

## The upper and lower Nunivak sedimentary geomagnetic transitional records from southern Sicily

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### ABSTRACT

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The detailed paleomagnetic records of the upper and lower Nunivak polarity transitions have been determined from Pliocene marine marls in southern Sicily. Magnetites are the most important carrier of the remanence in both records. The transitions are recorded in two components; a low temperature and a high temperature component. The two components do not represent the geomagnetic field because the changes in these components take place at lithological boundaries. In addition, the directional changes do not completely match directional changes of the same transitions recorded in Calabria, some 250 km away. The character of the virtual geomagnetic poles paths is probably caused by smoothing of the stable directions before and after the transitions. The directional changes as well as the smoothing mechanism can be explained by the diagenetic magnetite formation model (Van Hoof et al., 1992) in which shortly after burial, the remanence carried by newly formed secondary magnetites is superposed on the initial remanence carried by primary magnetite.

### 1. Introduction

The records of polarity reversals in sediments have recently led some authors to conclude that the paths of the virtual geomagnetic poles (VGPs) during transitions follow great circles that are preponderantly located over the Americas or over its antipode (Clement, 1991; Tric et al., 1991). Laj et al. (1991) related this to regions of higher seismic velocities in the lower mantle and to the pattern of fluid motion in the outer core. Hoffman and Slade (1986) and Rochette (1990), however, have earlier cautioned that sedimentary records may contain artefacts. Indeed, Langereis et al. (1992) indicate that the confinement of the VGP paths can also be explained by the process

of smoothing the stable non-antipodal directions before and after the transitions. In addition, a record of the natural remanent magnetization (NRM) may be much more complex than a straightforward registration of the geomagnetic signal. Some reversal records have shown that geomagnetic variations may be recorded by different magnetic components with varying lock-in depths (Dijksman, 1977; Channel et al., 1982; Van Hoof and Langereis, 1991). This indicates that the different components at the same stratigraphic level do not acquire their remanences at the same time. Dijksman (1977) and Channel et al. (1982) found the two components residing in magnetite and haematite, while Van Hoof et al. (1992) found these components both residing in magnetite. The varying lock-in depth of magnetite, the process of smoothing and the excursions that often precede or follow polarity transitions can be explained by the diagenetic mag-

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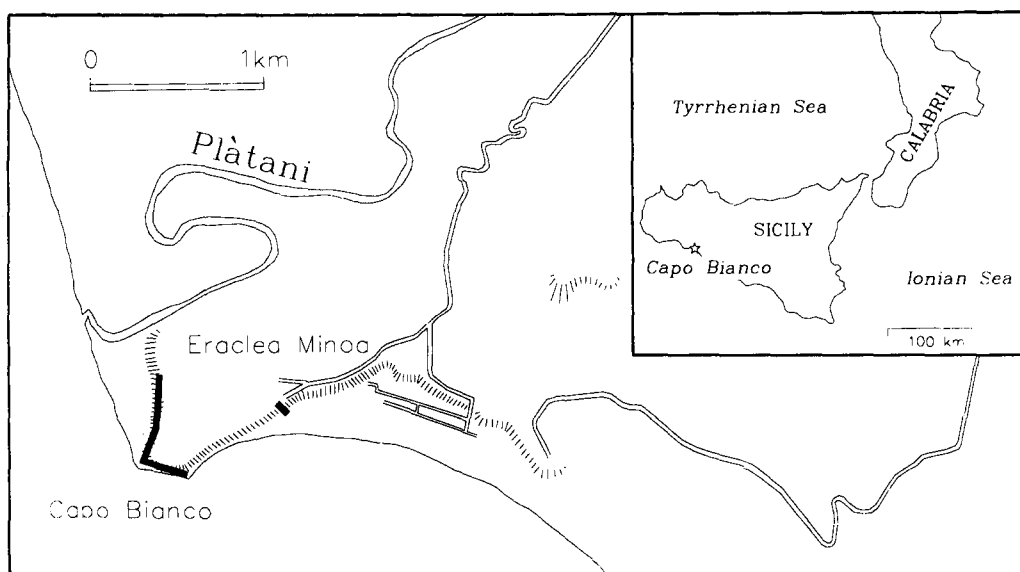


Fig. 1. Location of the Capo Bianco section in Sicily (Italy). Inset: Map of Sicily and Calabria. The transitions of this paper were sampled at the Capo Bianco section on Sicily and will be compared with the same transitions sampled by Linssen (1991) in Calabria, some 250 km away from Eraclea Minoa.

netite formation model (Van Hoof et al., 1992). This geochemically constrained model describes the formation of 'secondary' magnetites under suboxic conditions after burial of an organic-rich grey layer, and its remanence is superposed on the remanence of the 'primary' magnetite, acquired during deposition of the sediment.

Van Hoof and Langereis (1991) reported the varying lock-in depth in two sedimentary registrations of geomagnetic transitions from reversed to normal (R–N) polarity. One of those transitions was the lower Nunivak (LN) transition and only the declination and inclination records were shown. In this paper, we present the complete paleomagnetic record from the LN transition,

together with the data from the subsequent N–R Upper Nunivak (UN) transition. The results will be compared to the study of the records of the same transitions reported by Linssen (1991) sampled in Calabria, some 250 km away from the sampling places in southern Sicily.

