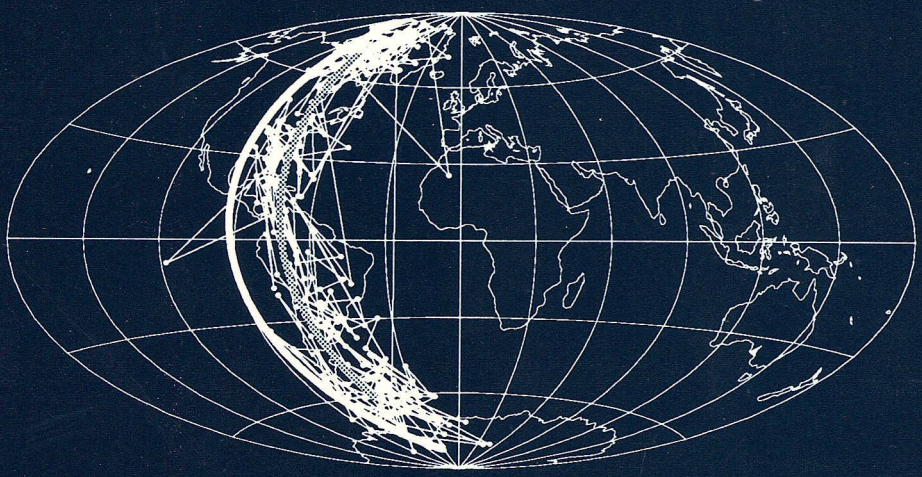
GEOLOGICA ULTRAIECTINA

Mededelingen van de
Faculteit Aardwetenschappen der
Rijksuniversiteit te Utrecht

No. 100

GEOMAGNETIC POLARITY TRANSITIONS OF THE GILBERT AND GAUSS CHRONS RECORDED IN MARINE MARLS FROM SICILY



A. A. M. VAN HOOF

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GEOMAGNETIC POLARITY TRANSITIONS OF THE GILBERT AND GAUSS CHRONS RECORDED IN MARINE MARLS FROM SICILY

Geomagnetische polariteitsomkeringen uit de Gilbert en Gauss Chrons geregistreerd in marine mergels van Sicilië

(met een samenvatting in het Nederlands)

PROEFSCHRIFT

TER VERKRIJGING VAN DE GRAAD VAN DOCTOR AAN DE RIJKSUNIVERSITEIT TE UTRECHT, OP GEZAG VAN DE RECTOR MAGNIFICUS, PROF. DR. J.A. VAN GINKEL, INGEVOLGE HET BESLUIT VAN HET COLLEGE VAN DEKANEN IN HET OPENBAAR TE VERDEDIGEN OP 18 MAART 1993 DES NAMIDDAGS TE 14.30 UUR

DOOR

ANTHONIUS ADRIANUS MARIA VAN HOOF

geboren op 31 oktober 1959 te Veghel

PROMOTOR: PROF. DR. J.D.A. ZIJDERVELD
CO-PROMOTOR: DR. C.G. LANGEREIS

Zeker weten doe ik nooit iets
en zelfs dat weet ik niet zeker.

voor mijn overleden moeder
voor mijn vader, mijn broers en zussen

Dit proefschrift had nooit tot stand kunnen komen zonder de samenwerking met dr. C. G. (Cor) Langereis. Zijn kritiek, suggesties, discussies, ideeën en niet in de laatste plaats zijn persoon zijn de oorzaken voor een bijzonder plezierige, en daardoor succesvolle periode als student en promovendus op "het Fort".

Piet-Jan Verplak heeft veel van de ontelbare kerntjes gemeten waarvan de data de basis van dit proefschrift vormen. Bovendien zorgde hij er mede voor dat de vele veldwerken de hoogtepunten van het onderzoek waren.

Prof. Dr. J. D. A. Zijderveld dank ik voor zijn steun tijdens het onderzoek en zijn kritiek die hij gaf op de manuscripten.

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Curriculum Vitae

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Samenvatting

Een van de meest opvallende waarnemingen in de geofysica is het feit dat het aardmagneetveld in het geologisch verleden herhaaldelijk van polariteit is gewisseld. De polariteitsovergangen van de laatste 165 miljoen jaar zijn nauwkeurig gedocumenteerd door hun continue registratie als mariene magnetische anomalieën aan weerszijden van spreidende mid-oceanische ruggen. De (veranderingen in de) frequentie waarmee polariteitsomkeringen optraden gedurende het geologische verleden kunnnen ons iets vertellen omtrent langperiodische veranderingen - gedurende miljoenen jaren - in processen die plaatsvinden in de ondermantel en de buitenkern van de aarde, en daarmee iets over de oorsprong van processen die van invloed zijn op het aardmagneetveld en het ontstaan daarvan. Het onderwerp van dit proefschrift houdt zich echter met een totaal andere tijdsdimensie bezig: dat van de individuele polariteitsomkeringen die plaatsvinden in 'luttele' duizenden jaren.

