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Preliminary results of a study of four successive sedimentary geomagnetic reversal records from the Mediterranean (Upper Thvera, Lower and Upper Sidufjall, and Lower Nunivak)

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The results of a study of four successive Early Pliocene geomagnetic reversal records are presented. The polarity transitions have been recorded in the Trubi formation of Calabria (S. Italy) which is composed of alternating carbonate-poor and carbonate-rich marly layers with moderately high (>5 cm ka^{-1}) accumulation rates. The multicomponent NRM contains a well-defined primary component in which two dominant magnetic phases can be recognized: a magnetite and a low-temperature (LT; i.e., removed at moderate temperatures) phase. Most probably the LT phase is pyrrhotite. The combined presence of pyrrhotite and magnetite in the sediment restricts the validity of a X_{ARM}–X test for paleointensity studies. The magnetite magnetizations reveal a smoother record relative to the records of the LT phase magnetizations. Between the two phases, no significant delay in the moment of remanence acquisition is observed. The durations of the major directional changes for the reversals are estimated to be in the range 0.3–3.0 ka. The VGP path of the Upper Thvera record follows a great-circle 90° W of the site meridian, indicating a clearly non-zonal transitional field, whereas the VGP paths of the other reversal records are close to the site meridian, indicating a mainly zonal transitional field. The lower Sidufjall and Lower Nunivak records are nearly identical and have a zonal harmonic content similar to records reported for the Matuyama–Brunhes polarity transition.

1. Introduction

One of the most outstanding physical properties of the Earth is that its magnetic field reverses polarity at irregular intervals. This feature was not known until the beginning of this century; at that time it was noticed that certain extrusive rocks contain remanent magnetizations with oppositely directed magnetizations. These reversed magnetizations were proved to be caused by the normal or reversed geomagnetic fields that existed during the formation of these rocks. The recordings in rocks of past reversals are probably the only source of information concerning this phenomenon. In recent years, geomagnetic reversal records in various kinds of rocks have been the subject of extensive studies (e.g., Valet, 1985; Mankinen et al., 1985; Prevot et al., 1985). A review of recent developments is provided by Bogue and Hoffman (1987).

So far, the only generally accepted conclusion to be drawn concerning geomagnetic field behaviour during a polarity transition is that the dipole field disappears. This conclusion was first suggested after the observation of diverging virtual geomagnetic pole (VGP) paths calculated from data of two Matuyama–Brunhes polarity transition records located at remote sites on the Earth's surface (Hillhouse and Cox, 1976). Meanwhile, many other similar observations have confirmed the validity of this conclusion (see, for example, Hoffman, 1977).

Various studies have revealed characteristics of reversal records which may well be properties of the transitional geomagnetic field. For instance, it

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has been observed that directional and intensity variations in successive reversals records at a single site are not always identical (Valet, 1985; Valet et al., 1988). This may indicate that the mechanism which causes the observed transitional magnetic field directions and intensities changes with time.

The reversal process appears to be fast relative to the geological time-scale. The various observations suggest that the durations of reversals are not always identical. For instance, the results from the Steens Mountain record of a Miocene reversal (Mankinen et al., 1985) indicate that main directional changes associated with the reversal take place in ~ 0.7 ka, whereas results from a sedimentary record of the Upper Olduvai reversal indicate a duration of ~ 11.1 ka (Clement and Kent, 1985). The period in which anomalous intensities associated with the reversals occurs is a multiple of the duration of the directional changes.

At first sight, diverging data of different reversals suggest that the transitional field cannot be generalized. However, we are dealing with a geometrical problem; a geomagnetic reversal may give different effects at distant sites on Earth. The standing field model (Hillhouse and Cox, 1976) and the flooding model (Hoffman, 1977, 1979; Hoffman and Fuller, 1978), which are essentially phenomenological, predict the transitional geomagnetic field at spatially distributed sites on the Earth's surface as well as between successive reversals at the same site.

The study of similarities and dissimilarities in reversal records at geometrically distributed sites may shed light on the physical origin of the reversal process. The strategy of systematically studying geographically distributed records of the same reversal, or of studying successions of reversal records at one site has been followed by some investigators (e.g., Valet, 1985; Clement and Kent, 1986). However, most of their conclusions have been of a different tenor. Some successions of reversal records suggest a mainly zonal transitional field geometry (Bogue and Coe, 1982, 1984); others indicate significant non-zonal field components (Clement and Kent, 1984).

The harmonic content of reversal records with zonal transitional field geometries is also subject to some discussion. For instance, Bogue and Coe (1982, 1984) suggested that a specific reversal started with an octupolar transitional field, whereas near the termination of the reversal a quadrupolar zonal harmonic component dominated. In some successions of reversal records antipodal transitional fields have been observed (Valet and Laj, 1984). Another succession of reversal records provided evidence that the transitional field was invariant (Clement and Kent, 1985). These observations suggest different types of reversal mechanisms (Bogue and Hoffman, 1987); this idea has also been suggested on theoretical grounds (Olson, 1983).

The fact that these extreme speculations are based on a limited number of (sequences of) reversal records illustrates the need for more data from reversal records.

Reversals of the geomagnetic field have been recorded in sediments and in intrusive and extrusive igneous rocks. Although the high-quality spot readings of the geomagnetic field in extrusive igneous rocks can be successfully used in the study of reversal records, it should be realized that high flow rate sequences of lavas with reversal records are seldomly found (Bogue and Coe, 1982, 1984; Mankinen et al., 1985; Prevot et al., 1985). Geomagnetic reversals are most successfully studied in sedimentary records since they are widespread.

Sediments that are used in reversal studies have to satisfy certain conditions: the sedimentary reversal record must reveal the major directional and intensity variations of the transitional geomagnetic field. For instance, this implies that the accumulation rate for the sediment must be at least 1 cm ka^{-1}. The accumulation rate near the reversal record is usually extracted from the boundary ages of the underlying and overlying polarity intervals and the stratigraphic lengths of the recorded polarity intervals. If temporal information is to be extracted from the record, the variation in the accumulation rate on the scale of the reversal record must be small.

Sediments often contain a variety of magnetic minerals (Freeman, 1986). The origin of these minerals may be depositional (authigenic or secondary). Each of these minerals may contribute in a different degree to the NRM. As a result the total (multicomponent) NRM may be very com-

Fig. 1. Location of Lower Pliocene Trubi formation sampled for this study.

plex. Reversal records in red beds (Herrero-Bervera and Helsley, 1983; Tauxe and Badgley, 1984) suggest that backtracking, i.e., the varying predominance of components acquired at slightly different times in a rapidly evolving transitional field, may occur. In order to study a sedimentary reversal record properly one needs to try and isolate the remanence directions of all primary magnetic minerals with progressive demagnetization techniques and to determine the relative timing of the magnetic remanence.

Investigators from the Utrecht paleomagnetic laboratory have been able to identify a great number of reversal records in the late Miocene marine deposits of Crete (Langereis, 1984). Detailed studies have been made of the rock-magnetic properties of the sediment (Langereis, 1984; Linssen, 1984; Valet, 1985; Chang and Kirschvink, 1985). The reversal records were extensively and successfully investigated at the paleomagnetic laboratory in Gif-sur-Yvette (Valet, 1985). The magnetostratigraphically well-defined (Zijderveld et al., 1986) Lower Pliocene marine deposits of southern Italy (38°15′N, 16°12′E; See Fig. 1), known as the Trubi formation (see, for example, Cita, 1973; Hilgen, 1987), appeared to be very suitable for making a study of a succession of undisturbed, well-dated reversal records. The sediments from the Trubi formation have rock-magnetic properties quite similar to those of the Cretan sediments.

In this paper we present the results of four successive reversal records from the Gilbert chron sampled in the Trubi formation: the Upper Thvera, Lower and Upper Sidufjall and the Lower Nunivak. Before discussing the reversal records the magnetic minerals which contribute to the NRM are discussed. In the general discussion an attempt will be made to determine more accurately the accumulation rate of the sediment in which the reversals are recorded.

2. Geology

The Trubi formation consists of alternating layers of marine marls with high and low carbonate contents; these layers are denoted as High Carbonate Units (HCU) and Low Carbonate Units (LCU) respectively. The well-defined unit boundaries facilitate the recognition of the few minor faults. The thickness of the individual units is laterally constant (on the scale of the exposure: a few hundred metres), the layer thickness of different units ranges from 0.1 to 1.3 m. Each of the four reversals are recorded within a single unit, none of them at a boundary. The bedding plane of the sediment at the site dips 6.5° to the east. Magnetostratigraphic results (Zijderveld et al., 1986) show that between the Upper Thvera and the Lower Nunivak the mean accumulation rate in the part of the section sampled for this study increased from 4.9 to 7.6 cm ka^{-1}.

3. Sampling

The sampling equipment consisted of a portable generator, electric drilling machines and water-cooled brass drills with diamond cutting edges. The drills have a diameter of 2.54 cm. Planes were cut perpendicular to the bedding plane at least 20 cm deeper than the shallowest fresh rock surface. In the Upper Thvera and Lower Nunivak records, sampling levels were marked on these planes at stratigraphic intervals ranging from 2.5 to 5.0 cm; at each level 3–5 cores were drilled in a direction parallel to the bedding plane. The Lower and Upper Sidufjall records and a part of the Upper Thvera record were sampled using a slightly different method: the number of cores per level was 1–2, the spacing between the levels was varied from 0.5 to 2.5 cm. A few additional verti-

cally drilled samples were obtained from the Lower Sidufjall record. For this purpose, special drills were used with a length of 20 cm.

4. Instrumentation

Thermal demagnetizations were performed in tunnel furnaces with mu-metal shielding. The alternating field demagnetizations were applied to the specimens in stationary holders.

Most of the magnetization measurements were performed with a 2G-Enterprises cryogenic magnetometer having a sensitivity of 10^{-11} A m^2; a small number of samples were measured with a modified Jelinek spinner magnetometer having a sensitivity of 10^{-10} A m^2. The reproducibility of both magnetometers is better than 1° for magnetizations exceeding 10^{-9} A m^2.

5. NRM components

5.1. Magnetic phases of the NRM components

The demagnetization diagrams and derived decay curves of Fig. 2 show rock-magnetic characteristics that are representative for the sediment. The components which point to the origin of the diagrams have near-reversed axial dipole field (RADF) directions; these components are therefore almost certainly primary ones (see also Zijderveld et al., 1986).

Figure 2a, c show demagnetization diagrams of two specimens from nearly the same horizon in a sedimentary interval above the Upper Thvera reversal record. Specimen UT.31.7A has been subjected to a thermal demagnetization and specimen UT.32.5A to an alternating field (AF) demagnetization treatment. Both demagnetization diagrams show a primary component and a viscous component. The viscous component is shallow and westward directed. The latter component is an artefact since its direction is highly correlated with the drilling direction. Similar viscous components, although generally with a lower magnitude, are observed in the demagnetization results of many specimens. The normalized decay curves (Fig. 2b, d) show that the viscous component is removed in an alternating field of 15 mT and at a temperature of 240 °C.