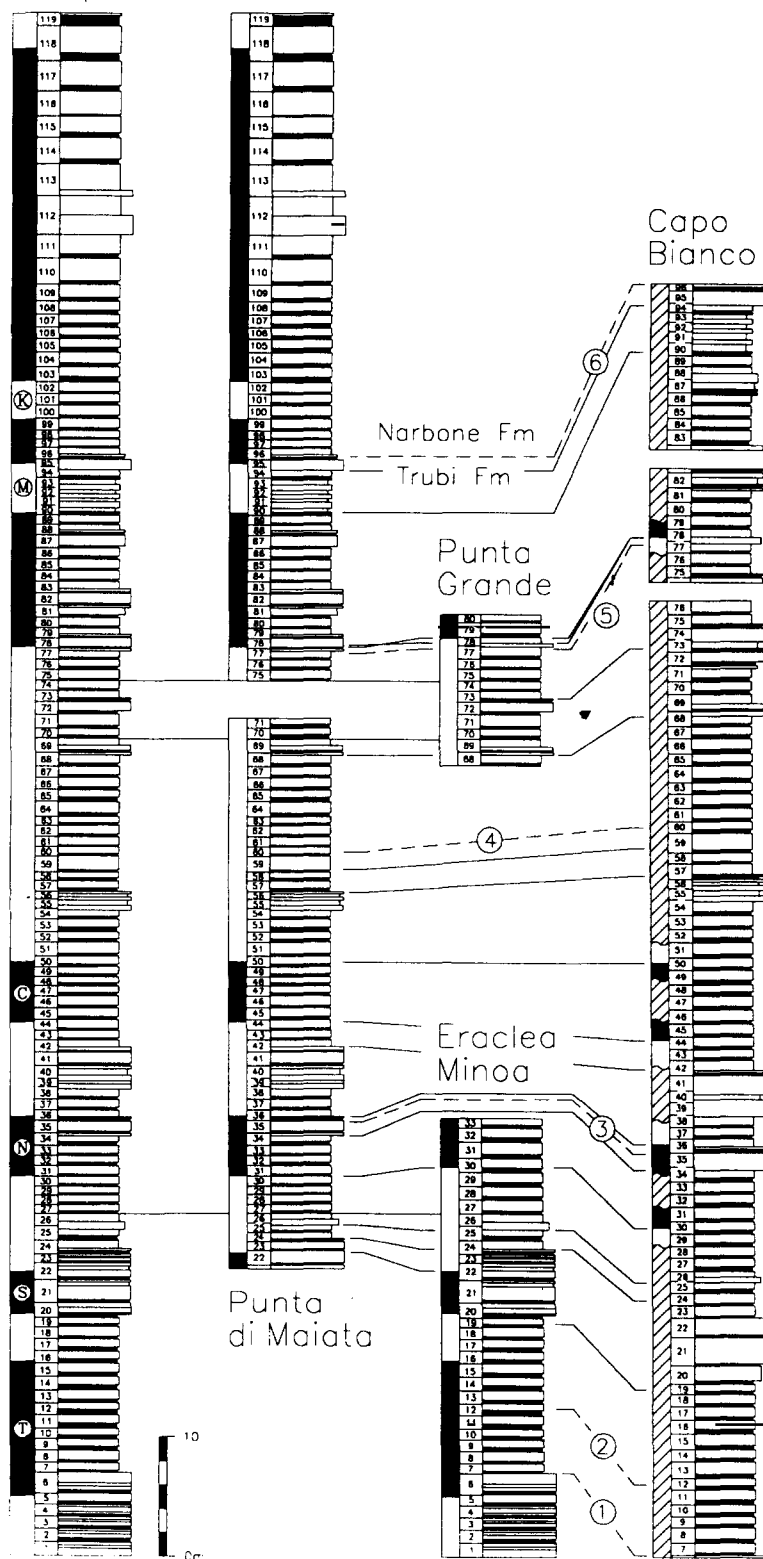
## 2. Geological setting and sampling

The Nunivak subchronozone in the Gilbert Chronozone was identified by a detailed magnetostratigraphic study of the Eraclea Minoa section and the Punta Maiata section in the Caltanissetta basin of southern Sicily (Fig. 1). These

Fig. 2. Lithostratigraphy and magnetostratigraphy of the Capo Bianco and parallel sections that form the Rossello composite section (Langereis and Hilgen, 1991). Figures in circles refer to occurrences of microfauna. The subdivision of small scale cycles is after Hilgen (1987). The weathering colours (carbonate contents) of each sedimentary cycle are grey (70%), white (80%), beige (60%) and white (80%), whereas fresh colours merely show gradual changes from dark-blue to light-blue. The lower Nunivak transition is determined in small scale cycle 30 and was sampled in detail over an interval of 1.2 m. The upper Nunivak transition is determined in small scale cycle 30 and was sampled over 1 m. In this paper the zero levels of the two transition records are arbitrarily chosen at pronounced layer parallel sedimentary lines. For the lower Nunivak this zero level was chosen at a clear brown layer in the beige layer of cycle 30. The zero level of the upper Nunivak record is the bottom of the beige layer of cycle 36. The white marls have a higher carbonate content, while the grey and beige marls are relatively poor in carbonate (Hilgen and Langereis, 1989).

Rossello  
composite

Punta  
Piccola



two sections contain the basal part of the Rossello composite section (Hilgen and Langereis, 1988; Langereis and Hilgen, 1990). The average sedimentation rate per polarity zone can accurately be determined and is  $4.5 \text{ cm kA}^{-1}$ . The lithology consists of marine marls of the Pliocene Trubi formation (Fig. 2). This formation is composed of sedimentary cycles, consisting of carbonates (60–

80%  $\text{CaCO}_3$ ) and a mixture of clay minerals. Hilgen and Langereis (1989) recognized a long succession of these small scale sedimentary cycles, so-called quadruplets. The succession starts with cycle 1 at the Mio/Pliocene boundary; the top is at cycle 119 with an age of 2.589 Ma. Based on the pattern of the quadruplets and the earlier magnetostratigraphic results, Hilgen correlated

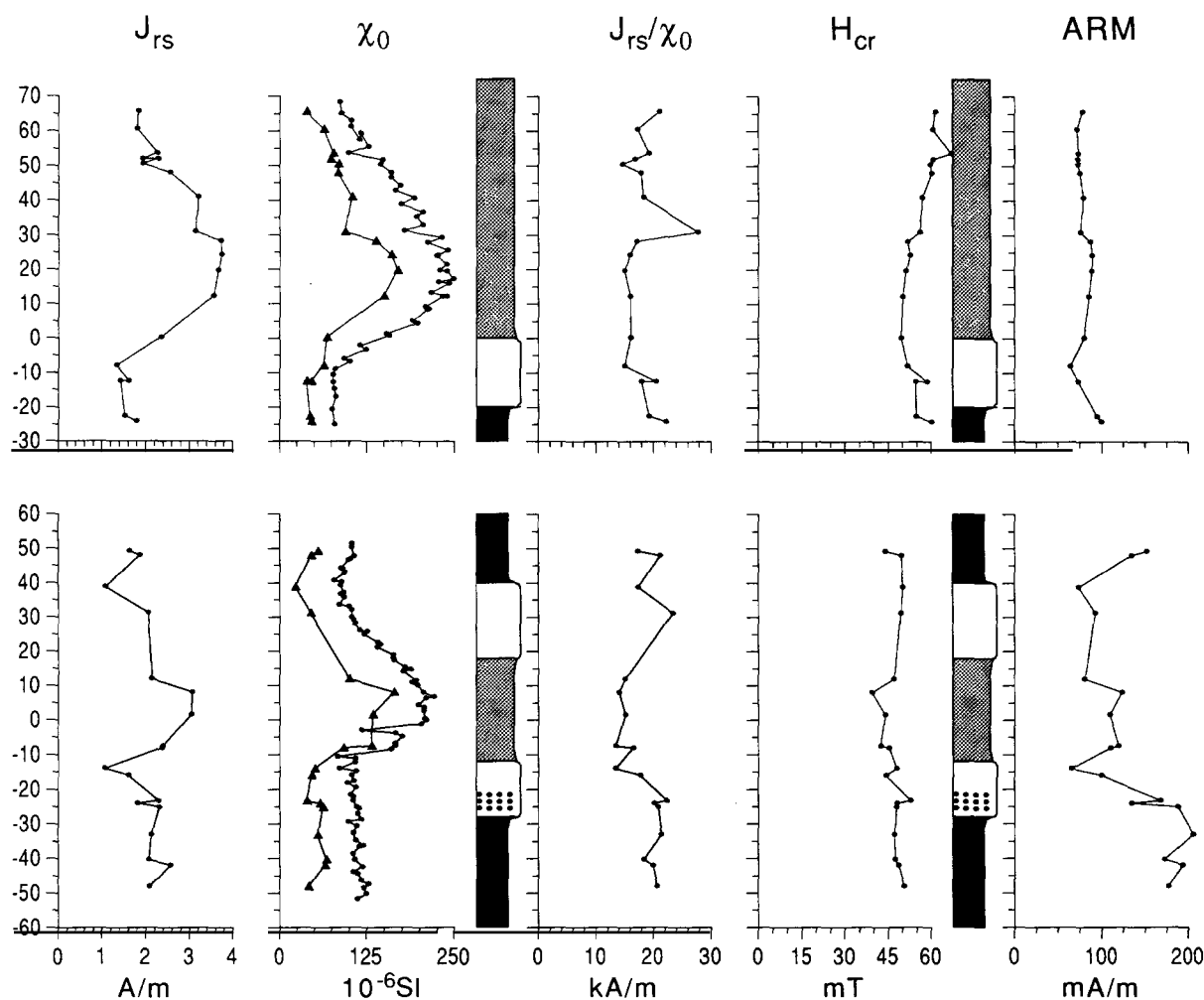


Fig. 3. Variation of magnetic parameters saturation IRM ( $J_{rs}$ ), initial susceptibility ( $\chi_0$ ), the ratio  $J_{rs}/\chi_0$ , remanent coercive force ( $H_{cr}$ ) and Anhysteretic Remanent Magnetization (ARM) as functions of the stratigraphical levels and lithology. The upper part of the figure is the upper Nunivak section; the lower is the lower Nunivak section. Black: grey coloured layer, white: white layer, grey: beige layer. Dots in the lower white layer of the lower Nunivak section refer to brown oxidation spots in the fresh blue marl. Triangles in the  $\chi_0$  diagrams are the corrections for the high field susceptibility (Fig. 4).  $J_{rs}$  and  $\chi_0$  have maxima in the beige parts of the lithology. The ratio of these parameters ( $J_{rs}/\chi_0$ ) lies in the range of fine-grained magnetite. After correction for the high field susceptibility these ratios will increase some 20%, to a range of even finer grained magnetites. The remanent coercive forces  $H_{cr}$  also lie in the range of fine grained magnetite. The dependency on lithology of the ARM is totally different from  $J_{rs}$  and  $\chi_0$ .

the Capo Bianco section near Eraclea Minoa to the basal part of the Rossello composite section, and he indicated the cycles (numbers 30 and 36) in which the LN and UN reversal boundaries were to be expected in the Capo Bianco section. The bedding plane of the Capo Bianco section has a strike and a dip of  $289.5^\circ$  and  $25.4^\circ$  N. The colour layering (Fig. 2) and the cyclicity have been described earlier (Hilgen and Langereis, 1989). The reversal records were sampled with a detail of a few millimetres. Considerable effort was taken to remove the weathered surface (up to 1 m) in order to expose the fresh (blue coloured) sediment. This method proved successful for almost all the sampled intervals, except for a 10 cm thick part of the lower white layer in quadruplet 30 where some brown spots were persistent in the fresh blue sediment. In the lithological column of the LN section this part is shown with black spots (Fig. 3).