De karakteristieke veranderingen van het aardmagneetveld vlak voor, tijdens en vlak na een polariteitsomkering kunnen ons helpen iets meer te weten te komen over de oorsprong van het aardmagneetveld, en met op welke wijze een polariteitsovergang plaatsvindt. De natuurlijke remanente magnetisatie (NRM) van lavas en sedimenten bevat informatie omtrent het aardmagneetveld ten tijde van het ontstaan van deze gesteenten. Een opeenvolging van lavastromen levert weliswaar zeer betrouwbare maar discontinue informatie vanwege het feit dat het aardmagneetveld alleen dan geregistreerd wordt op het moment dat er een lavastroom is uitgevloeid. Dat geschiedt in het algemeen echter zelden ten opzichte van de snelle geomagnetische veranderingen tijdens een polariteitsomkering. Een opeenvolging van sedimenten daarentegen, representeert een vrijwel continue registratie van veranderingen in het aardmagneetveld, omdat sedimentatie in het algemeen beschouwd wordt als een continue proces. In dit proefschrift worden polariteitsomkeringen uit de Gilbert en Gauss Chrons onderzocht die zijn geregistreerd in de mariene mergels van Sicilië en Calabrië (Zuid Italië). Samen met acht omkeringen die al eerder door J.H. Linssen (1991) voor dit tijdsinterval onderzocht zijn, vormen zij een volstrekt unieke serie van 13 opeenvolgende transities. Van de 13 verschillende transities zijn er vijf dubbel bemonsterd waarvan steeds één in Sicilië en één in Calabrië. De geografische afstand tussen de beide gebieden is ongeveer 250 km. De Gilbert en Gauss Chrons vormen een aaneengesloten tijdsinterval van 5.4 tot 2.6 miljoen jaar geleden. De 13 omkeringen vonden plaats tussen 5.27 en 2.92 miljoen jaar geleden, en een dergelijke opeenvolging geeft in principe een uitmuntende gelegenheid om het gedrag van het aardmagneetveld tijdens omkeringen over een langere tijd te vervolgen, en te zien of zich daarin veranderingen voltrekken, bijvoorbeeld op een termijn van enkele miljoenen jaren. De dubbele registraties vormen dan een extra controle op de betrouwbaarheid waarmee de onderzochte sedimenten het veranderende geomagnetische veld vastleggen. Immers, op wereldschaal vormen beide gebieden een practisch eenzelfde locatie alwaar geomagnetische veranderingen identiek zouden moeten zijn.

De Pliocene sedimenten van Sicilië waren eerder magnetostratigrafisch in detail onderzocht (Langereis en Hilgen, 1991), en dus was de exacte locatie van de omkeringen in het sediment goed bekend. De sedimenten bestaan uit mariene mergels en bevatten een afwisseling van kleinschalige cycli die samenhangen met de precessiecyclus van de Aarde, met een gemiddelde periode van ongeveer 21.700 jaar. Elke cyclus heeft een gemiddelde dikte van ongeveer een meter en bestaat uit een grijze, witte, beige en witte mergellaag. In de sedimenten van Calabrië is de wit-beige-wit opeenvolging echter aanwezig als slechts één enkele witte laag. De grijze lagen correleren met minima in de precessiecyclus (Hilgen, 1991) en dit gegeven maakt het in theorie mogelijk om eventuele veranderingen van het aardmagneetveld zeer nauwkeurig in de tijd te plaatsen, zowel qua ouderdom als qua duur. Belangrijker nog was het feit dat de magnetostratigrafische resultaten een perfecte correlatie met de (standaard) magnetische tijdschaal laten zien - hetgeen een primaire oorsprong van de magnetisatie zeer aannemelijk maakt - en tezamen met de goede magnetische eigenschappen van het sediment leken deze bij uitstek geschikt voor een gedetaileerd onderzoek van het gedrag van het aardmagneetveld tijdens polariteitsomkeringen.