The primary component has a unimodal decay when demagnetized in an AF and a bimodal decay when thermally demagnetized. Apparently, the primary component is carried by a low-temperature (LT) magnetic phase of which the remanence is removed below 360 °C and by a high-temperature (HT) magnetic phase of which the remanence is removed just below 600 °C. It is also clear from Fig. 2b that the thermal decay curves of the viscous component and of the LT phase overlap to a significant extent. The overlap of the decay curve of the viscous component and of the primary component in the AF decay curve (Fig. 2d) is less significant.

Figure 2e, g show demagnetization diagrams of two specimens from neighbouring sites from an interval above the Upper Sidufjall reversal record. Specimen US.96A has been subjected to a thermal demagnetization and specimen US.96B to an AF demagnetization treatment. The demagnetization diagrams and the decay curves (Fig. 2f, h) also show a primary and a viscous component. A few other characteristics distinguish them from the previous examples: the thermal demagnetization diagram has a primary component with a trimodal decay due to the presence of a third magnetic phase, the remanence of which is removed just below 700 °C. The LT phase has a slightly higher inclination than the phases whose remanences are broken down at higher temperatures. Table I gives the directions of the phases with near-RADF directions from Fig. 2. Note that the inclination of the primary component from the Upper Sidufjall specimens obtained with AF demagnetization is intermediate to the directions of the LT and HT phases obtained with thermal demagnetization. Note also that the decay of the primary component in the AF decay curves (Fig. 2d, h) is unimodal. Apparently the LT and HT phases have (almost) completely coinciding coercive force spectra in the range 20–70 mT.

Primary components with a similar decay are common in marine marls (Langereis, 1984; Valet, 1985). The two magnetic phases whose remanences are broken down just below 600 and 700 °C are most probably magnetite (Curie temperature:

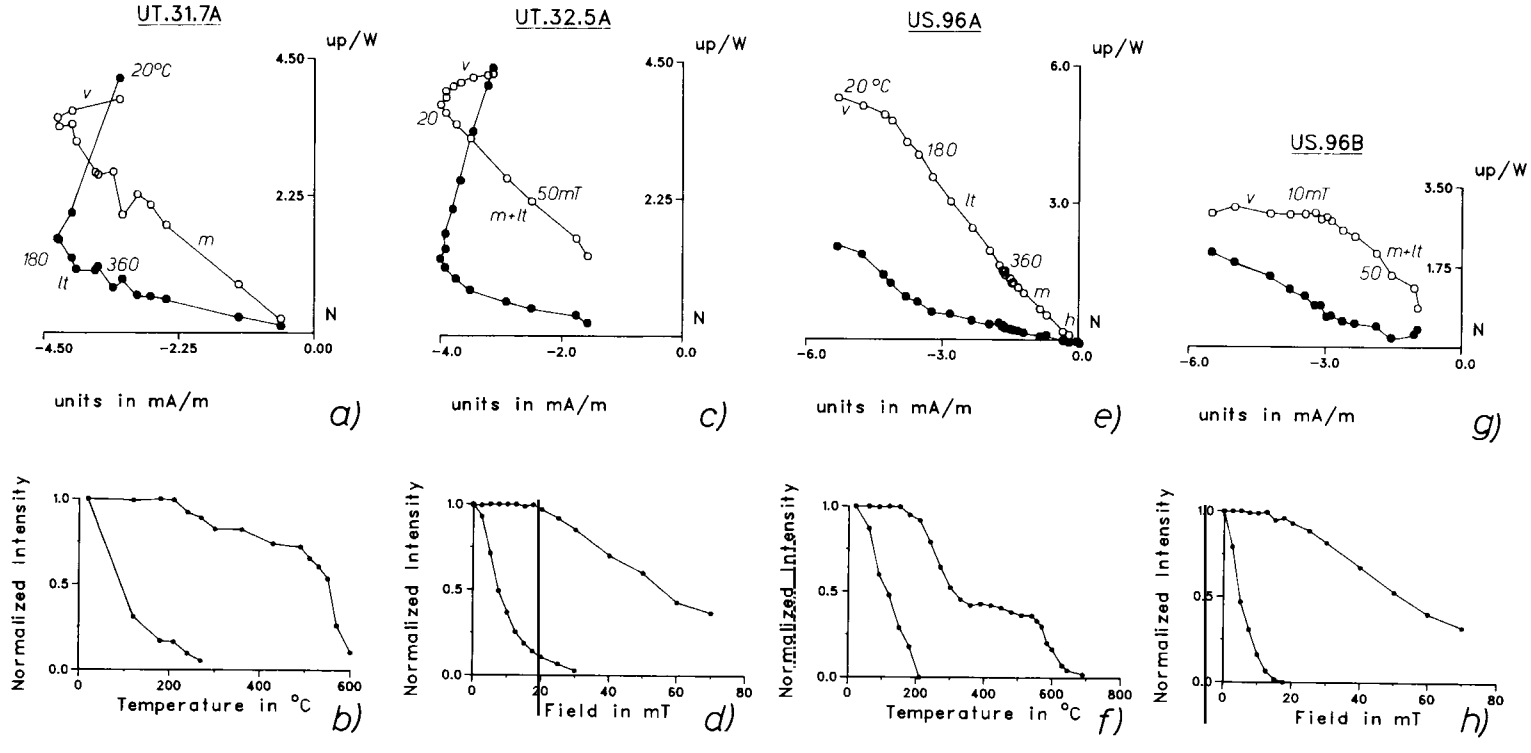

Fig. 2. Orthogonal projection diagrams (Zijderveld, 1967) and corresponding decay curves. In the demagnetization diagrams the closed (open) symbols denote the projection on the horizontal (vertical) plane. Specimens are from levels above the Upper Thvera (UT) and Upper Sidufjall (US) reversal records. Figure (a) and (e) are thermal demagnetization diagrams and (c) and (g) are alternating field demagnetization diagrams. 'm', 'lt' and 'v' denote respectively: the magnetite phase, the low-temperature (LT) phase and the viscous component. The normalized decay curves (b, d, f, h) show the decay of the viscous component and the primary component (m + lt).

TABLE I

Least-squares declination and inclination values of the primary components from the demagnetization diagrams depicted in Fig. 2

Specimen	Total reversed component		LT component		HT component	
	D	*I*	*D*	*I*	*D*	*I*
UT.31.7A			194.1	−38.5	192.3	−36.2
UT.32.5A	190.7	−40.1				
US.96A			190.4	−51.2	187.8	−40.1
US.96B	192.3	−45.3				

Column 1: demagnetization data points exceeding 240 °C or 15 mT.
Column 2: demagnetization data between 240 and 360 °C.
Column 3: demagnetization data after 360 °C.

578 °C) and hematite (Néel temperature: 680 °C). The magnetite phase is observed in all thermally treated specimens. The hematite phase is observed occasionally and always has a low intensity; therefore, the hematite phase is not discussed in the following sections. The determination of the LT phase is discussed in Section 6.

5.2. *Directions of NRM components outside the reversal records*

The study of viscous and primary NRM components outside the reversal records provides a tool for distinguishing these components in the reversal record where we are looking for primary components with unknown transitional directions.

Throughout the sedimentary intervals above and below the reversal records it was observed that the directions of the primary component obtained with thermal demagnetization comprise two linear segments that make an angle of 0–20 °. The linear segments correspond to the LT phase and the magnetite phase. Often, the angle between the linear segments is only visible in the projection on the vertical plane. Generally, the magnetite phase has inclinations that are ~ 10 ° lower than the local axial dipole field inclination (52 °). The LT phase has inclinations that are intermediate to the inclinations of the magnetite phase and to the axial dipole field inclination. The systematic inclination error of the magnetite phase was suggested to be caused by compaction of the sediment (Laj et al., 1982; Langereis, 1984).

The inclination error of magnetic phases with near normal axial dipole field (NADF) directions allows them to be distinguished from secondary

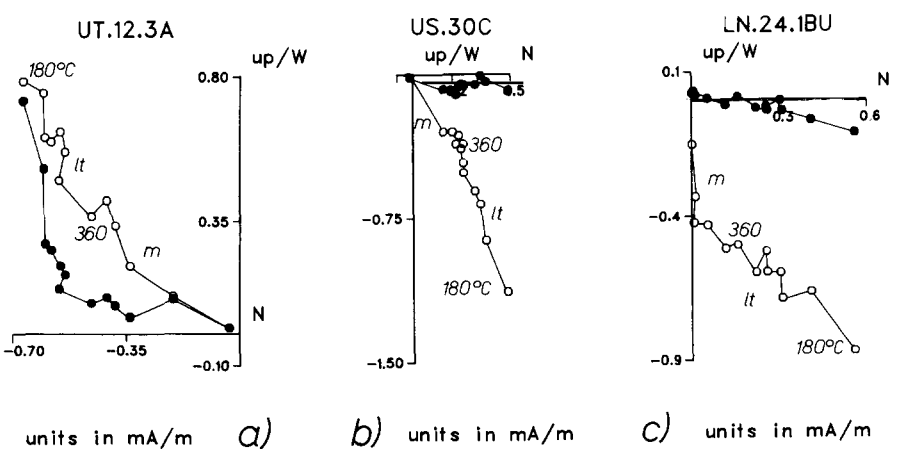

Fig. 3. Thermal demagnetization diagrams of specimens from the (a) Upper Thvera, (b) Upper Sidufjall and (c) Lower Nunivak transition records. The demagnetization data for temperatures below 180 °C were omitted. All diagrams show linear segments corresponding to the LT and the magnetite phase (denoted as 'lt' and 'm' respectively).

magnetizations. Secondary magnetizations cannot be distinguished from an LT phase with an NADF direction (without an inclination error). The observation that (1) the sedimentary interval which contains such phases is unaltered and rock-magnetically homogeneous and (2) a neighbouring rock-magnetically similar sedimentary interval contains LT phases that carry near-RADF directions demonstrates that the component with the NADF direction is a primary one.

Viscous components are easily recognized by the strong relation between their directions and the storage orientation of the specimens. The observed viscous components have relaxation times exceeding a few days. An example of a viscous component has been given in section 5.1. The thermal decay curves of the viscous component and the LT phase of the primary component have a significant overlap. Consequently, the relative magnitudes of the LT phase and the viscous component determine how well the direction of the magnetization of the LT phase can be determined. Fortunately, viscous component with high intensities and LT phases with very low intensities were never observed in the same specimens.

5.3. Directions of NRM components in the reversal records

Within the reversal records an anomalous direction of a remanent magnetization is defined as a part of a primary component when (1) its thermal decay is identical to the decay of the LT or the magnetite phase (see Fig. 2), (2) when it is found in other specimens from approximately the same stratigraphic level, or (3) it is part of a directional variation of identical phases as a function of stratigraphic height.