### 3. Rock magnetism

The magnetostratigraphy of the Rossello composite section is based on the thermal demagnetization of a magnetite or high temperature (HT)

component removed mainly between 500 and  $600^\circ\text{C}$  (cf. Hilgen and Langereis, 1988). A low temperature (LT) component, demagnetized between 330 and  $500^\circ\text{C}$ , shows the same direction as the HT component well outside a transitional interval. However, near a transition, considerable differences in the directions of the LT and HT component occur that cannot be simply reconciled with the idea of a secondary or viscous overprint (Van Hoof and Langereis, 1991). Therefore, the detailed study of polarity transitions in these marine sediments requires more than an interpretation of the HT component only, because Van Hoof and Langereis (1991) have recognized a depth lag of the HT component relative to the LT component. This depth lag is not constant and, moreover, the LT component has acquired its remanence before the HT component. Rock magnetic investigations are needed to determine in some detail the magnetomineralogy. We have determined the initial susceptibility ( $\chi_0$ ) and the (remanent) saturation magnetization ( $J_{rs}$ ,  $J_s$ ), the (remanent) coercive force ( $H_{cr}$ ,  $H_c$ ) and the anhysteretic remanent magnetization (ARM).

The saturation isothermal remanent magnetization ( $J_{rs}$ ) acquired in a maximum DC field of 2

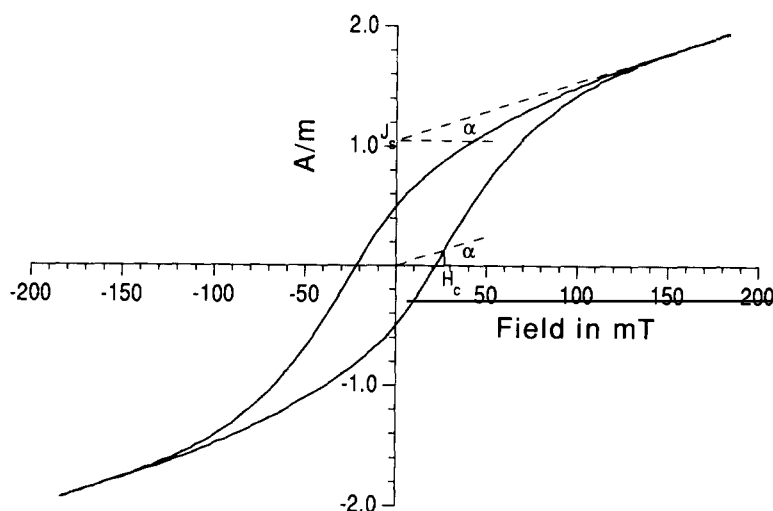


Fig. 4. Hysteresis loop. Horizontal (vertical) axis: inducing field (induced remanence). The linear trend at higher fields (dashed lines) results from the paramagnetic susceptibility of the clay minerals. It can be derived from the curve as the tangent (of  $\alpha$ ).  $J_s$ : saturation remanence determined where the dashed line crosses the vertical axis;  $H_c$ : coercive force determined where the loop crosses the horizontal axis after correction for the paramagnetic susceptibility.

T shows prominent maxima in the beige layers (Fig. 3(a)). Minima are usually found in the white layers just above the grey layers. The lithological dependence of the  $\chi_0$  (Fig. 3(b)) is in the UN section similar to that of  $J_{rs}$ , with a maximum of  $250 \times 10^{-6}$  (SI units) in the beige layer. In the LN section, the maximum of  $\chi_0$  in the beige layer is more pronounced than the maximum of  $J_{rs}$ . Because the clay fraction of the lithology may give a substantial paramagnetic contribution, it will increase the bulk susceptibility  $\chi_0$ . We have used hysteresis loop experiments to quantify the amount of the paramagnetic contribution (Fig. 4). Correction for the paramagnetic susceptibility decreases the  $\chi_0$  over both entire intervals (triangles in Fig. 3(b)). The maxima in the beige layers,

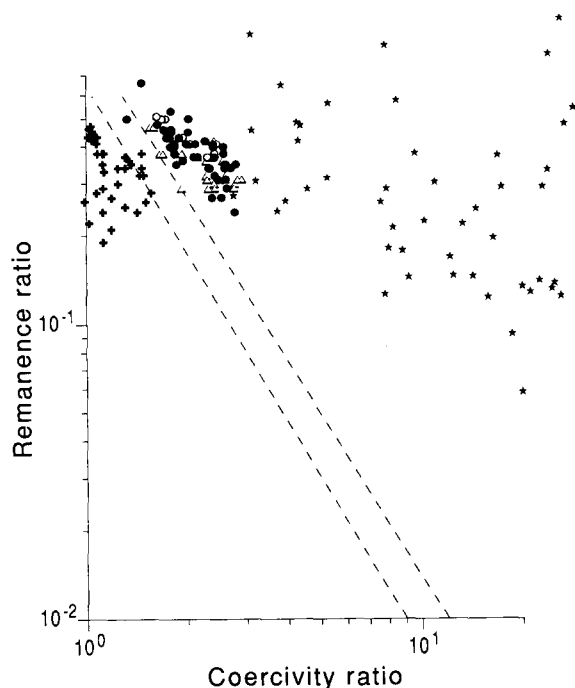


Fig. 5. Double logarithmic plot of coercivity ratio  $H_{cr}/H_c$  vs. remanence ratio  $J_{rs}/J_s$  after Van Velzen and Zijdeveld (1990). Literature data for magnetite of known grain sizes fall on a single trend indicated by two dashed lines. Stars (crosses) denote data from goethite (pyrrhotite) (Dekkers, 1988). The data from this study (closed circles) fit the values from the Trubi sediments by Van Velzen and Zijdeveld (1990) (open circles and triangles) very well. These authors concluded that fine grained magnetites were the most important magnetic minerals in the sediment.

however, are clearly retained and indicate an increase not only resulting from the clay fraction but also because of ferrimagnetic minerals. Both  $\chi_0$  and  $J_{rs}$  are dependent on the nature and concentration of the magnetic minerals. The ratio  $J_{rs}/\chi_0$  — largely independent of concentration provided that the dominant magnetic mineral is magnetite as is the case here (Van Velzen and Zijdeveld, 1990; Van Hoof et al., 1992) — may give an indication of grain size changes. The ratio shows typical values of about  $20 \text{ kA m}^{-1}$  (Fig. 3(c)). Fine grained magnetites have values larger than  $20 \text{ kA m}^{-1}$  up to  $30 \text{ kA m}^{-1}$ , but when corrected for the clay susceptibility the ratio will increase by some 50 %.

The remanent coercive force  $H_{cr}$  is independent of the concentration of magnetic material and is not influenced by paramagnetic clay minerals. Typical  $H_{cr}$  values of fine grained magnetites are 40–60 mT (Day et al., 1977; Hartstra, 1982; Dunlop, 1986) which are the generally rather constant values observed in Fig. 3(d).

In their rock magnetic study of the Trubi marls of Eraclea Minoa, Van Velzen and Zijdeveld (1990; 1992) determined the ratios  $H_{cr}/H_c$  and  $J_{rs}/J_s$  and concluded that the magnetic minerals were dominated by fine grained magnetites. A small discrepancy between their data and those from Dunlop (1986) was explained by the presence of goethite and some superparamagnetic material. The values of  $H_{cr}/H_c$  and  $J_{rs}/J_s$  fit the values from Van Velzen and Zijdeveld (1990) extremely well (Fig. 5).