Detail onderzoek van omkeringen vereist een gedegen aanpak van elke aspect vanaf het moment van bemonsteren - waaraan zeer veel aandacht werd besteed - tot aan de uiteindelijke interpretatie van de grote hoeveelheden gegevens die noodzakelijk zijn voor dit type onderzoek. In het bijzonder moet eerst nauwkeurig bepaald worden in hoeverre het sediment geschikt is, óók

voor het registreren van zeer snelle geomagnetische veranderingen die - naar men aanneemt - optreden tijdens een omkering. Het eerste deel (Part 1) van dit proefschrift behandelt daarom hoe de onderzochte sedimenten hun remanente magnetisatie verkregen.

De kenmerken die te voorschijn kwamen uit met name de uitgebreide demagnetisatieprocedures lieten zien dat de diepte waarop de NRM in het sediment vastgelegd wordt tijdens de afzetting, niet constant is (Chapter 1). De variatie hierin is zelfs zodanig dat een en dezelfde omkering op verschillende intervallen in de sedimentkolom geregistreerd wordt. Opvallend is tevens dat sterke richtingsveranderingen in de NRM vaak voorkomen op lithologische grenzen, namelijk op die van de verschillend gekleurde mergellagen. Hieruit kon al snel geconcludeerd worden dat de waargenomen richtingsveranderingen voornamelijk het gevolg zijn van processen tijdens de sedimentatie, en niet het de details van het werkelijke gedrag van het aardmagneetveld weergeven.

Men spreekt van intermediaire richtingen wanneer ze duidelijk verschillen van de richting die men kan verwachten als het aardmagneetveld in een stabiele toestand verkeert, namelijk de richting behorend bij een dipoolveld dat ongeveer volgens de rotatieas van de Aarde gericht is. Nu blijkt echter dat wanneer men de waargenomen stabiele richtingen van vóór en nà een omkering filtert met een 'moving window', men een (model)registratie van de omkering verkrijgt met eigenschappen die zeer goed overeenkomen met de werkelijk waargenomen registratie (Chapter 2). Dit suggereert dat de waargenomen intermediaire richtingen tenminste voor een belangrijk deel het resultaat zijn van de filterwerking van het sediment. Het maakt het tevens waarschijnlijk dat eventuele snelle veranderingen van het omkerende veld grotendeels of zelfs geheel verborgen blijven.

De dubbele registraties, met een onderlinge afstand van 250 km, laten niet die overeenkomsten zien die men mag verwachten als slechts geomagnetische veranderingen vastgelegd zijn. Bijvoorbeeld, de intervallen met intermediaire richtingen vertonen duidelijke verschillen in lengte (duur), of dezelfde omkeringen vertonen niet dezelfde intermediaire richtingen, terwijl zelfs het niveau in het sediment waar de omkering plaatsvindt (of lijkt te vinden) niet altijd overeenkomt.

Om die redenen is mede op basis van gesteentemagnetische en geochemische waarnemingen een model ontwikkeld (Chapter 3) dat als doel heeft de verschillen in (de diepte van het) vastleggen te verklaren, en om de filterwerking van het sediment te verduidelijken. Volgens dit model wordt een magnetisatie die gedragen wordt door zogeheten 'secundaire magnetiet' gesuperponeerd op een magnetisatie in 'primaire magnetiet' die al eerder - vrijwel meteen tijdens afzetting - ontstond. De NRM gedragen door secundaire magnetiet wordt een aantal duizenden jaren na die van primaire magnetiet verkregen. Dit houdt in dat wanneer deze secundaire magnetiet gevormd wordt *na* een omkering, terwijl (op hetzelfde niveau) de primaire magnetiet reeds de richting van *voor* de omkering heeft vastgelegd, de resulterende magnetisatie de vectoriële som is van twee min of meer antiparallelle richtingen. De mate waarin superpositie, of zelfs substitutie plaatsvindt hangt klaarblijkelijk af van de reducerende en oxiderende omstandigheden in het sediment, terwijl de maximum diepte tot waarop dit proces invloed heeft ongeveer een meter (of soms meer) is. De details van het omkeringsproces worden hierdoor in belangrijke mate verstoord, hoewel het sediment verder even geschikt blijft voor magnetostratigrafische doeleinden. Desalniettemin lijkt het erop dat de sedimentaire registraties van deze omkeringen toch nog belangrijke informatie bevatten omtrent het omkeringsproces. Het blijkt dat opvallende kenmerken die waargenomen zijn in (als betrouwbaar geldende) vulkanische gesteenten ook te zien zijn in deze sedimenten, zoals typische en vaak terugkerende configuraties ('clusters') van het omkerende veld (Hoffmann, 1992). Aangezien tevens de geografische posities van die clusters sterk overeenkomen met die van andere geofysische waarnemingen, zoals gebieden met hogere seismische snelheden in de ondermantel, lijkt er niet van toeval sprake te zijn en wint derhalve een oorzakelijk verband aan waarschijnlijkheid.