In the sedimentary intervals outside the reversal records, the magnetizations of the LT phase and the magnetite phase can have different directions. This characteristic is more extremely encountered in specimens from the reversal record; Fig. 3 shows three examples. The difference in direction between the LT and magnetite phase can be caused (1) by different lock-in depths of these phases or/and (2) by different lock-in intervals of these phases. The implication of the first possibility is a backtracking of the local geomagnetic field directions recorded by the two phases. In that case, there will be two identical records of the geomagnetic field behaviour during its reversal between which an apparent time-lag exists. The latter possibility that the two phases have different lock-in intervals will result in two records which are smoothed to different extents. Only the study of the directional and intensity variations with time (stratigraphic distance) of the LT and magnetite phase can indicate to what degree these possibilities reflect the real remanence acquisition processes.

6. Bulk-rock-magnetic investigation of the sediment

A detailed study has been made of the bulk-rock-magnetic properties of the Trubi sediment (Linssen, in preparation). The relevant information is given below.

Specimens of the four studied reversal records were subjected to bulk-rock-magnetic experiments so that the nature and the grain sizes of the LT and magnetite phases which carry the primary component could be examined. Rock-magnetic information about the minerals in the sediment contributes to the interpretation of accumulation rate variations and of the continuity of sedimentary deposition. It should also indicate whether it is feasible to relate remanent magnetizations and/or rock-magnetic parameters to paleointensities.

6.1. Continuous thermal demagnetization and low-temperature treatment

We subjected representative specimens from the sampled reversal records to successive thermal treatments in order to investigate alteration effects upon heating. Prior to each heating or cooling cycle the specimens were given an isothermal saturation remanence (ISR) in a 2 T dc field. The term ISR is used for convenience although it will be demonstrated in section 6.2 that the specimens are not completely saturated in 2 T.

The first heating cycle for specimen US.109.1B shows (Fig. 4a) that the ISR is dominated by two

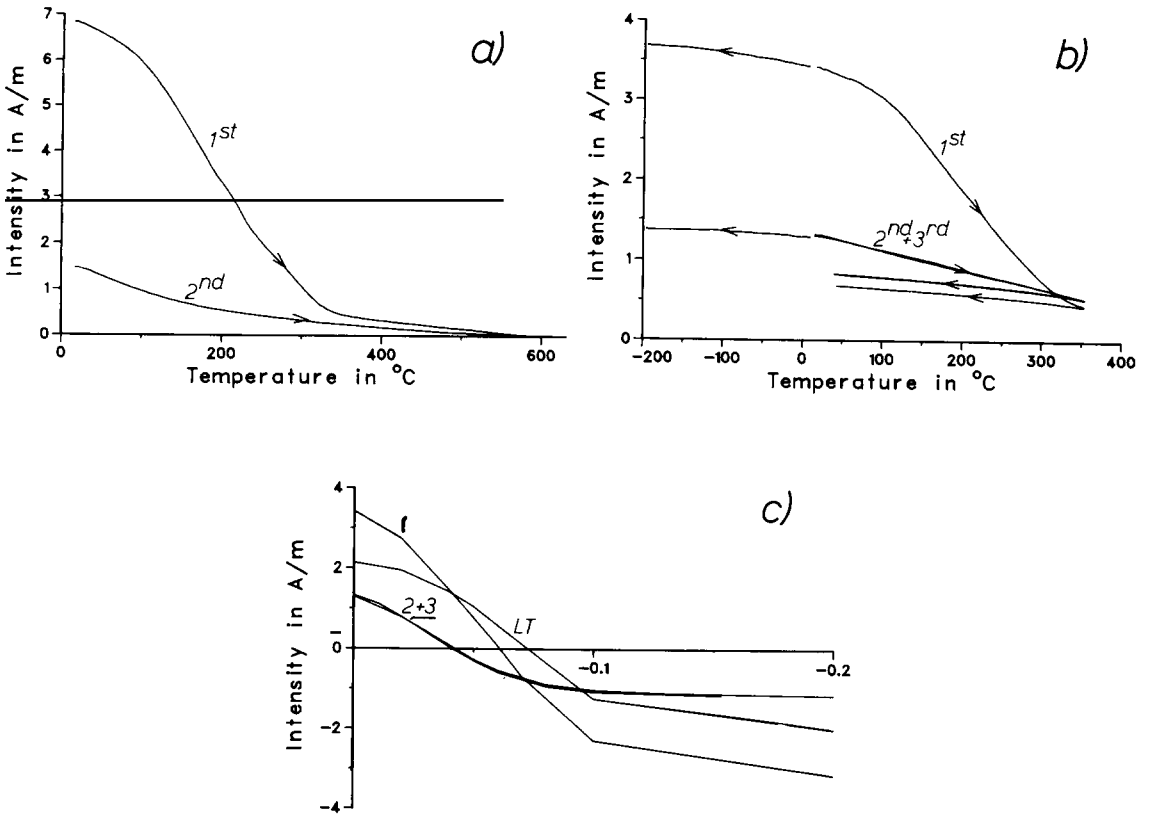

Fig. 4. (a, b) Continuous thermal treatments of specimens from the Upper Sidufjall reversal record. The specimen acquired an ISR in a 2 T field before each thermal treatment. The heating cycles are numbered. (c) Remanent hysteresis curves determined before each of the thermal treatments of (b). Magnetite and the LT phase contribute to curve 1, original and newly formed magnetite contribute to curve 2 and 3. The curve denoted 'LT' remains after subtraction of curve 1 and 2. It represents the contribution of newly formed magnetite and the LT phase. Abscissa: back field in Tesla.

magnetic phases the remanence of which is completely removed at 360 and 580 °C respectively. Since these breakdown temperatures are similar to those of the primary component of the NRM, one must conclude that they are carried by the same phases. During the second heating cycle applied to the same specimen (Fig. 4a), it is demonstrated that the LT phase is absent and that the ISR which is left after 360 °C has decreased. Obviously, during the first heating cycle the LT phase was physically removed. The decreased ISR left after 360 °C points to a partial alteration of magnetite. There is no evidence that new magnetic minerals have been produced as a result of the alterations. Figure 4b depicts the results of a similar experiment. However, in this case the specimen (US.113B), which acquired an ISR, is successively cooled to −190 °C, reheated to room temperature, regiven an ISR and heated to 360 °C. The latter heating step is assumed to cause a complete alteration of the LT phase on the basis of the results for the similar specimen US.109.1B. It should be noted that the remanence increased significantly with cooling. Afterwards, the same specimen is twice subjected to the identical experimental procedure. The results for these experiments confirm that the LT phase was altered during the first thermal treatment. The significant remanence increase that occurred upon cooling has now disappeared; apparently, this remanence increase is a characteristic of the LT phase. The second and the third heating curves are identical

and have a slightly higher ISR left at 360 °C than the first heating curve. Apparently, during the first heating cycle, a new magnetic mineral was produced. The fact that the second and the third heating curves are identical suggests that they represent true demagnetization behaviour and are not affected by alteration.

The remanent hysteresis curves of specimen US.113B determined prior to each heating step show (Fig. 4c) lower H_{cr} values after the first heating cycle. The first remanent hysteresis curve represents the LT and magnetite phase, the second curve represents the magnetite phase and the new magnetic phase that was produced during the first heating cycle to 360 °C. Subtraction of the second from the remanent hysteresis curve produces an estimate of the remanent hysteresis curve of the LT phase; it is an estimate since the curve is slightly biased by the contribution of the new magnetic material.

The initial susceptibilities (X_i) of other specimens, similar to specimen US.113B, show a slight (< 5%) increase during a stepwise thermal demagnetization treatment up to 300 °C. The X_i decreases significantly after heating the specimen to 360 °C (> 20%). Further heating causes the X_i to increase again to values that are 2–4 times above the X_i of the non-heated specimen.

Heating experiments with specimens from sediments with similar rock-magnetic contents (Linssen, 1984) suggested that, with increasing temperature, magnetite was produced from the non-magnetic matrix. Hysteresis experiments provided evidence that the grain size of the newly formed magnetite increased (due to growth) from superparamagnetic (SP) grain-size ranges, to the single-domain (SD) grain-size range and subsequently to the pseudo-single-domain (PSD) or even the multi-domain (MD) grain-size range. In these sediments the transition of the magnetite through the SD grain-size range always occurred at 450 °C and caused a maximum of ISR (which was 2–10 times greater than the ISR of the non-heated specimen). Heating the specimens to temperatures above 500 °C caused the old and new magnetite to alter to hematite.

In summary, it is concluded that the observed breakdown of the remanence carried by the LT phase (magnetite) is in the main (partly) due to alteration of these minerals rather than to their demagnetization. The LT phase alters completely below 360 °C; the magnetite alters partly between 360 and 600 °C. There is evidence that magnetite is produced below 360 °C and that some magnetite is altered after the specimens are heated to 600 °C. Low-temperature treatment shows (1) a significant remanence increase of the LT phase with cooling and (2) hardly any effects for the remanence of magnetite. The last observation indicates a small grain size (< 5 μm) for magnetite since the loss of remanence upon cooling to the isotropic point of magnetite decreases with grain size (Hartstra, 1982).

The LT phase in specimen US.113B has an H_{cr} value of ~ 72 mT; this value is slightly biased by the production of the newly formed fine-grained magnetite. Since fine-grained magnetite has an H_{cr} lower than 70 mT (Hartstra, 1982), the true H_{cr} value of the LT phase will probably be > 72 mT.

6.2. *ARM and ISR characteristics*

It was investigated whether anhysteretic remanent magnetization (ARM) and ISR characteristics allowed a more quantitative determination of the contribution of the LT and magnetite phase to the NRM. The specimens for the ISR and ARM experiments have been selected from the same sedimentary interval as those from the previous section. In Fig. 5a the thermal decay curves are depicted of (1) an ARM acquired in an alternating field (AF) of 100 mT, (2) an ARM acquired in an AF field of 350 mT, (3) an ISR acquired in a 2 T field, and (4) a primary component of an NRM. The ARM-inducing direct field was 0.036 mT. Since the previous sections provided evidence that the main magnetic phases are an LT and a magnetite phase, a bimodal decay is expected. The bimodality is evident in the thermal decay of the NRM and ISR. The thermal decay curves of ARM are only slightly bimodal. The decay curve of ARM acquired in 350 mT shows hardly a more pronounced bimodality than the decay curve of the ARM acquired in 100 mT. The thermal decay of the NRM is not identical to the thermal decay