In those parts of the record where the LT and HT components have completely different directions, the two components can easily be separated, resulting in two separate decay curves (Fig. 6(b)). The decay of the NRM of the other samples was not separated into LT and HT portions, but these curves also suggest the presence of an LT component by the two-fold decay (Fig. 6(b)). To investigate the presence of an LT component in the entire section we determined the ARM (Fig. 3(e)) as the remanence acquired in a 0.037 mT DC field superimposed on an AF field of maximum 250 mT. Subsequently, we thermally demagnetized the ARM from samples that were taken from the same cores as the ones used for

the NRM decay curves. These samples originated from grey and white (respectively levels  $-40$  and  $-25$  cm, Fig. 6) and from the interface between beige and white (level  $-12$  cm, Fig. 6) in the LN section. In addition, the susceptibility  $\chi$  during thermal demagnetization was monitored. The relative temperature dependence of  $\chi$  as well as ARM was identical for the three levels. The ARM decreases with a plateau in the decay curve between  $300$  and  $400^\circ\text{C}$ , just like the NRM decay curve. The  $\chi$  increases from  $350$  to a maximum at  $450$ – $500^\circ\text{C}$ , followed by a relative minimum at  $550^\circ\text{C}$  and a strong increase at higher temperatures. The maximum at  $450$ – $500^\circ\text{C}$  is an indication of the presence of iron-bearing sulphur minerals (Dekkers, 1990). The final increase is caused by the change of fine grained magnetites into super-paramagnetic magnetites (Van Velzen and Zijdeveld, 1992). Van Velzen and Zijdeveld (1990) found that there was no indication for any sulphur-bearing magnetic minerals (i.e. greigite or pyrrhotite) in the Trubi sediments.

#### 4. NRM components

Thermal demagnetization diagrams of the NRM generally show three prominent features: a viscous or secondary magnetization which is demagnetized below  $200$ – $250^\circ\text{C}$ , a LT component removed between  $200$ – $250^\circ\text{C}$  and  $480$ – $510^\circ\text{C}$ , and a HT component with unblocking temperatures between  $480$ – $510^\circ\text{C}$  and  $600^\circ\text{C}$ . Both components are carried by magnetites (this study; Van Hoof et al., 1992; Van Velzen and Zijdeveld, 1992). The LT and HT components are most clearly seen in the diagrams where they are not parallel (Figs. 7(b), 7(c), (d), (h) and (i)) and the two-fold decay of the two NRM components is depicted in Fig. 6(b). During demagnetization of the LT component, the direction of the remanence in some samples start fluctuating at temperatures of  $330$ – $390^\circ\text{C}$  (Fig. 7(h)). At this trajectory the susceptibility starts increasing (see rock magnetic section) so the fluctuations are most likely caused by oxidation of an iron-sulphur (pyrite) mineral and

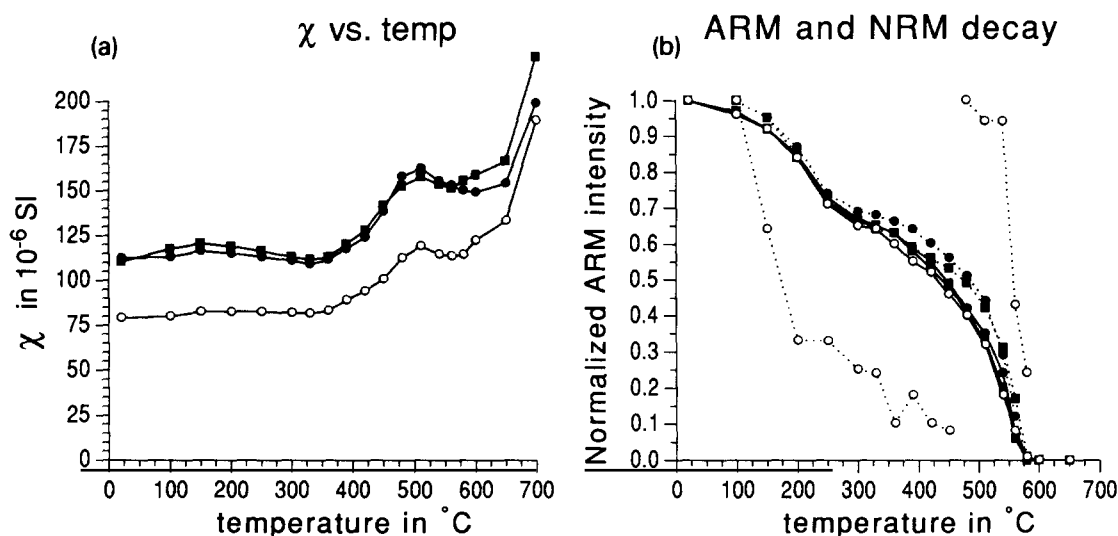


Fig. 6. Temperature dependency of  $\chi$ , ARM and NRM. Black dots and squares are from levels  $-40$  and  $-25$  cm, respectively, from the lower Nunivak section where the HT and LT components were parallel. Open circles show the measurements from a sample from level  $-12$  where LT and HT were anti-parallel. (a) Changes in  $\chi$  are identical in the three samples. Increase at  $450^\circ\text{C}$  indicates the presence of sulphur bearing minerals (Dekkers). Increase at  $650^\circ\text{C}$  indicates the generation of super-paramagnetic magnetites (Van Velzen and Zijdeveld, 1992). (b) Dotted line: NRM decay. Continuous line: ARM decay. The NRM of the samples from level  $-12$  was easily separated into two anti-parallel components. Decay of ARM and NRM are identical for all levels.

