds) *Epicontinental Triassic, Vol. 1. Zentralblatt für Geologie und Paläontologie*, **7-8**, 475–496.
- McFADDEN, P.L. 1990. A new fold test for palaeomagnetic studies. *Geophysical Journal International*, **103**, 163–169.
- McFADDEN, P.L. & McELHINNEY, M.W. 1988. The combined analysis of remagnetization circles and direct observations in palaeomagnetism. *Earth and Planetary Science Letters*, **87**, 161–172.
- McFADDEN, P.L. & McELHINNEY, M.W. 1990. Classification of the reversal test in palaeomagnetism. *Geophysical Journal International*, **103**, 725–729.
- MICHALSKI, K. & LEWANDOWSKI, M. 2004. Palaeomagnetic results from the Middle Carboniferous rocks of the Hornsund region, southern Spitsbergen: preliminary report. *Polish Polar Research*, **25**, 169–182.
- MOLINA-GARZA, R.S., GEISSMANN, J.W., LUCAS, S.G. & VAN DER VOO, R. 1996. Paleomagnetism and magnetostratigraphy of Triassic strata in the Sangre de Cristo Mountains and Tucumari Basin, New Mexico, USA. *Geophysical Journal International*, **124**, 935–953.
- MØRK, M.B.E. 1999. Compositional variations and provenance of Triassic sandstones from the Barents Shelf. *Journal of Sedimentary Research*, **69**, 690–710.
- MØRK, A. & BJØRØY, M. 1984. Mesozoic source rocks of Svalbard. In: SPENSER, A.M. ET AL. (ed.) *Petroleum Geology of the North European Margin*. Norwegian Petroleum Society. Graham and Trotman, London, 371–382.
- MØRK, A., KNARUD, R. & WORSLEY, D. 1982. Depositional and diagenetic environments of the Triassic and Lower Jurassic succession of Svalbard. In: EMBRY, A.F. & BALKWILL, H.R. (eds) *Arctic Geology and Geophysics*. Canadian Society of Petroleum Geologists, Memoirs, **8**, 371–398.
- MØRK, A., EMBRY, A.F. & WEISCHAT, W. 1989. Triassic transgressive–regressive cycles in the Sverdrup Basin, Svalbard and the Barents Shelf. In: COLLINSON, J.D. (ed.) *Correlation in Hydrocarbon Exploration*. Norwegian Petroleum Society. Graham and Trotman, London, 113–130.
- MØRK, A., VIGRAN, J.O. & HOCHULI, P.A. 1990. Geology and palynology of the Triassic succession of Bjørnøya. *Polar Research*, **8**, 141–163.
- MØRK, A., VIGRAN, J.O., KORCHINSKAYA, M.W., PCHELINA, T.M., FEFILOVA, L.A., VAVILOV, M.N. & WEITSCHAT, W. 1993. Triassic rocks in Svalbard, the arctic Soviet Islands and the Barents Shelf: bearing on their correlations. In: VORREN, T.O., BERGSAGER, E., DAHL-STAMNES, Ø.A., HOLTER, E., JOHANSEN, B., LIE, E. & LUND, T.B. (eds) *Arctic Geology and Petroleum Potential*. Norwegian Petroleum Society, Special Publications, **2**, 457–479.
- MØRK, A., DALLMANN, W.K. & DYPVIK, H. ET AL. 1999. Mesozoic lithostratigraphy. In: DALLMANN, W.K. ET AL. (ed.) *Lithostratigraphic Lexicon of Svalbard. Review and Recommendations for Nomenclature Use. Upper Palaeozoic to Quaternary Bedrock*. Norsk Polarinstittutt, Tromsø, 127–214.
- MUTTONI, G., KENT, D.V., DI STEFANO, P., GULLO, M., NICORA, A., TAIT, J. & LOWRIE, W. 2001. Magnetostratigraphy and biostratigraphy of the Carnian/Norian boundary interval from the Pizzo Mondello section (Sicani Mountains, Sicily). *Palaeogeography, Palaeoclimatology, Palaeoecology*, **166**, 383–399.
- MUTTONI, G., KENT, D.V., OLSEN, P.E., DI STEFANO, P., LOWRIE, W., BERNASCONI, S.M. & HERNANDEZ, F.M. 2004. Tethyan magnetostratigraphy from Pizzo Mondello (Sicily) and correlation to the late Triassic Newark astrochronological polarity time scale. *Geological Society of America Bulletin*, **116**, 1043–1058.
- NAWROCKI, J. 1999. Paleomagnetism of Permian through Early Triassic sequences in central Spitsbergen: implications for palaeogeography. *Earth and Planetary Science Letters*, **169**, 59–70.
- OPDYKE, N.D. & CHANNELL, J.E. (EDS) 1996. *Magnetic Stratigraphy*. International

- Geophysical Series, **64**.
- ORCHARD, M.J. & TOZER, E.T. 1997. Triassic conodont biochronology, its calibration with the ammonoid standard and a biostratigraphic summary for the western Canada sedimentary basin. *Bulletin of Canadian Petroleum Geology*, **45**, 675–692.