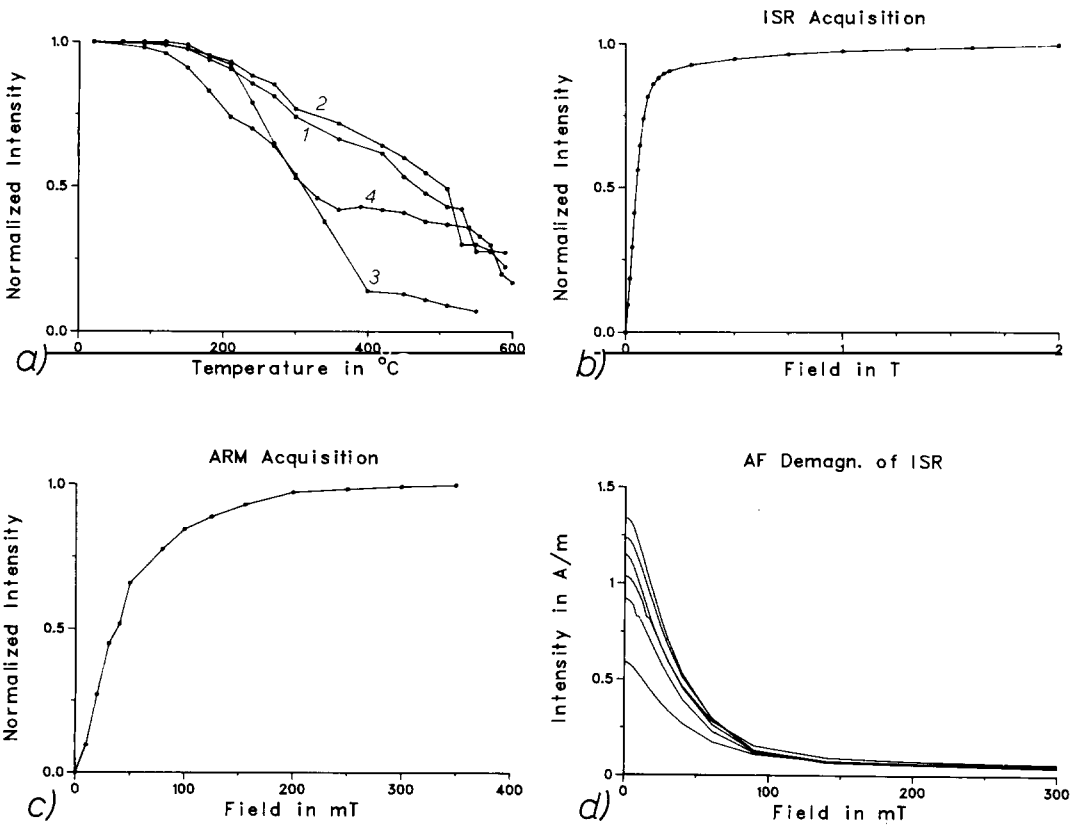

Fig. 5. (a) Thermal decay curves of specimens from the US reversal record with (1) ARM acquired in an AF field of 100 mT, (2) ARM acquired in an AF field of 350 mT, (3) ISR acquired in 2 T, (4) primary component of the NRM. (b) ARM acquisition curve. (c) ISR acquisition curve. (d) AF demagnetization curves of ISR. Continuous thermal treatment (with exact temperature readings) of specimens from the same sedimentary intervals as the specimens of (a)–(d) demonstrated that the magnetizations of all specimens are entirely removed at 600 °C.

of the ARM or the ISR. The thermal decay curves show that, with respect to the NRM, the LT phase acquires less ARM and more ISR.

The normalized ISR and ARM acquisition curves (Fig. 5b and c) demonstrate that specimens from this sediment are not completely saturated respectively in 2 T dc fields and 350 mT AF fields. The ISR and ARM acquisition curves show, in spite of the presence of the LT and the magnetite phase, no sign of bimodality.

In some sedimentary intervals the intensities of the primary component show a major variation. A closer examination demonstrates that this intensity variation often is caused by a change in rock-magnetic properties in either the magnetite or the LT phase. This is illustrated by the AF decay curves (Fig. 5d) of specimens from such a sedimentary interval. The specimens acquired an ISR in a 2 T field. The figure shows that the high coercive fractions have equal intensities, unlike the low coercive fractions. Thermal demagnetization of other specimens from the same sedimentary interval showed that an acquired ISR was entirely removed below 600 °C. These observations suggest that the observed NRM variation is due to rock-magnetic changes of magnetite and is not caused by the geomagnetic field. A minor contribution to the high-coercive fraction of small amounts of high-coercive minerals other than the LT phase (e.g., hematite) cannot be excluded.

In summary, the thermal demagnetization of ARM shows only a small contribution of the LT

phase, although this contribution increases slightly with the applied AF. The contribution of the LT phase to the ARM is lower than to the NRM, whereas its contribution to the ISR is higher than to the NRM.

The sediment under study, containing a LT and a magnetite phase, does not give bimodal ARM and ISR acquisition curves. The LT phase is not saturated in 2 T dc fields. Variations in the intensity of the magnetite correlate with variations in the low coercive fraction of the ISR, whereas they do not correlate with the high coercive fraction of the ISR.

6.3. *Determination of the LT phase*

The fact that the LT phase alters completely upon heating to 360 °C points to pyrrhotite (Schwarz, 1975) and maghemite (Dankers, 1978). In Fig. 6a the results of the AF demagnetization of the ARM of some natural pyrrhotites is shown (Dekkers, unpublished data). Figure 6b depicts the grain size variation of the Median Destructive Field of ARM ($H_{1/2A}$) for natural magnetite, maghemite and pyrrhotite. All minerals show increasing $H_{1/2A}$ values with decreasing grain sizes. The magnetite and maghemite behaviour is similar; the $H_{1/2A}$ increase for pyrrhotite is more significant. Considering the $H_{1/2A}$ data for maghemite it is expected that an ARM acquisition in an alternating field of 350 mT will saturate or nearly saturate small (say 1–5 μm) grains of this mineral. The thermal demagnetization data of ARM (Fig. 5a) indicate otherwise; a pronounced bimodal thermal decay would be expected if the LT phase is maghemite. With regard to Fig. 6b, the quoted thermal demagnetization data of ARM makes sense if the LT phase is pyrrhotite. The ISR results for pyrrhotite (Dekkers, 1988) resemble the experimental results presented in the previous sections (High H_{cr}, ISR not completely saturated at 2 T). Finally, in the studied specimens we observed a typical pyrrhotite characteristic: spurious remanences are produced when > 80% of the remanence of the pyrrhotite phase is demagnetized with alternating fields (Schwarz, 1975).

Given the present information, pyrrhotite is considered to be the most likely carrier of the LT phase. This mineral (most probably) is produced authigenically; chemical characteristics of seawater are such that it would oxidize detrital pyrrhotite. Post-sedimentary or early diagenetic processes in the presence of organic matter encourage the formation of sulphidic minerals, which indeed are abundant in the sediment (Linssen, 1984).

Therefore, it is tentatively concluded that the LT phase is pyrrhotite. Low-temperature characteristics and H_{cr} suggest that the grain size of

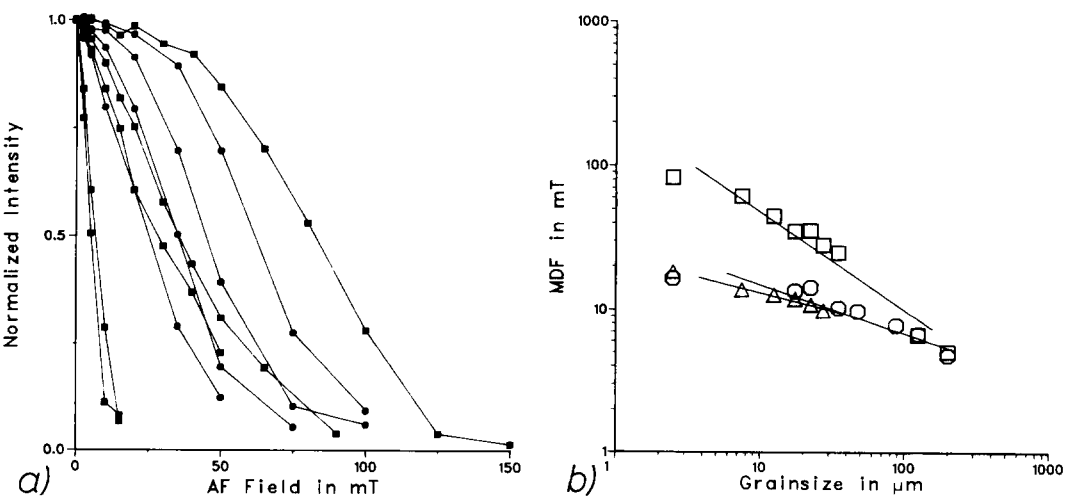

Fig. 6. (a) AF decay curves of ARM in synthetic specimens containing natural pyrrhotite of various grain size ranges. ● denotes EOR pyrrhotite, ■ denotes TTE pyrrhotite (see Dekkers, 1988). (b) Median destructive field of ARM. □ denotes natural pyrrhotite (TTE, EOR; see a), ○ denotes natural magnetite (HM4T; Hartstra, 1982), △ denotes natural maghemite (DKA, Dankers, 1978).

the pyrrhotite and magnetite in the sediment is < 5 μm.

6.4. Implications

King et al. (1983) developed a rock-magnetic test to discover whether magnetic remanence of a sediment can be related to paleointensity. Although this test is used by investigators working on sediments which are rock-magnetically similar to the sediment of the Trubi formation, the experiments suggest that its use is not really justified because in our view the first condition of this test, namely that the dominant magnetic mineral should be magnetite, is not satisfied. In the range 2–20 μm, the X_{ARM} of natural pyrrhotite tends to increase with grain size (Fig. 7a). A maximum may be present near 20–30 μm. This trend is opposite to the trend observed for magnetite (King et al., 1983). In Fig. 7b the susceptibility data for pyrrhotite (Dekkers, 1988) and the above ARM data are combined in an X_{ARM}–X plot according to King et al. (1983). When these data are compared with the magnetite data of King it becomes clear that the presence of pyrrhotite biases the grain size of magnetite interpreted from the X_{ARM}–X diagram to lower values. The apparent variations in the concentrations of magnetite that are interpreted from the X_{ARM}–X plot can easily be influenced by variations in pyrrhotite content of the sediment. The obvious way to solve this problem would be to determine X_{ARM} and X_i after the alteration of pyrrhotite upon heating the specimens to 360 °C. However, the experiments also provided evidence that small-grained magnetite has already been produced at this temperature. Kings' model may only be used when the contribution to X_{ARM} and X_i by the newly formed magnetite can be calculated or is proven to be minor. The correction for the presence of the newly formed small-grained (say 0.01–5 μm) magnetite is problematical since the grain-size variation of the ARM carried by this magnetite is large (King et al., 1983). The non-uniformity of mineral contents in this type of sediment prevents the successful use of standard tests for determining the paleointensities in sediments. Alternatively, to test whether one of the two phases is suitable for representing the ancient geomagnetic field intensity, one has to use rock-magnetic methods with which one can discriminate the two magnetic phases. Neither X nor ARM has such characteristics.