In het tweede deel van dit proefschrift worden de individuele registraties besproken, inclusief hun details op gesteentemagnetisch gebied. Een zekere dichotomie in de interpretaties van gelijksoortige waarnemingen is onvermijdelijk aanwezig, aangezien een aantal registraties, c.q. hoofdstukken uit dit proefschrift, reeds gepubliceerd waren voordat de konsekwenties van de vroeg-diagenetische veranderingen (de secundaire magnetiet zoals hierboven genoemd) volledig bekend waren. Daarnaast werd pas in een later stadium gerealiseerd - naar aanleiding van de zeer onlangs gepubliceerde 'Hoffmann clusters' (Hoffmann, 1992) - dat enkele opvallende eigenschappen in de sedimentaire registraties opmerkelijk goed correleren met deze - waarschijnlijk relatief langdurige - clusters uit vulkanieten.

Om die redenen, tenslotte, is in het laatste hoofdstuk getracht de meest opvallende kenmerken van de volledige sequentie van 13 opeenvolgende polariteitsomkeringen - met inbegrip van de registraties uit Calabrië - op een rij te zetten, zowel ten aanzien van de sedimentaire artefacten als ten aanzien van werkelijk geomagnetisch gedrag. Echter, blijkt ook dat een echt betrouwbare vergelijking met vulkanitische registraties bemoeilijkt wordt door het niet of nauwelijks beschikbaar zijn van registraties in lavas van juist de omkeringen die het onderwerp van de huidige studie zijn. Pas als die beschikbaar komen kan in betrouwbare mate de vraag beantwoord worden in hoeverre de onderzochte sedimenten werkelijk geschikt zijn voor het bestuderen van (kenmerkende eigenschappen van) het aardmagneetveld tijdens omkeringen.

Hilgen, F.J., Extension of the astronomically calibrated (polarity) time scale to the Miocene/Pliocene boundary, *Earth planet. Sci. Lett.*, 107, 349-368, 1991.

Hoffman, K.A., Dipolar reversal states of the geomagnetic field and core-mantle dynamics, *Nature*, 359, 789-794, 1992.

Langereis, C.G. and Hilgen, F.J., The Rossello composite: A Mediterranean and global reference section for the Early to early Late Pliocene. *Earth planet. Sci. Lett.*, 104, 211-225, 1991.

Linssen, J.H., Properties of Pliocene sedimentary geomagnetic reversal records from the Mediterranean, (PhD. Thesis, University of Utrecht), *Geologica Ultraiectina*, 80, 230 pp., 1991.

Summary

One of the most fascinating phenomena of geophysics is the fact that in the geological past the Earth's magnetic field has frequently reversed its polarity. These polarity transitions are accurately established during at least the past 165 Myr - from their recording in the ocean floor: the marine magnetic anomalies. The (changes in) reversal frequency during the geological past may tell us something about long-term changes in the lower mantle and outer-core processes, and hence constrain the origin of the field itself. The subject of this thesis, however, deals with an entirely different scale of time: that of individual polarity reversals, taking place in several thousands of years.

The characteristics of the geomagnetic field during a change in polarity may help us to unravel the processes that govern the origin of the geomagnetic field and its transitions. The natural remanent magnetisation (NRM) of lava's and sediments contains information of the Earth's magnetic field at the time of formation of these rocks. A sequence of lava flows represents spot readings scattered in time since the Earth's magnetic field is only recorded when intermittent individual lava flows cool after their extrusion. A sedimentary sequence may give a more continuous registration of the Earth's magnetic field in the past because sedimentation is - assumedly - a continuous process. In this dissertation, geomagnetic polarity transitions from the Gilbert and Gauss Chronozones, recorded in marine marls from Southern Italy will be discussed. Together with eight transitions from the thesis by J.H. Linssen (1991) they form a sequence of 13 successive transitions, five of which were sampled as duplicate records from two parallel sections, one on Sicily and one in Calabria. The distance between the two sections is approximately 250 kilometres. The 13 transitions have taken place between 5.27 and 2.92 million years before present. This sequence would give a unique opportunity to study the time dependent transitional behaviour of the geodynamo. The duplicate records are an extra control for the reliability of the sediment to record the transitional geomagnetic field, because on a global scale the sampling sites of the duplicate records are at the same location and hence the directional behaviour should be identical.