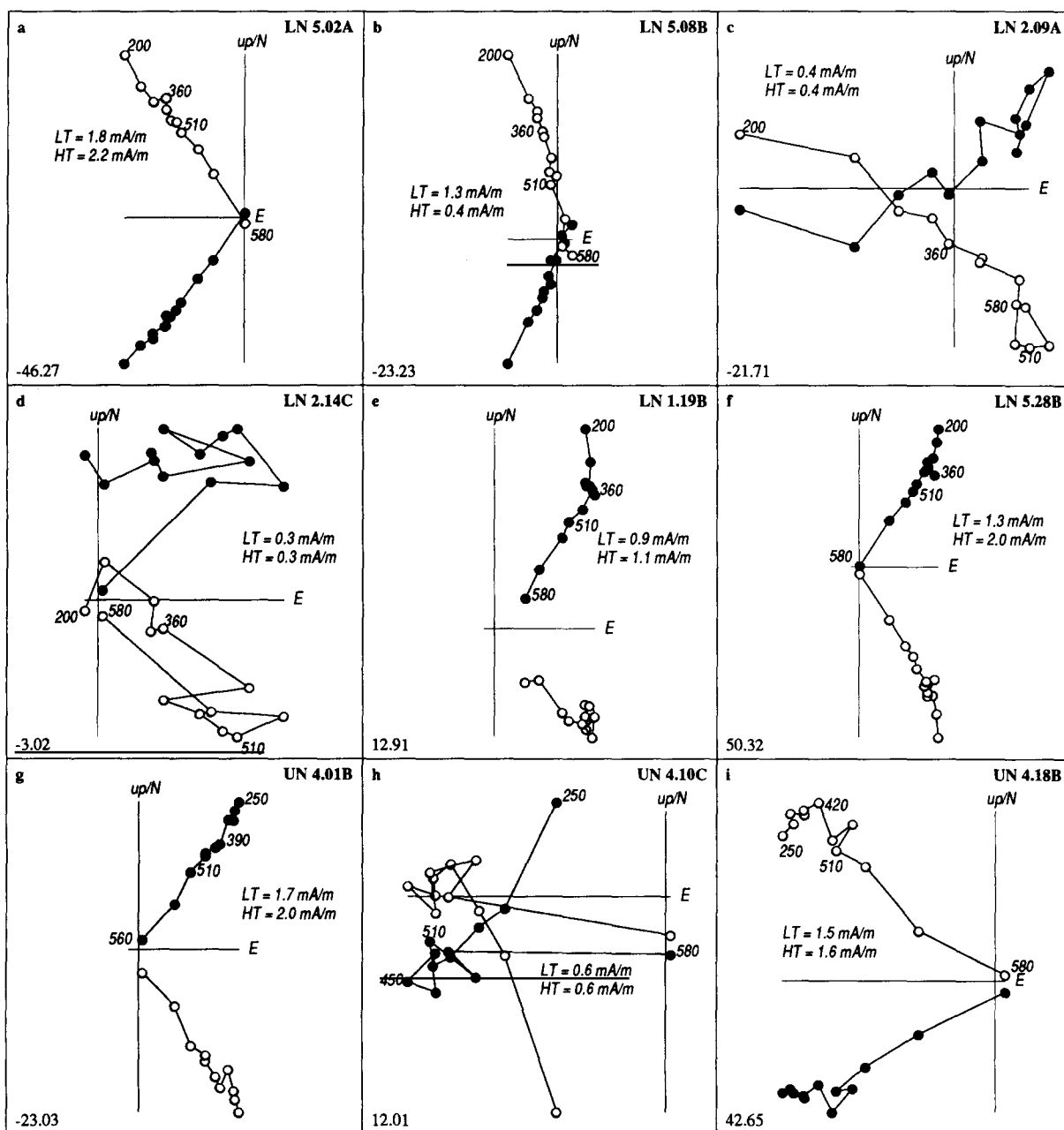


Fig. 7 Representative thermal demagnetization diagrams of the lower (LN samples (a)–(f)) and upper (UN samples (g)–(i)) Nunivak reversals. Stratigraphical level (down left) refers to the stratigraphic columns of Fig. 4. Solid (open) symbols are horizontal (vertical) projections. Temperature steps below 200°C are not shown to enhance the detail at the higher temperatures. Intensities of LT and HT are given in each diagram. UN: (a) the HT and LT components are both clearly reversed and include the 35° rotation; (b) the same, but at the highest temperatures (> 540°C) there is a tendency to normal directions and the intensity of the HT component decreases significantly; (c) the HT component has a normal polarity, while the LT component is still clearly reversed; both components have a low intensity; (d) the HT component is normal, and the LT component shows an intermediate direction: W and up; (e) the LT component has a northerly direction, but the inclination is still very shallow; (f) finally, both components have approximately the same normal polarity direction, including the familiar 35° rotation, and the intensities have largely recovered to pre-reversal values. LN; (g) both components are normally directed; (h) intermediate directions in LT and HT components; (i) reversed HT component and component could not be determined.



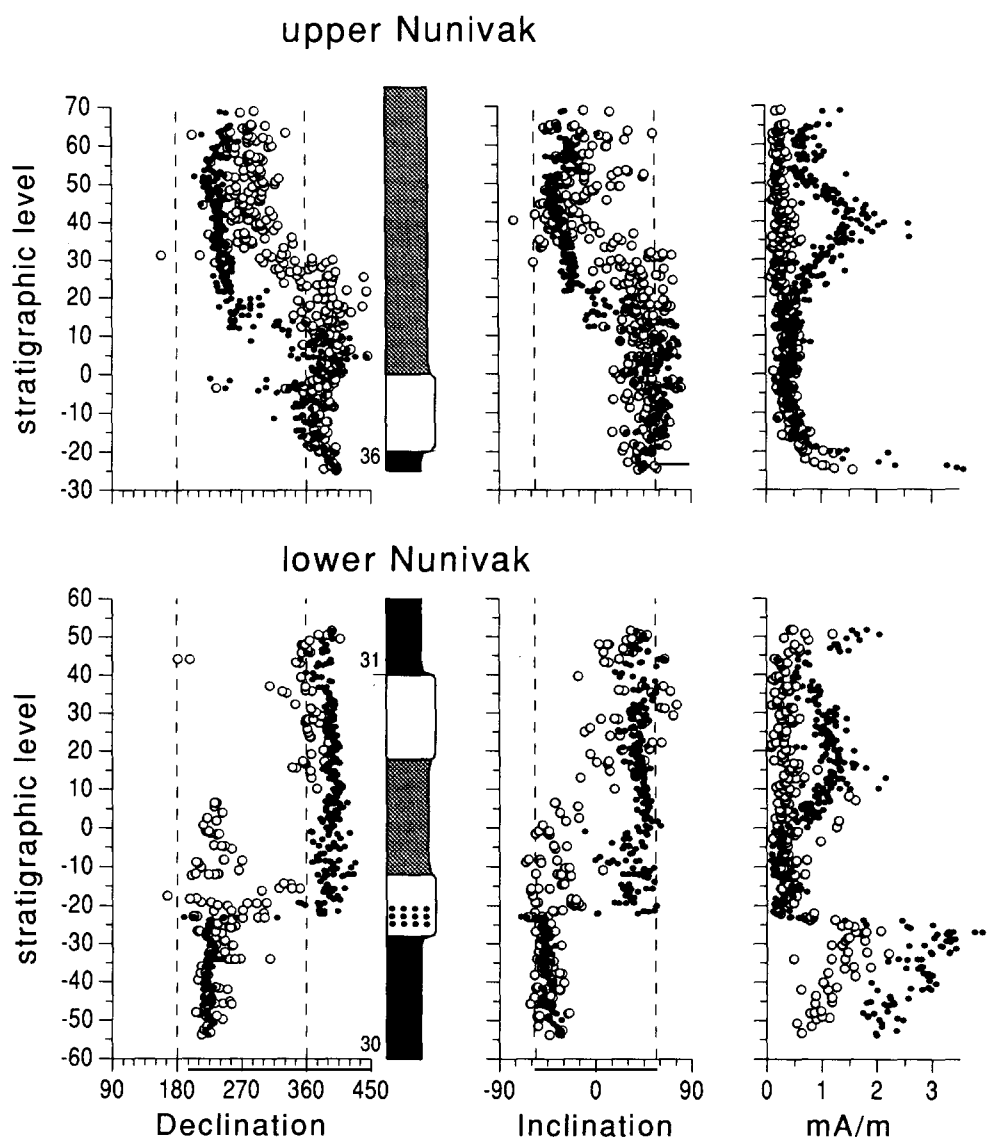


Fig. 8. Records of the declination, inclination and intensity obtained after thermal demagnetization. For explanation of sedimentary column, see Fig. 3. Figures along the stratigraphic column refer to the small scale sedimentary cycles (Hilgen, 1991). Circles denote the LT component; black spots the HT component. Dashed lines indicate declination and inclination ( $57.5^\circ$ ) of the geocentric axial dipole for the present latitude of the locality. Lower Nunivak transition: The change reversed to normal directions in HT component is abrupt and takes place deeper in the sediment than the gradual change of the LT component. The intensities of HT as well as LT decrease at the depth of the transition of the HT component. The LT intensities do not recover; the HT intensities recover partially and with a second minimum at level 40. Upper Nunivak transition: the reversal in the HT component is preceded by an excursion in declination. Between levels 5 and 20 cm the HT component changes from normal to reversed relatively quickly. Above level 20 cm the change is very slow, until level 50 cm. Also here the transition of the LT component is higher in the sediment and slower. The intensities both drop at level -20 and as in the lower Nunivak only the HT component recovers partially followed by another minimum.

the subsequent formation of (SP) magnetite (Van Velzen and Zijdeveld, 1992). Where possible, directions of the LT and the HT components have both been determined.

The demagnetization diagrams show that the stable reversed and normal polarity directions (Fig. 7(a), (b), (e), (f), (g) and (i)) show a clockwise rotation as observed in the other Trubi sections on Sicily (Langereis and Hilgen, 1991), and this is a strong indication of the primary nature of both the LT and HT components. In addition, the perfect correlation of the magnetostratigraphy of the Rossello composite section with the geomagnetic polarity time scale (GPTS) (Langereis and Hilgen, 1991) marks the primary character of the components.