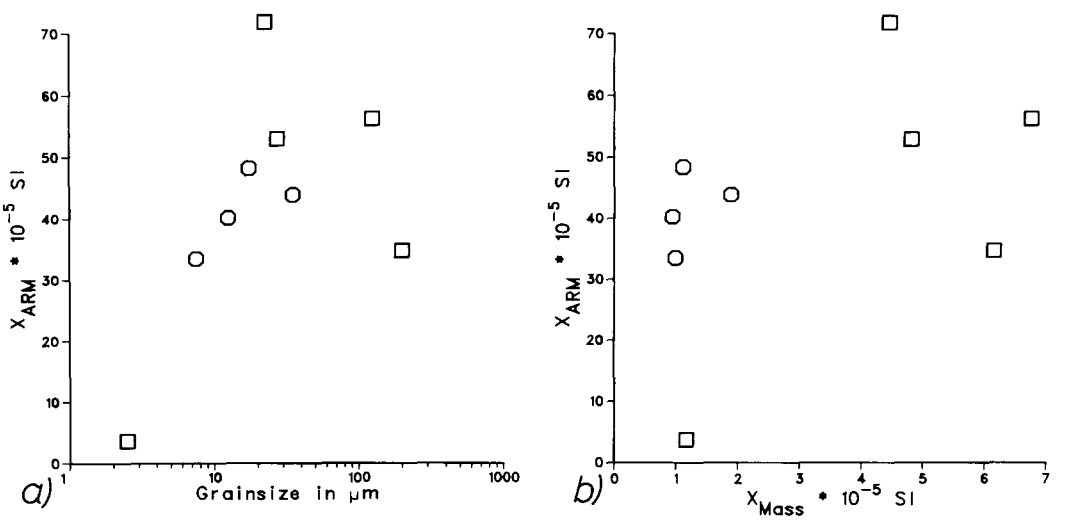

Fig. 7. (a) X_{ARM} of EOR (○) and TTE (□) natural pyrrhotites versus grain size. (b) X_{ARM} – X diagram according to King et al. (1983) of EOR (○) and TTE (□) natural pyrrhotites. For susceptibility data of pyrrhotite see Dekkers (1988).

7. Demagnetization procedures and analysis of demagnetization data

Viscous behaviour of thermally treated specimens, due to the production of small SD magnetite grains with relaxation times of < 100 s (section 5.3; Linssen, 1984), was successfully suppressed by heating specimens for no longer than the time necessary to create a homogeneous (< 5 °C) temperature distribution in the specimen and to cool the specimens rapidly with air. Heating times varied from 30 min at 100 °C to 20 min at 600 °C. A homogeneous temperature distribution in a specimen is required to prevent unwanted smoothing effects in the demagnetization results caused by the removal of remanent magnetizations with different directions and thermal decay. Heating experiments with thermocouples in dummy specimens demonstrated that a temperature inhomogeneity in the specimens, at all temperatures, is never in excess of 5 °C after 20 min heating time. Furthermore, the temperature of the specimens as a function of the heating time was determined. At heating times of 20–30 min, the results of the experiments showed a maximum bias of 0–20 °C to lower temperatures (relative to the temperature indication of the furnace).

Partial thermal remanent magnetization (PTRM) can be produced during cooling of the newly formed 'sensitive' small SD magnetite in a non-zero (< 25 nT) magnetic field. The recognition of such a PTRM is important in this study since we are looking for (small) remanences with anomalous (transitional) directions. This PTRM could be recognized and accounted for since the position of the specimens in the furnace was varied and registered at each heating step. The magnitude of a PTRM in a specimen was never in excess of 0.05 mA m^{-1}.

The large number of demagnetization data have been analysed with an in-house developed interactive computer program that makes use of graphical projection routines. In this program linear segments are 'least-squares fitted' to data points selected by the operator. The directions of these linear segments are employed to construct (Zijderveld, 1967) and calculate intensities. The interactive computer program displays demagnetization diagrams and decay curves and facilitates the calculation of the directions and intensities.

The minor but indefinite viscous and laboratory-induced remanences exceed the small measurement error of the cryogenic and spinner magnetometers. For this reason the demagnetization data were not weighted with the measurement error. The statistical error which was determined is the maximum angular deviation (MAD; Kirschvink, 1980).

8. Reversal records

The thermal demagnetization results are presented for the Upper Thvera (UT), Lower and Upper Sidufjall (LS and US) and the Lower Nunivak (LN) reversal records. The demagnetization results for the LS and LN records comprise the directions and intensities of the magnetite phase, whereas the results for the UT and the US records also include the demagnetization results for the LT phase. The LT phase records for the LS and LN reversals have not yet been determined. The AF results will not be discussed in this paper. One should note that the magnetite and the LT phase records have inclination errors relative to NADF or RADF directions (Section 5.2). The inclination error is about 10° for magnetite and probably less for the LT phase. The LT phase has a decay curve with a small temperature interval of non-overlap with either the viscous component and the magnetite phase. Relative to the magnetite results, a high statistical error is produced if the limited number of demagnetization data points are least-squares fitted.

For lack of a better method, the intensities of the magnetite and LT phase are compared with an ISR acquired in a 2 T field and besides, for the UT and US records, with the remanence which remains after a 300 mT AF demagnetization of the ISR (ISR_{AF}). The ISR and ISR_{AF} are considered to be related with the contribution to the primary component of respectively: LT phase + magnetite and LT phase (sections 5 and 6). The ISR_{AF} records must be considered with caution since locally minerals with high coercive forces (e.g., hematite) may contribute to the ISR_{AF}. One

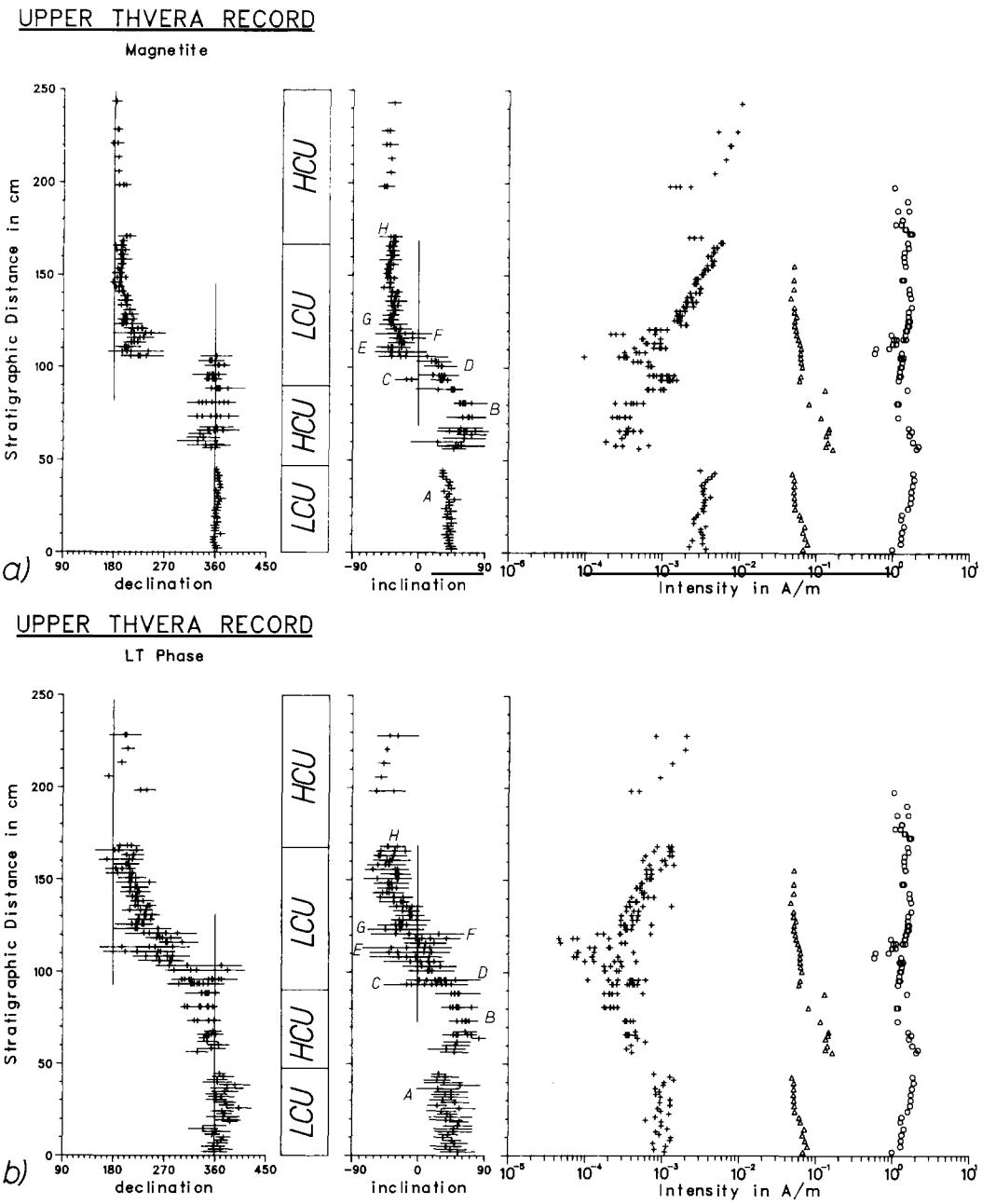

Fig. 8. Upper Thvera reversal records for (a) magnetite and (b) LT phase. The error bar represents the maximum angular deviation (Kirschvink, 1980). ○, denotes the ISR; △, denotes the ISR_{AF}. Characters refer to virtual geomagnetic poles in Fig. 12.

should also notice that the ISR fraction which remains after an AF demagnetization in a 300 mT field is small (< 10%) relative to the initial ISR.

The demagnetization results of specimens from sedimentary intervals near the boundaries between the HCUs and LCUs often do not answer to the

TABLE II

Durations (in ka) for observed changes (rows 1–4) in the studied reversal records

Reversal	Durations							
	UT		LS		US		LN	
	(a)	(b)	(a)	(b)	(a)	(b)	(a)	(b)
1. Period with anomalous intensities	26.0	26.0	1.3	1.9	5.0	7.7	2.1	4.0
2. Period with anomalous directions	17.0	17.0	2.5	3.8	10.0	15.4	4.1	8.0
3. Major declination change	1.0	1.0	0.6	1.0	0.5	0.8	0.3	0.7
4. Major inclination change	3.0	3.0	1.3	1.9	2.0	3.1	2.1	4.0

Columns (a) denote durations relative to the accumulation rates from this study. Columns (b) denote durations calculated with accumulation rates from Zijderveld et al. (1986).

criteria of section 5.3. For this reason they were omitted. The accumulation rate variation will be discussed in Section 10. Prior to this discussion the duration of the observed directional and intensity variations will be given in centimetres. Table II provides an overview.