The Pliocene sediments from Sicily have been the subject of detailed magnetostratigraphic studies (Langereis and Hilgen, 1991), and which studies have revealed the the precise location of the transitions. These sediments are made up of an alternation of small scale cycles that are related to the precessional cycle of the Earth with an average period of approximately 21.7 kyr. Each individual cycle consists of a grey, white, beige and white layer and has an average thickness of one metre. In the Calabrian sediments a white, beige, white alternation is represented by one single white layer. The grey layers can be correlated with the minima in the precessional cycles (Hilgen, 1991) and this makes it possible to establish very precise time constraints of the observed directional changes. In addition, the perfect correlation of the magnetostratigraphic results to the standard geomagnetic polarity time scale, as well as the high (paleo)magnetic quality of the data, made the sediments seem extremely suitable for studies of geomagnetic polarity reversals.

The detailed study of sedimentary reversal records requires careful attention to every aspect along the trajectory from sampling to interpretation of the final results. In particular, the ability of the sediment to record (rapid) geomagnetic changes must be thoroughly investigated. For this reason, Part 1 of this thesis is dedicated to the mechanisms of NRM acquisition in these sediments.

The characteristics of the demagnetisations of the remanence have revealed that - during deposition - the depth below the sediment/water interface where the direction of the geomagnetic field is acquired, is not constant (chapter 1). A same transition can be recorded at different levels in the sediment column. In addition, several directional changes appear to take place at lithological boundaries of the sedimentary cycles. Therefore, the directional changes of the remanence as a function of the stratigraphic level are sedimentary artefacts rather than a true representation of the changing geomagnetic field.

Directions are intermediate when they are clearly deviating from the direction of the Earth's stable axial dipole field. It appears that if a moving window is applied that filters the stable directions observed before and after a transition,

synthetic reversal records are produced that show virtual geomagnetic poles (VGPs) fitting the observed VGP paths very well (chapter 2). This suggests that the observed intermediate directions during the transitions are at least partly result of a filtering mechanism in the sediment and that the real (high frequency) features of the geomagnetic field are obscured by this process.

The duplicate records with a separation of some 250 km do not show the similarities which are expected if they were accurate registrations of the geomagnetic field. For instance, the lengths of stratigraphic intervals with intermediate directions are different, the same transitions from the two locations do not have the coeval intermediate directions and the levels in the sediments where the transitions are recorded are different.

Based on the results of ample rock magnetic and geochemical studies, a model has been developed (chapter 3) to explain the variation of the depths at which the remanence is acquired and to clarify the filtering process. In this model, it is suggested that a remanence residing in "secondary magnetites" is superposed on the remanence of "primary magnetites" acquired during deposition of the sediment. The remanence of the secondary magnetites is acquired (several) thousands of years after that of the primary magnetites. Hence, if the secondary magnetites were formed *after* a polarity transition occurred, and the primary magnetites *before* that transition, then the resulting remanence is the vector sum of the two - more or less antipodal - directions. The extent to which superposition takes place is dependent on the reducing and oxidizing environment of the sediment. The maximum depth at which superposition can take place is approximately one metre (or more) below the sediment/water interface. The registrations of details of polarity transitions, therefore, may be partly distorted, while the sediment is still very suited for magnetostratigraphic work. It cannot be ruled out, however, that the sedimentary transition records contain some information of the reversing magnetic field, since some characteristics observed in volcanic records are also visible in these Southern Italian records. For example, typical recurring VGP clusters based on evidence from some lava flows (Hoffman, 1992) are also seen in our sedimentary records. Further, the coincidence of these clusters with other geophysical observations, such as higher seismic velocity patterns in the lower mantle, is remarkable and strongly suggests a relationship.

In Part 2 of this thesis, the individual records are presented, and their details - including (often) the rock magnetic properties of the sediments - are discussed. There is inevitably a certain dichotomy, since several records (chapters) have been published before the consequences of the early diagenetic magnetite formation (the "secondary magnetite" as referred to above) was fully realised. In the later established records, for instance that of the Gilbert/Gauss transition (chapter 8), it was recognized that particular features are present in our sedimentary records that correlate remarkably well with the suggested long-lived VGP from lava flows.