- ÖZDEMİR, Ö., DUNLOP, D.J. & MOSKOWITZ, B.M. 1993. The effect of oxidation on the Verwey transition in magnetite. *Geophysical Research Letters*, **20**, 1671–1674.
- PCHELINA, T.M. 1980. New data on the Triassic/Jurassic boundary beds in the Svalbard archipelago. In: SEMERDKIY, D.V. (ed.) *Geology of the Sedimentary Deposits of the Archipelago of Svalbard. Collection of Scientific Papers*. NIIGA, Leningrad, 44–60 [in Russian].
- ROGHI, G. 2004. Palynological investigations in the Carnian of the Cabe del Predil area (Julian Alps, NE Italy). *Review of Palaeobotany and Palynology*, **132**, 1–35.
- SANDAL, S.T. & HALVORSEN, E. 1973. Late Mesozoic palaeomagnetism from Spitsbergen: implications for continental drift in the Arctic. *Physics of the Earth and Planetary Interiors*, **7**, 125–132.
- SIMMS, M.J., RUFFEL, A.H. & JOHNSON, L.A. 1995. Biotic and climatic changes in the Carnian (Triassic) of Europe and adjacent areas. In: FRASER, N.C. & SUES, H.D. (eds) *In the Shadow of the Dinosaurs—Early Mesozoic Tetrapods*. Cambridge University Press, Cambridge, 352–365.
- SMIRNOV, A.V. & TARDUNO, J.A. 2002. Magnetic field control of the low-temperature magnetic properties of stoichiometric and cation-deficient magnetite. *Earth and Planetary Science Letters*, **194**, 359–368.
- SPALL, H. 1968. Anomalous palaeomagnetic poles from Late Mesozoic dolerites from Spitsbergen. *Earth and Planetary Science Letters*, **4**, 33–78.
- TAUXE, L. 2005. Inclination flattening and the geocentric axial dipole hypothesis. *Earth and Planetary Science Letters*, **233**, 247–261.
- TAUXE, L. & KENT, D.V. 2004. A simplified statistical model for the geomagnetic field and the detection of shallow bias in palaeomagnetic inclinations: was the ancient geomagnetic field dipolar? In: CHANNELL, J.E.T., KENT, D.V., LOWRIE, W. & MEERT, J.G. (eds) *Timescales of the Paleomagnetic Field*. American Geophysical Union Monograph, **145**, 101–116.
- THEVENIAUT, H., BESSE, J., EDEL, J.B., WESTPAHL, M. & DURINGER, P. 1992. A Middle Triassic palaeomagnetic pole for the Eurasian plate from Heming (France). *Geophysical Research Letters*, **19**, 777–780.
- TORSVIK, T.H., LØVLIE, R. & STURT, B.A. 1985. Palaeomagnetic argument for stationary Spitsbergen relative to the British Isles (Western Europe) since the late Devonian and its bearing on North Atlantic reconstruction. *Earth and Planetary Science Letters*, **75**, 278–288.
- TORSVIK, T.H., VAN DER VOO, R., MEERT, J.G., MOSAR, J. & WALDEHAUG, H.J. 2001. Reconstructions of the continents around the North Atlantic at about the 60th parallel. *Earth and Planetary Science Letters*, **187**, 55–69.
- VAN VEEN, P.M., SKJOLD, L.J., KRISTENSEN, S.E., RASMUSSEN, A., GJELBERG, J. & STØLAN, T. 1992. Triassic sequence stratigraphy in the Barents Sea. In: VORREN, T.O., BERGSAGER, E., DAHL-STAMNES, Ø.A., HOLTER, E., JOHANSEN, B., LIE, E. & LUND, T.B. (eds) *Arctic Geology and Petroleum Potential*. Norwegian Petroleum Society, Special Publications, **2**, 515–538.
- VIGRAN, J.O., MANGERUD, G., MØRK, A., BUGGE, T. & WEITSCHAT, W. 1998. Biostratigraphy and sequence stratigraphy of the Lower and Middle Triassic deposits from the Svalis Dome, Central Barents Sea, Norway. *Palynology*, **22**, 89–141.
- VISSCHER, H. & BRUGMAN, W.A. 1981. Ranges of selected palynomorphs in the alpine Triassic of Europe. *Review of Palaeobotany and Palynology*, **34**, 115–128.
- WATSON, G.S. & ENKIN, R.J. 1993. The fold test in palaeomagnetism as a parameter estimation problem. *Geophysical Research Letters*, **20**, 2135–2137.
- WEITSCHAT, W. & DAGYS, A.A. 1989. Triassic biostratigraphy of Svalbard and a comparison with N.E. Siberia. *Mitteilungen aus dem Geologisch-Paläontologisches Institut der Universität Hamburg*, **68**, 179–213.
- WEITSCHAT, W. & LEHMANN, H. 1983. Stratigraphy and ammonoids from the Middle Triassic Botneheia Formation (Daonella Shales) of Spitsbergen. *Mitteilungen aus dem Geologisch-Paläontologisches Institut der Universität Hamburg*, **54**, 27–54.

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