## 5. The transition records

The directions of the LT and HT components for the two transition records are shown in Fig. 8. In both records, the HT component reverses lower in the sediment than the LT component, as described earlier by Van Hoof and Langereis (1991). In the LN record, the HT component changes abruptly from reversed (R) to normal (N) directions at level  $-23$  cm in the lower white layer. The LT component changes gradually some 35 cm above the HT reversal. The intensities of both components collapse at the level where the HT component reverses. Only the intensities of the HT component recover twice (levels 20 cm and 50 cm) but they do not regain the pre-transitional values.

The actual transition of the UN record is preceded by a change in declination from  $35^\circ$  to  $360^\circ$  back to  $35^\circ$  in the white layer, between levels  $-25$  cm and 5 cm. At the  $-25$  cm level the intensities of both components also collapse. A sudden jump to south/up directions is observed at the  $-5$  cm level. During these directional changes, the inclination is somewhat steeper and the intensities do not recover in this interval. As in the LN record, the reversal in the HT component is much more abrupt than in the LT component; there seems to be an intermediate 'cluster' with a direction of approximately  $250^\circ$  in declina-

tion and  $-30^\circ$  in inclination, between levels 10 and 20 cm. Next, the HT component changes slowly to reversed directions at level 50 cm. During this gradual change in direction, the HT intensity recovers to a relative maximum at the 40 cm level, followed by a decrease. The transition of the LT component is initiated at the 20 cm level and it finishes at the 50 cm level. The LT component does not reach a completely reversed direction; it shows more scatter and its intensities, collapsed at the  $-20$  cm level, remain low.

## 6. Discussion

### 6.1. The Sicilian lower and upper Nunivak records

The reliability of the NRM of sediments representing the directions of geomagnetic field during its reversals is subject to discussion (Van Hoof and Langereis, 1991; Langereis et al., 1992; Mary and Courtillot, 1993; Quidelleur et al., 1992; Van Hoof et al., 1992). Because the (non-transitional) directions of both the LT and HT show the clockwise rotation of the sedimentary basin, an inclination error and a perfect correlation to the GPTS, the acquisition of these components took place a very short time after sedimentation. However, their transitional characteristics are very different. This implies that at least one of the two components does not reflect the geomagnetic behaviour during the reversals. The direction and intensity of the HT component in the LN record changes very abruptly at a sedimentary level that contains very persistent brown spots in the fresh blue clay, making the component suspect. This will be discussed in more detail in the context of the diagenetic magnetite formation model. At the brown spotted level, the intensity of the LT component also collapses, strongly indicating that the intensities of either component also do not represent the relative intensity of the geomagnetic field.

### 6.2. Comparison of the Nunivak records from Sicily and Calabria

We have compared the Sicilian and Calabrian records from the same Nunivak transitions, be-

cause both records should be identical over such a small geographical distance (250 km). The Calabrian Nunivak transition records were earlier

reported as two of five successive transitions (Linssen, 1991); they were sampled at the Monte Singa section in southern Italy. The grey to

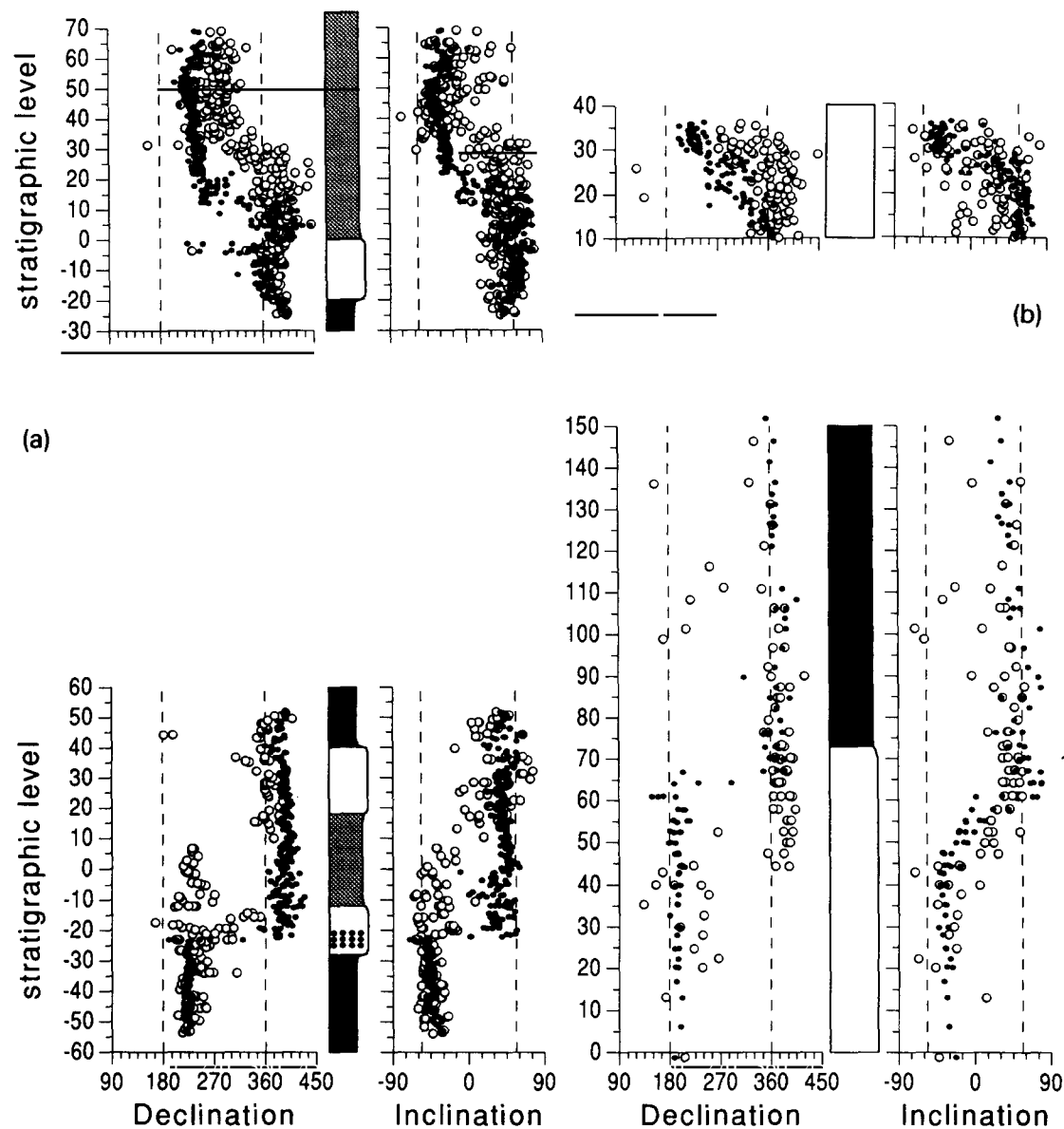


Fig. 9. Comparison of the two transitions from this paper at the left hand side and the same ones from the Calabrian Trubi formation (Linssen, 1991) at the right-hand side. (a) Lower Nunivak: the vertical scale has been changed to calibrate the same lithological boundaries from white to grey at level 40 cm of the Sicilian record and level 73 of the Calabrian record. The boundary from grey to white does not come together with a boundary in the Calabrian Trubi. In the Calabrian record the HT component changes above the transition in the LT component, contrary to the Sicilian record. The HT as well as the LT component change gradually. (b) Upper Nunivak: transition of the HT component is less far below the change of the LT component than in the Sicilian transition.

white–beige–white sequence in the Sicilian Trubi is equivalent to a grey–to–white sequence in the Calabrian Trubi (Hilgen and Langereis, 1989). Although the Calabrian records have a somewhat lower resolution than the Sicilian ones, the characteristics of the transitions should be identical. It appears, however, that in the Calabrian lower Nunivak record the transition of the HT component is gradual in inclination and instantaneous in declination, whereas the same transition recorded in Sicily is instantaneous in inclination and declination (Fig. 9(a)). More important, the LT component in the Calabrian lower Nunivak transition changes polarity lower in the sediment than the HT component, quite contrary to the Sicilian record.