8.1. Upper Thvera polarity transition

The length of the sampled and investigated sedimentary interval is 250 cm. It comprises two LCUs, one complete HCU and a lower part of another HCU. Figure 8a, b shows the thermal demagnetization results for the magnetite and the LT phases respectively. Transitional magnetite and LT phase magnetizations related to the Upper Thvera N–R reversal are observed in the lower HCU (43 cm) and in the upper LCU (78 cm). It should be noted that the directional changes in the LT phase record have higher amplitudes than those in the magnetite record. For both records similar directional changes seem to take place at identical sedimentary levels. Starting at the base of the lower LCU, the directional changes are as follows: the record starts with near-NADF directions showing a slight oscillation. At the base of the lowermost HCU the directional changes observed in the declination record intensify. They are accompanied by a gradual steepening of the inclination in the interval from 55 to 75 cm. Subsequently, the inclination decreases to near-zero values in the interval from 75 to 95 cm. Above this interval the inclination returns to near-NADF directions before it changes to negative values at the 106 cm level. Above the 106 cm interval the declination and inclination show a damped oscillation which almost disappears at the 150 cm level. Above the 150 cm level only near-RADF directions are observed.

The magnetite and LT phase intensity records show lowered values between the 50 and 170 cm levels. A minimum intensity is present at the 106 cm level in the magnetite record and at 118 cm in the LT phase record. Both minima do not correlate with the ISR minimum at 110 cm. The records demonstrate that the minimum intensity is ~ 5% of the maximum intensity. The LCU unit between 93 and 158 cm is characterized by fairly constant ISR_{AF} values. The experiments which underlie Fig. 5d (AF decay curves of ISR) are made with specimens from the interval near the ISR minimum at the 110 cm level. The constant ISR_{AF} values in this interval suggest that the minimum is caused by bulk-rock-magnetic changes (in concentrations and/or grain sizes) of magnetite. The intensity variation of the magnetite and LT phase in the lower HCU unit is somewhat obscured by bulk-rock-magnetic changes. Since the ISR_{AF} values in this interval are high relative to the ISR values, one is led to the assumption that this maximum is caused by bulk-rock-magnetic changes of the LT phase. This correlates with the observation that the intensity decrease in the lower HCU is more significant for the magnetite record than for the LT phase record.

8.2. Lower Sidufjall polarity transition

The sedimentary interval, containing the Lower Sidufjall R–N reversal record, has a length of 50

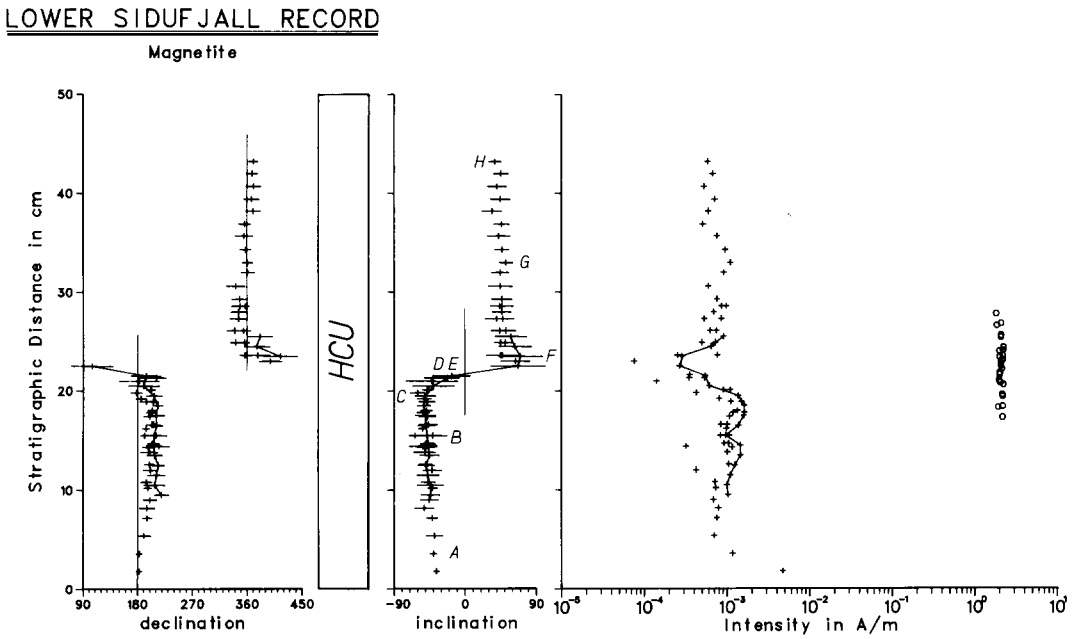

Fig. 9. Lower Sidufjall reversal record for the magnetite phase. See also caption to Fig. 8.

cm. The record comprises the lowermost part of an HCU (Fig. 9). A preceding investigation already ascertained the fact that the reversal is recorded in a small sedimentary interval. Therefore, this interval was sampled closely both with horizontally and vertically drilled cores. Figure 9 shows the thermal demagnetization results for magnetite. The demagnetization results for specimens (slices with a thickness of 0.9 cm) from the vertical core are connected by straight lines.

The reversal is recorded as follows. In the interval between the 10 and 18 cm levels the inclination steepens slightly (from $-40°$ to $-60°$) whereas the intensity increases ~ 20%. At the 18 cm level the intensity starts to decrease. From the 20 to the 23 cm level the intensity decrease is accompanied by a change to eastward and steeply downward directed magnetizations. From the 23 to the 27 cm interval the intensity increases whereas the declinations and inclinations turn to near-NADF directions. The intensity at the 10 and 27 cm level are nearly identical. The intensity minimum at the 23 cm level is 20% of the intensities at the 10 and 27 cm levels whereas it is 15% of the intensity at the 18 cm level. The ISR is constant in the interval from the 17 to the 28 cm level.

Note that the minimum in intensity coincides with steeply downward directed magnetizations.

8.3. *Upper Sidufjall polarity transition*

The investigated sedimentary interval has a length of 150 cm, containing an HCU (104 cm) and an overlying LCU (46 cm). The Upper Sidufjall N–R reversal is recorded in the HCU. The thermal demagnetization results of the magnetite and the LT phase are depicted in Fig. 10a, b respectively.

The reversal proceeds as follows. The record starts with near-NADF directions. After an initial increase, the inclinations shallow near the 20 cm level. Low inclinations are recorded in the interval between the 20 and 40 cm levels. The lowered inclination values are accompanied by an intensity increase.

Near the 40 cm level the intensities decrease. The decrease is sudden and sharp in the LT phase record whereas it is gradual in the magnetite

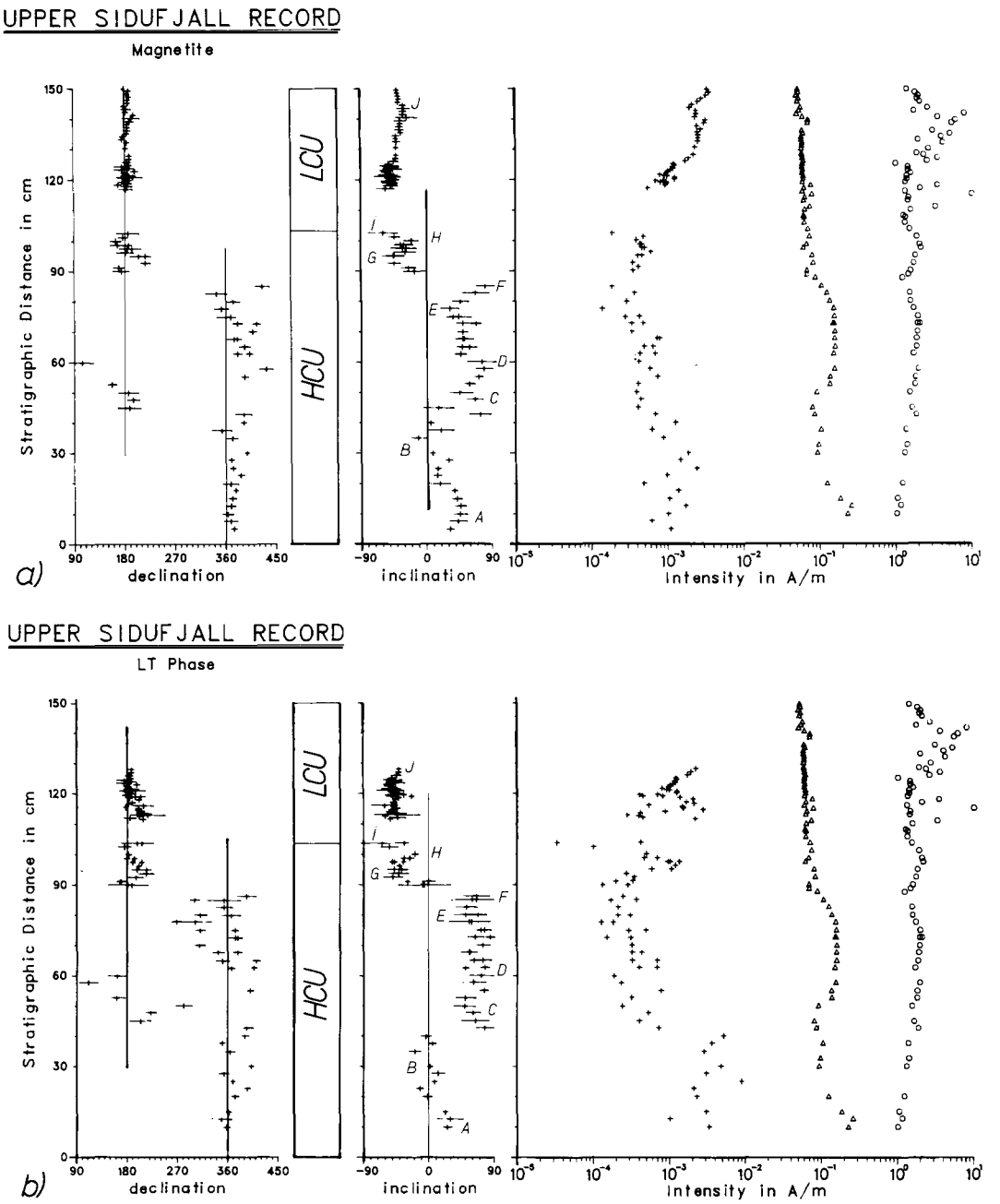

Fig. 10. Upper Sidufjall reversal records for the (a) magnetite and (b) LT phase. See also caption to Fig. 8.

record. At the level where the intensities start to decrease the magnetizations become more steeply downward directed. Again, the change is rapid in the LT phase record and gradual in the magnetite record. Up to the 44 cm level the declination values show an increasing rate of variation. At the 44 cm level the declinations swing to near-south (magnetite record) and south-west (LT phase record). The declination returns to north near the 60 cm level. The sense of movement of the declina-

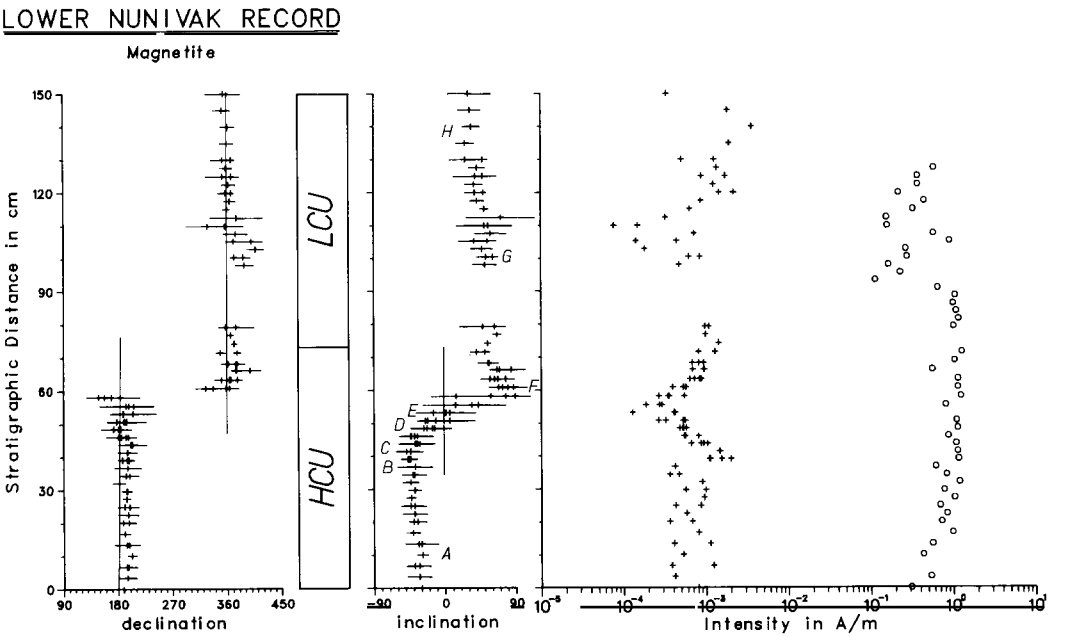

Fig. 11. Lower Nunivak reversal record for the magnetite phase. See also caption to Fig. 8.

tion near the 40 and 60 cm levels suggests that the first swing is eastward directed, whereas the return is westward directed. The interval between the 42 and the 87 cm levels is characterized by a low intensity, steeply downward directed magnetizations and a major variation of the declination. Concurrently with high inclinations, declinations are of course always poorly defined. In the interval between the 42 and 87 cm levels the variation of the inclination in the magnetite record is between 35 and 80°, whereas it is between 50 and 85° in the LT phase record.

Between the 87 and 93 cm levels the inclinations and declinations change to near-RADF directions in both the magnetite and the LT phase records. However, in the interval from the 93 to the 100 cm level the inclination again shallows. In the LT phase record the shallowing of inclination is accompanied by an intensity high. The intensity high is hardly observed in the magnetite record. At the 100 cm level the magnetization becomes steeply upward directed. This change is more rapid in the LT phase record than it is in the magnetite record. In the LT phase record the steepening of inclination is accompanied by an intensity low.

Above the hiatus in the record of demagnetization results the directions are close to the RADF, whereas the intensities tend to increase. The intensity records of the magnetite and the LT phase in the HCU are accompanied by constant or gradually changing ISR and ISR_{AF} records. The ISR_{AF} record of the LCU shows constant values, whereas the ISR record of the LCU shows strong variations ($> 10\times$). The ISR variations at the top of the LCU somewhat obscure the observed intensity increase in that part of the magnetite record. However, the nearly constant ISR_{AF} values of the LCU and the intensity increase in the LT phase record suggest that the increased intensities in the magnetite record are not (only) caused by bulk-rock-magnetic changes.

8.4. Lower Nunivak polarity transition

The sedimentary interval which was investigated has a length of 150 cm. It contains an HCU (73 cm) and an LCU (87 cm). The Lower Nunivak R–N reversal is recorded in the HCU. Figure 11 shows the thermal demagnetization results for the magnetite phase.

The directional and intensity changes which accompany the reversal are as follows. From the base of the record, where near-RADF directions are observed, the magnetization becomes gradually more steeply upward directed (up to 50°). At the 42 cm level the inclination starts shallowing. Near the 53 cm level the inclination is zero. Between the 53 and 60 cm intervals the magnetization becomes progressively more steeply downward directed. At the 60 cm level the mean inclination value is > 70°. Above the 60 cm level the inclination shallows to near-NADF directions. Near the 60 cm level the declination record shows only one sudden change with no intermediate directions. The level at which this change occurs nearly coincides with the level at which the inclination becomes steeply downward directed. The intensity record shows a V-shaped intensity low.

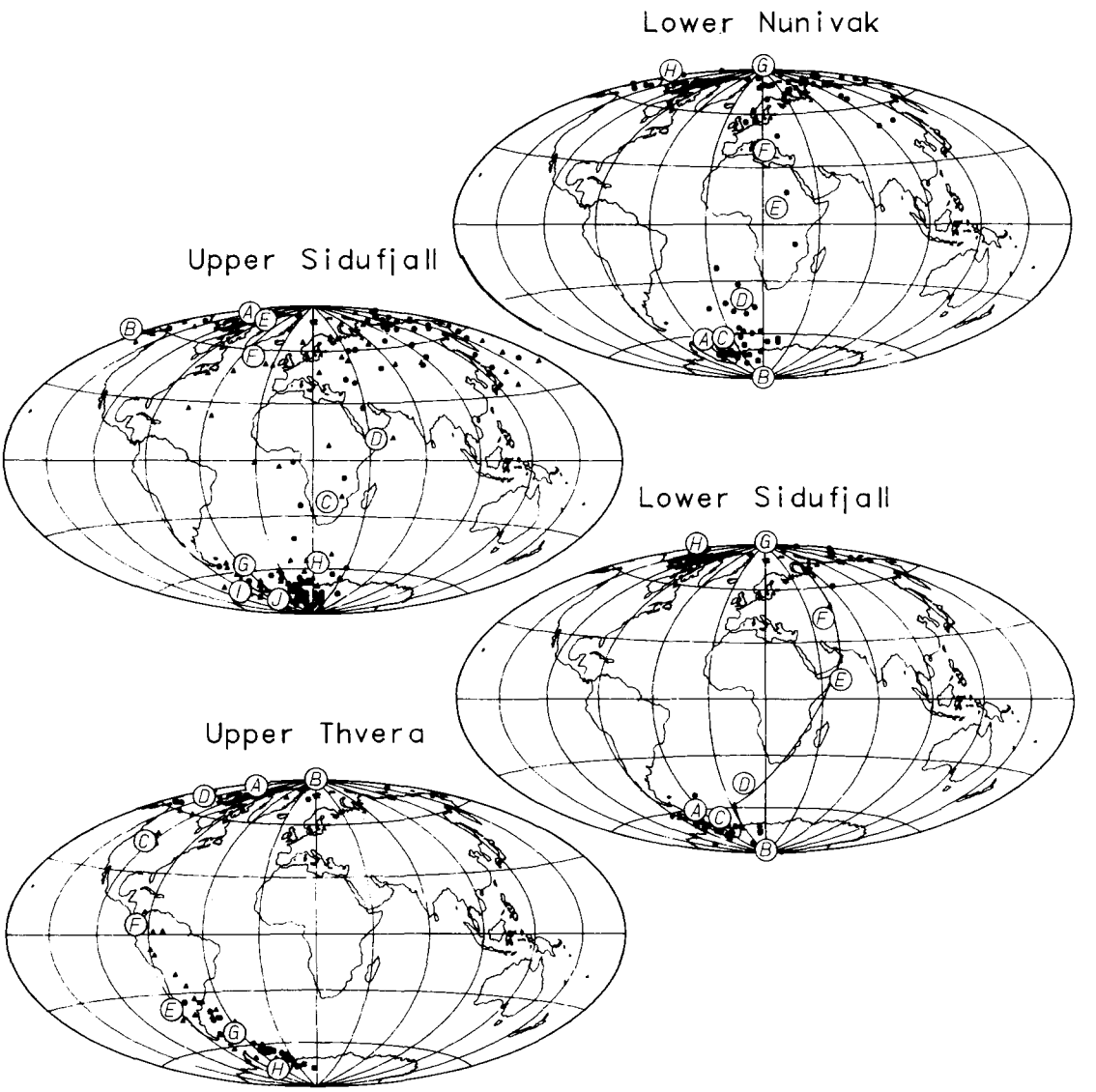

Fig. 12. Virtual geomagnetic poles of the reversal records in Aitoff projections of the Earth centred around the site meridian. Characters refer to corresponding labelled directions in the reversal records (Figs. 8–11).

The intensity minimum is near the 56 cm level. The nearly constant ISR record from the 40 to the 80 cm level suggests that the intensity low is not caused by bulk-rock-magnetic changes. The minimum intensity is ~ 10% of the intensity prior to and after the reversal.

9. Virtual geomagnetic poles

Figure 12 shows the virtual geomagnetic poles (VGPs) for the demagnetization results from section 8. One should note that the results have an inclination error. As a result of this error the VGPs of the near-RADF or near-NADF directions are respectively near- and far-sided. The corresponding characters in the projections of the VGPs and in the records of demagnetization results relate specific directional features.

Respectively, the VGPs of the Upper Thvera record move: from a far-sided location at 80° N (A), to the geographic pole (B), to a far-sided location at 70° N (C) and back to the far-sided location at 80° N (D). The westward declinations between the directional features (D) and (E) cause the VGPs to be confined to a great circle ~ 90° west of the site meridian. The oscillation (E–F–G) proceeds along the same great circle. The directions become stable at a near-sided location at 70° S (H).

The VGPs of the Lower Sidufjall and the Lower Nunivak record move: from a near-sided location at 70° S (A), to the geographic pole (B), to a near-sided location at 70° S (C), along a path 30° E of the site meridian (LS) or along the site meridian (LN) (D, E), to a location at the latitude of the site (F), to the geographic north pole (G) and at last to a far-sided location at 80° N (H).

The VGPs of the Upper Sidufjall move: from a far-sided location at 80° N (A), to a far-sided location at 60° N (B), near a location on the site meridian at 30° S (C), to a location near the equator (D), back to a far-sided location at 80° N (E), to a near-sided location at the latitude of the site (F), to a near-sided location at 60° S (G), to a near-sided location on the site meridian at 60° S (H), to a location near the antipodal site (I), and at last to a near-sided location at 70° S (H).

10. Discussion

Sedimentary reversal records are formed during a period of high geomagnetic field variation per unit time. Reports on depositional realignment (Tucker, 1980) and smoothing (Hoffman and Slade, 1986) provide valuable information about the acquisition of remanence in sediments. The synthetic records in these reports have been actually observed. For instance, Tauxe and Badgley (1984) found four magnetic phases in red beds each with a different timing of remanence acquisition giving delayed transitional paths for these phases.

In the present study we observed a different behaviour for the two magnetic phases that comprise the primary component (sections 5.2, 5.3, 8.1 and 8.3). Different directions are observed during periods of stable polarity and of polarity reversal. In view of the directional records of magnetite and the LT phase of the same reversal in which corresponding directional changes occur at the same sedimentary level, one is led to the conclusion that these phases do not produce significantly delayed records. However, the same records demonstrate convincingly that the directional variations of the LT phase exceed those of the magnetite phase. It is therefore tentatively concluded that the remanences of the LT phase produce a less-smoothed record of geomagnetic field behaviour than the remanences of the magnetite phases.