Both LT and HT components in the Calabrian upper Nunivak transition show some similarities to the record from Sicily; the transitions are gradual and have west/shallow intermediate directions. However, the timelag between the two components in the Calabrian upper Nunivak is somewhat less while the scatter in the LT component is larger, especially in the inclination where no fully reversed or normal directions are reached.

### 6.3. The diagenetic magnetite formation model

Changes in rockmagnetic parameters and of the NRM intensities and directions near a reversal often coincide with the lithological changes in these marls (Van Hoof et al., 1992 and references therein). This correspondence has been explained by Van Hoof et al. (1992) and will be summarized below.

In their 'diagenetic magnetite formation model' they describe the formation of magnetite during early diagenesis, and the new remanence is superposed on the old remanence direction acquired during deposition.

The diagenetic conditions in the sediment varied with the amount of metabolizable organic carbon (Froelich et al., 1979). During deposition (primary) magnetite was formed and preserved in the sediment. After deposition, the diagenetic conditions were the most reducing in the grey layer where sulphate reduction took place. How-

ever, magnetite was probably preserved because iron hydroxides (which are less stable) react more easily with  $\text{HS}^-$  (Canfield and Berner, 1987). After the deposition of a grey layer a progressive oxidation front ('burn-down') is formed as oxygen re-enters the sediment (Wilson et al., 1985). This causes the formation of an iron and manganese enrichment just above the grey layer, in the lower part of a white<sub>(1)</sub> layer.

After burial of the grey/white<sub>(1)</sub>/beige/white<sub>(2)</sub> sequence, suboxic conditions develop as sulphate reduction ceases (Berner, 1981). These diagenetic circumstances favour the formation of (secondary) magnetite (Karlin and Levi, 1985; Karlin, 1990) especially in the previously formed iron (hydroxide) enriched layer, but also below the buried grey layer. The extent of secondary magnetite formation depends on the availability of reactive amorphous iron hydroxides and the distance of  $\text{Fe}^{2+}$  migration from the grey layer.

This implies that in these sediments there is a time lag of at least one cycle white<sub>(1)</sub>/beige/white<sub>(2)</sub> between the formation of secondary and primary magnetite. If during this time interval the polarity of the earth magnetic field changes, the secondary magnetite will record the new direction. The dominance of either primary magnetite or secondary magnetite will determine the final direction of the remanence.

In the LN and UN transitions, the 'diagenetic magnetite formation' closely fits the observed characteristics of the inclination and declination. According to the model, in the upper grey layers of both the UN (grey 37, not shown) and the LN (grey 31, Fig. 8) only primary magnetite was formed, representing the polarity after the reversal. Below the grey layers, however, the remanence carried by the secondary magnetite may be dominant as the observed remanence has the post-transitional direction. Assuming that the lower grey layers, as predicted by the model, (respectively, grey 30 in LN and grey 36 in UN) have recorded the pre-transitional directions during deposition, the reversal must have taken place between the grey layers. The concurrence of the LT and HT components in the lower grey layers gives us additional evidence that they predominantly contain primary magnetites.

There is a lag between the LT and HT components in both reversal records (Fig. 8). The transitions recorded by the HT component are lower in the sediment than the transitions recorded by the LT component, suggesting that the diagenetic formation of HT magnetite extends deeper than the diagenetic formation of the LT magnetite.

The abrupt change in the recorded directions of the HT component in the LN record is a good example of the preferential formation of diagenetic magnetite in the iron enrichment zone that was formed during the burn-down stage in early diagenesis (Van Hoof et al., 1992). This preferential formation during suboxic diagenesis is enhanced by iron hydroxides, still clearly visible as brown spots in the white<sub>(1)</sub> layer.

The transition is less abrupt in the UN reversal record. This can be explained by the lower concentration of reactive amorphous iron hydroxides than found in the iron enriched front of the white<sub>(1)</sub> in quadruplet 30. In addition, because of the larger distance to the grey 37 layer, diagenetic conditions were less reducing and did not allow

high Fe<sup>2+</sup> concentrations. The amount of secondary magnetite formed was therefore limited and the remanence carried by the secondary magnetite is smaller. As a result, intermediate and/or pre-transitional directions recorded by primary magnetite will dominate post-transitional directions.

#### 6.4. Intermediate directions

In a statistical study of recently obtained transition records Tric et al. (1991) showed that VGP paths of two-thirds of the transitions follow more or less a great circle over North and South America or over its antipode. Laj et al. (1991) pointed out that the same bands of longitude are important in other geophysical observations, such as the pattern of fluid motion in the outer core and regions of higher seismic velocities in the lower mantle, and suggested a causal relationship. However, smoothing of non-antipodal direction before and after a transition also results in a VGP path with a similar longitudinal confine-

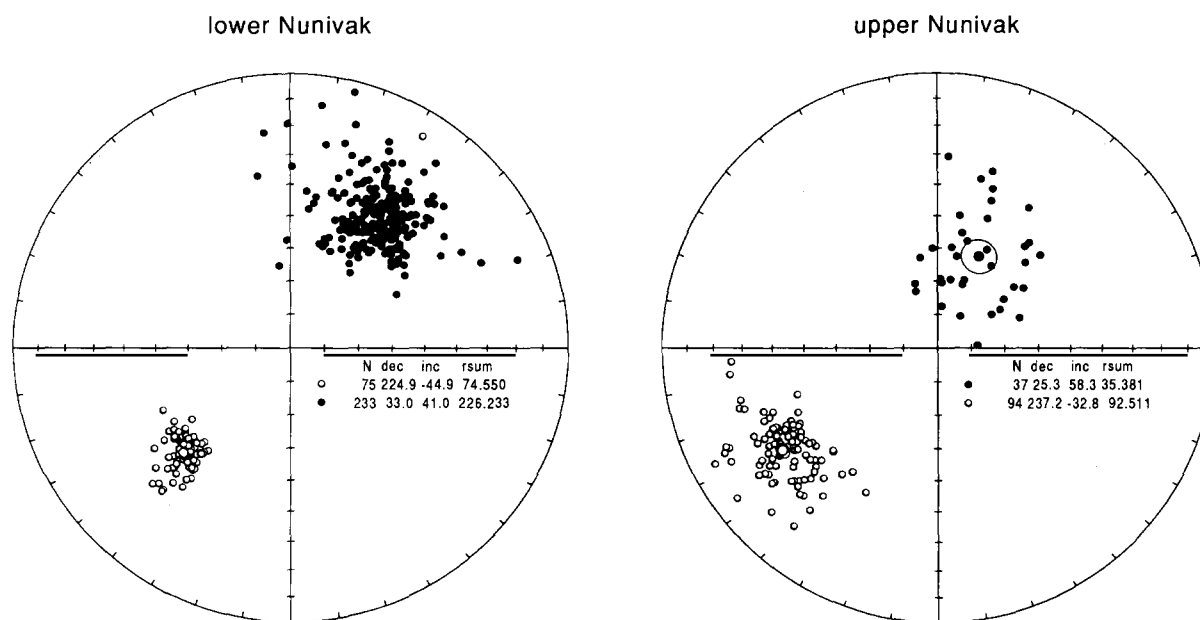


Fig. 10. Stable HT directions from the two transitions determined in the LN record below level -23 cm (before the transition) and above level -19 cm (after) and in the UN record between levels 0 and 8.5 cm (before) and above level 35 cm (after). The means of the stable directions before and after the transitions show a clear offset.