Each of the sampled polarity records resides for the greater part in a specific HCU or LCU. In order to make a more realistic estimate of the duration of the sampled polarity transitions one needs to have knowledge of the accumulation rates of the HCUs and LCUs separately. The accumulation rates were estimated by comparing the accumulated thicknesses of the HCUs and LCUs (provided by Zijderveld and corrected by the author for the exact location of each reversal record) with the known durations of the adjacent polarity intervals. In this section, the mean accu-

mulation rate for a polarity interval gradually decreases with increasing age. One can calculate mean accumulation rates for the HCUs and LCUs in adjacent polarity intervals if the assumption is made that the accumulation rates of the HCUs and of the LCUs are constant (two equations and two unknowns). If this method is performed on all neighbouring polarity intervals, the result is a moving average of the accumulation rate. Minor variations in the accumulation rate will be smoothed in the moving averages of the accumulation rate. However, an overall trend in the accumulation rate will be unaffected. The method has not been used with the Sidufjall subchron and the overlying reversed polarity interval since the mean accumulation rates in these polarity intervals are too distinct to yield realistic estimates.

The results are depicted in Fig. 13. It appears that the increase in the accumulation rate from 4.9 to 7.6 cm ka^{-1} is due to an increase in carbonate influx. This caused the accumulation rate of the HCU to dominate the increase of the accumulation rate of the LCU by a factor of three during the time interval between the UT and the LN reversal. The accumulation rates of the sediment containing the reversal records were obtained from a least-squares fit of the calculated HCU and LCU accumulation rates; although these values will probably not be absolutely accurate they are thought to be more realistic than those corresponding to the duration of the polarity intervals and total (HCU + LCU) sediment thicknesses.

This approach is supported by the fact that the lengths of the quite similar Lower Sidufjall and the Lower Nunivak reversal records, which are both recorded in a HCU, are different by a factor of three. The durations of these reversal records are identical if the above-described method is used to calculate the accumulation rates.

The durations of the directional and intensity changes related to the reversals records (section 8) are produced in Table II. The durations which were calculated from the accumulation rates of Zijderveld et al. (1986) as well as the durations calculated from the above-described method are given. It should be noted that the duration of the major directional changes that give rise to a change in magnetic polarity take place in very short time

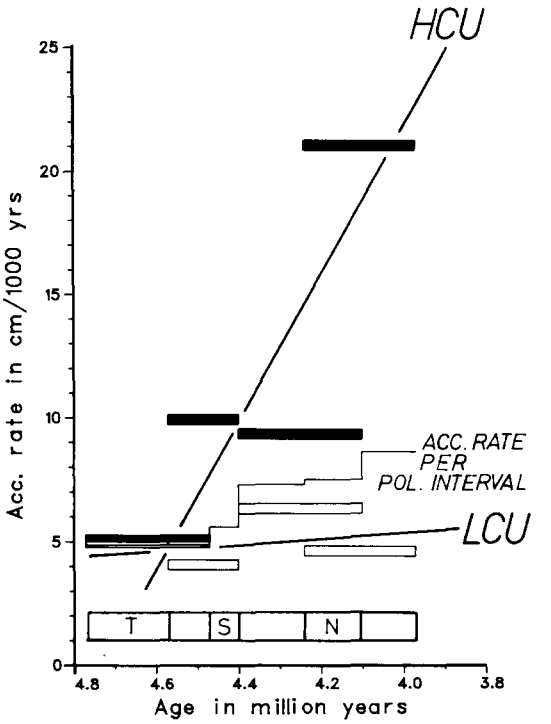

Fig. 13. Accumulation rates for six successive polarity subzones from the Gilbert Chron (thin solid line). The normal polarity intervals are indicated (in the inset) as T (Thvera), S (Sidufjall) and N (Nunivak). The calculated mean accumulation rates for the HCU (closed boxes) and the LCU (open boxes) in two successive polarity intervals have been least-squares fitted (thick solid lines).

intervals. This time interval is shortest for the change in declination for the three youngest reversals.

A comparison of the four studied reversal records demonstrates the high degree of similarity between the two R–N reversals records (LS and LN). The points of differences between the LS and LN reversal are the moment at which the intensity minimum is observed and the sense of movement in the declination records. The LS record suggests an intensity low at the moment when the inclination is nearly vertical, whereas the LN suggests an intensity low at the moment when the inclination is near-zero. It is hoped that further investigation (e.g., production of LT phase records and ISR_{AF} records) will provide information that indicates whether this effect is real or is

produced during the remanence acquisition. The declination record of the LS reversal suggests an eastward sense of movement (rapid and late in the record) whereas no intermediate declinations are observed in the LN record.

The UT reversal record shows a number of directional oscillations. During these oscillations there is an intensity low. A second short-period intensity low is observed at the level where the inclination changes sign. It is not clear whether this second intensity low is real or is caused by rock-magnetic changes.

The US reversal record seems to be most complicated. The record shows a characteristic phenomenon: a shallowing of the inclination accompanied by an intensity high and a subsequent steepening of the inclination accompanied by an intensity low. This feature is also observed above the major directional change (actual reversal?). There seems to be some kind of symmetry between these directional and intensity changes. When the dipole field is absent for a long period of time, as seems to be the case in the Upper Sidufjall transition, the record of the (steeply downwards directed) remaining field suggests a westward trend.

The projections of the VGPs calculated from the intermediate directions demonstrate that all directional changes in the reversals, including the rebounds, are confined to a path along a specific great circle. For the UT reversal this path is along a great circle ~ 90° W of the site meridian implying a non-zonal transitional field. The LS reversal is zonal with a minor contribution of non-zonal terms. The US reversal shows major directional changes in the vertical plane but the directional changes in the horizontal plane are either near-south or near-north; it indicates a mainly zonal transitional field behaviour. The LN reversal is clearly confined (within 30°) to the site meridian and again suggests the presence of a zonal transitional field geometry. One is led to the conclusion that there is a trend in the dominance of the zonal field. The zonal field seems to be less significant in the UT record whereas it is most dominant in the LN record.

The rebounds from the UT and US reversals are quite similar to the 'stop and go' movements observed in the Steens Mountain record (Mankinen et al., 1985; Prevot et al., 1985).

The phenomenological 'standing field' model of Hillhouse and Cox (1976) and the 'flooding' model of Hoffman (1977, 1979) provide transitional field geometries. The standing field model is based on the assumption that the dipole field decays and is subsequently rebuilt in the opposite sense while a standing field remains. The standing field may be zonal and/or non-zonal. In the flooding model it is assumed that the reversal process is initiated at a localized point in the fluid core and then floods through the rest of the core. The flooding model assumes zonal transitional fields although moderate non-zonal terms can also be incorporated. The standing field model predicts identical VGP paths for successive polarity transitions from the same site. Identical VGP paths for successive reversal records in the flooding model imply that, (1) in the case of an octupolar transitional field, the reversal process in one of the two identical transitions is simultaneously initiated in different hemispheres of the core or, (2) in the case of a quadrupolar transitional field the reversal process is initiated in one region but in different hemispheres of the core.

Regarding the sequence of four reversal records as a whole, neither the flooding model nor the standing field model easily fits the results. The flooding model does not fit the results since the UT reversal seems to be dominated by non-axisymmetric terms. The standing field model only fits the results if it can be assumed that the standing field changed from zonal to non-zonal between the Upper Thvera and the Lower Sidufjall.

The descriptive zonal harmonic model of the Matuyama–Brunhes (R–N) reversal provided by Williams and Fuller (1981) fits the LS and LN (R–N) reversal records. For the latitude of Calabria the model predicts (1) a V-shaped intensity low which is ~ 10% of the original field and (2) high inclinations late in the reversal (after the inclination changed sign). Clement and Kent (1984, 1985, 1986) report records of the Upper Olduvai (N–R) and Lower Jaramillo (R–N) reversals. The records are from sedimentary deep-sea cores from southern and northern latitudes. They

suggested that during these reversals a standing field was present with a harmonic content which was slightly different compared with the Matuyama–Brunhes reversal. Actually, the energy distribution to the different zonal terms was identical, they merely changed the signs of the zonal G_2^0 and G_4^0 terms. No reversal of the zonal G_3^0 term was observed. In doing so, the inclination record for different hemispheres mirrors with respect to the horizontal plane. The observation in the present study of a set of older reversals with a harmonic content nearly identical to the Matuyama–Brunhes reversal leads to the speculation that one or more of the low-order even zonal terms reverses polarity on a time-scale larger than that of the dipole field. The above-quoted reversal records span a period of ~ 3 Ma. It should be noted that the zonal harmonic contents of the Upper Olduvai and Lower Jaramillo reversal records provided by Theyer et al. (1985) and Herrero-Bervera and Theyer (1986) are different from the records provided by Clement and Kent. Within the context of the zonal harmonic approach there is no explanation for these differences.

It is expected (Bogue and Coe, 1984) that a zonal reversal process which is dominated by quadrupolar terms gives records in which the intensity decrease is strongly latitude dependent (near 45% of the original field at 0° latitude and near 35% at 35°: the latitude of Calabria). The relation between latitude and paleointensity in a reversal is nearly absent when octupolar terms dominate the reversal process (near 20% of the original field intensity for all latitudes).

The observed intensity minima in the studied reversal records range from 5 to 15% of the intensities that are observed at the onset of the reversal. If we consider that the relative intensity decreases are related to the paleointensities, then the records argue for a dominating presence of octupolar or even higher-order terms.

Bogue and Coe (1982, 1984) report results of a study of two successive (R–N–R) reversal records in lavas from Kauai, Hawaii, with a K–Ar age of 5.6–4.5 Ma for the R–N transition. This age corresponds either to the succession Lower Thvera, Upper Thvera transitions or to the succession formed by the last normal subchron of Chron 5. Since the K–Ar dating method may easily yield errors of 0.1 Ma for these ages (Mankinen and Dalrymple, 1979), the Kauai results may also correspond to the Lower and Upper Sidufjall successions.

The Kauai results strongly suggest near-sided axisymmetric geomagnetic field behaviour during both polarity transitions. The Calabrian Upper Thvera reversal is non-axisymmetric and consequently does not fit the Kauai results. The Calabrian Lower Sidufjall reversal has a slight offset from an axisymmetric transitional field, whereas the Upper Sidufjall reversal is, on the whole, axisymmetric. On the base of this information, the Kauai records might fit the Calabrian LS and US records. However, the paleointensity study for the Kauai R–N reversal, which provides evidence for a two-stage intensity decrease, does not fit the Calabrian LS results.

Since the studied reversal records appear to have intrinsic features which are unique, it seems worthwhile considering whether they can be used as dating tools for this region. If more remote records of these reversals become available and if they can be generalized, it might be possible to use reversal records as dating tools on a world-wide scale.

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