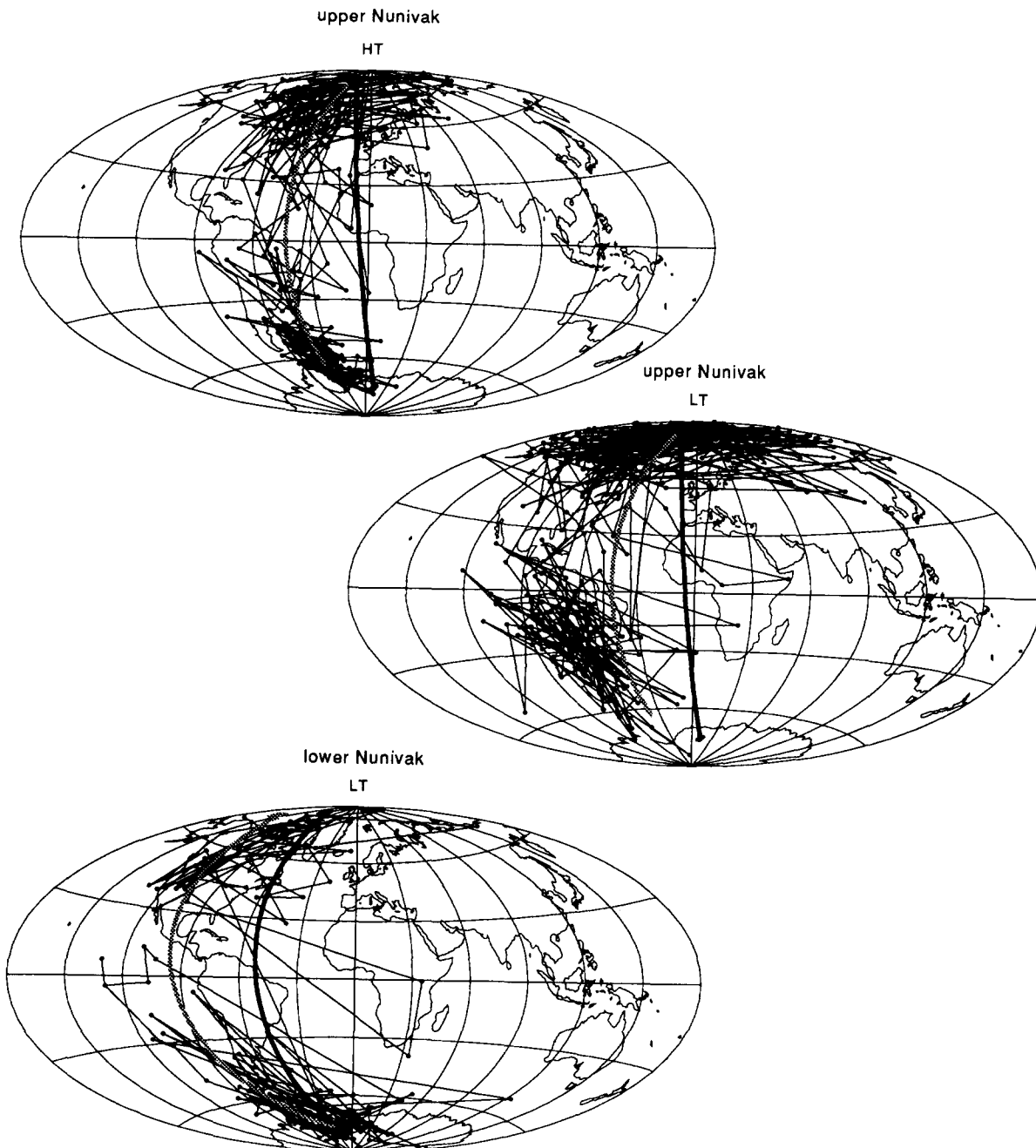


Fig. 11. VGP paths of the lower and upper Nunivak transitions. The solid black line shows the VGP path obtained by filtering the mean directions of under/overlying polarity zones resulting from magnetostratigraphy (Langereis and Hilgen, 1991), the shaded line represents filtering of near-transitional directions determined from the reversal records. The synthetic VGP paths of the mean non-transitional directions are in good correlation with the observed VGP paths while the coincidence with the synthetic VGP path of the smoothed near transitional directions is even better.

ment (Rochette, 1990). Using the same procedure of Rochette (1990) the mean stable directions before and after 23 transitions sampled in the Mediterranean were used by Langereis et al. (1992) to calculate synthetic VGP paths. These modelled VGP paths appeared to be confined to the Americas and were to a large extent identical to the observed VGP paths.

As discussed in the previous section, diagenetic formation of secondary magnetite can cause post-transitional directions to occur below the chronostratigraphic level of the actual geomagnetic reversal. The vector sum of pre- and post-transitional magnetizations (and even true intermediate directions) leads to smoothing, the extent of which depends on the relative contributions of the primary and secondary magnetite. The VGPs of the lower and upper Nunivak transitions have been calculated after applying a 35° correction for the clockwise rotation of the location. No VGP path has been calculated for the HT component in the LN transition, because this component shows no intermediate directions. The VGP paths of LT components of the two transitions are, with the exception of a few outliers, confined to the Americas, whereas the path of the HT component of the UN transition tends to lie more in the Atlantic (Fig. 10).

Mean stable directions of the polarity zones before, during and after the Nunivak subchronozones (mean non-transitional directions) were calculated using the magnetostratigraphic results of the Punta Maiata section (Langereis and Hilgen, 1991; Table 1). They have negative reversal tests (McFadden and McElhinney, 1990), thus indicating a significant non-antipodality.

TABLE 1

Mean stable directions and ages of the polarity zones before, during and after the Nunivak subchronozones after Langereis and Hilgen (1991)

Subchron (age in Ma)	DEC	INC	<i>N</i>	<i>R</i>
4.85–4.65	223.2	–44.9	13	12.79
Nunivak	33.5	50.9	12	11.93
4.53–4.35	214.7	–41.0	17	16.78

*N*, number of samples; *R*, unit vector sums.

Synthetic VGP paths were determined both from the mean non-transitional stable directions and by smoothing the mean near-transitional stable directions just before and after the transitions as determined from the present records (Fig. 10). The observed as well as the synthetic VGP paths (Fig. 11) show a remarkable coincidence as was earlier observed in most of the transitional records from the Sicilian Trubi marls (Langereis et al., 1992).

A gradual change in dominance of pre-transitional directions over post-transitional directions can explain the smoothed intermediate directions. This smoothing mechanism may thus obscure the real geomagnetic transitional directions. Hoffman (1992) found in transitional records from lava sequences, which are usually considered as not smoothed, long-lived and recurring VGP positions that cluster at locations on the globe coinciding with the preferential longitudinal bands over the Americas or its antipode as found by Laj et al. (1991) and Tric et al. (1991). If these clusters represent a stage of a non-axial dipolar transitional configuration, the VGPs will be independent of the sampling site on the globe. Filtering of a transition that includes such a cluster will evidently result in a VGP path within a longitudinal band passing that cluster. In spite of the smoothing mechanism of the sediment, some information about the transitional path may therefore still be registered by sediments. If the NRM is strongly smoothed, the clusters will not only confine the VGP paths over longitudinal bands passing their location, they will also bias the near-transitional directions. In this respect, it may be noted that both the HT and LT component from the upper Nunivak show a strong clustering just south of South America (HT) and — more scattered — over South America itself (LT). This may equally well be the result, however, of significant secondary magnetite formation in a particular lithological interval and causing smoothing of pre- and post-transitional directions to an intermediate cluster. The offset in the mean non-transitional directions is not easily explained as a result of smoothing of these cluster directions and the directions of the stable polarity zones, because it would cause the filter to be at least as

wide as half the polarity zones. In that case, the major directional changes in the transition records would occur over much larger intervals than presently observed. A smoothing mechanism therefore does not exclude a cluster of VGPs lying in southern America as was found by Hoffman (1992), but it cannot account for the non-antipodal offset of the non-transitional stable directions as found for the entire polarity zones before, during and after the Nunivak subchronozone.

### Conclusions

The transitional records of the upper and lower Nunivak transitions show that magnetites are the most important carriers of the remanence in both records. The transitions are recorded in two components; a low temperature (LT) and a high temperature (HT) component. The two components do not reverse simultaneously, nor are their transitional characteristics identical. There are major changes in direction (HT component) and intensity (LT and HT component) at lithological boundaries that make the two components unreliable in representing the geomagnetic field. In addition, the significant differences in directional changes between these records and the records from the same transitions sampled some 250 km away confirm the unreliability. The longitudinal confinement of the VGP paths is probably caused by smoothing of the stable directions before and after the transitions. The directional changes as well as the smoothing mechanism can be explained by diagenetic formation of magnetite (Van Hoof et al., 1992). Because of the smoothing, a long lived cluster of intermediate VGPs as found by Hoffman (1992) during the transition cannot be excluded.

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