Therefore, in the last chapter (Part 3) a review is presented that attempts to catalogue the characteristics of the entire sequence of reversal records - including those of Calabria - in relation both to sedimentary artefacts and true field behaviour. But a reliable comparison with lava records is hindered by the scarcity, or rather absence of such records of the same transitions as recorded here. Such a comparison must give the answer to the question: to which extent are sediments suitable for the study of (long-lived) features of the Earth's magnetic field during reversals.

Hilgen, F.J., Extension of the astronomically calibrated (polarity) time scale to the Miocene/Pliocene boundary, *Earth planet. Sci. Lett.*, 107, 349-368, 1991.

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Linssen, J.H., Properties of Pliocene sedimentary geomagnetic reversal records from the Mediterranean, (PhD. Thesis, University of Utrecht), *Geologica Ultraiectina*, 80, 230 pp., 1991.

PART 1

PITFALLS AND POTENTIAL OF SEDIMENTARY REVERSAL RECORDS

Chapter 1

Reversal records in marine marls and delayed acquisition of remanent magnetization

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RECORDS of geomagnetic reversals 'frozen' into the magnetic components of sediments provide a means to study the time-dependent behaviour of the geomagnetic field during polarity transitions. Although sedimentary records have the advantage of being readily available and continuous, the process by which they acquire a remanent magnetization is still not fully understood[1]. Magnetites, which lose their magnetization at relatively high temperatures (≲580 °C), are generally considered to carry the primary remanence of the sediment—that is, to record the palaeomagnetic direction. Here we describe two reversed-to-normal transitional records from marine marls in which this high-temperature component shows a delayed remanence acquisition relative to a lower-temperature component. In these samples, therefore, the high-temperature component does not reflect geomagnetic changes during the reversal. At least for marine marls, detailed palaeomagnetic, rock-magnetic and geochemical studies are apparently necessary to judge the validity of reversal records.

The two studied (R–N) reversals are the lower Nunivak and the lower Cochiti, and are from a succession of 14 reversals which were identified by detailed magnetostratigraphic studies of marine marls on the southern coast of Sicily[2]. These marls are exposed as sedimentary cycles[3], each of which shows a characteristic colour layering (see Fig. 2 legend). The sedimentation rate is ~5 cm kyr^{-1} (ref. 4).

The two reversals have been sampled in a cliff section near Eraclea Minoa[5]. Removal of the weathered surface exposes the (blue coloured) fresh marls. Oriented cores were taken more or less parallel to the bedding plane, with an accurately determined stratigraphic spacing. The spatial resolution is a few millimetres, corresponding to a temporal resolution of <100 yr. But the specimen diameter (25 mm) will smooth the remanent magnetization over at least 500 yr.

Typical thermal demagnetizations are characterized by a viscous and/or secondary component removed below 200 °C, a low-temperature (LT) component removed between 200 and 480 °C, and a high-temperature (HT) component carried by single-domain magnetite[6] and which is removed at 580 °C. Occasionally, slightly higher maximum blocking temperatures are found which indicate the presence of cation-deficient magnetite[7,8] (Fig. 1*c*, *e*).

The LT component decays most strongly between 200 and 330–360 °C. Often, there is almost no decay between 360 and 480 °C, but some scatter may be observed which is possibly due to magnetic minerals being formed during heating. The HT component decays most strongly between 540 and 580 °C. This results in a distinctive and narrow peak in the blocking temperature spectrum which is consistently found not only in the Pliocene marls of Sicily[2,5] and Calabria[8,9], but also in late Miocene marine marls elsewhere in the Mediterranean[10,11]. The typical HT and LT components recognized in these marls have essentially the same direction, including an inclination error; in southern Sicily these directions also include an average 35° tectonic rotation. In addition, the observed polarity patterns result in an excellent correlation with the geomagnetic polarity timescale[2]. Apparently, both components have been acquired during or shortly after deposition of the sediment. A slightly delayed acquisition of normal remanent magnetization (NRM)—having no serious effect on the magnetostratigraphic record—may significantly disturb a record of rapid geomagnetic field changes such as occur during a polarity reversal.

Demagnetization diagrams of the different stadia throughout the lower Nunivak transition (Fig. 1) show that the two components were acquired at different times (Fig. 2). The HT component is stably reversed until −23 cm, then there is an abrupt change to normal polarity taking place within 1–2 cm (for explanation of how levels were assigned, see Fig. 2). The transition to normal polarity of the LT component is located ~35 cm above the HT reversal and is more gradual. (Note also that the LT component shows normal declinations near −15 cm, in the upper part of the lower white layer.)

A stable HT component residing in magnetite is generally taken to be primary. In a record where the HT component is

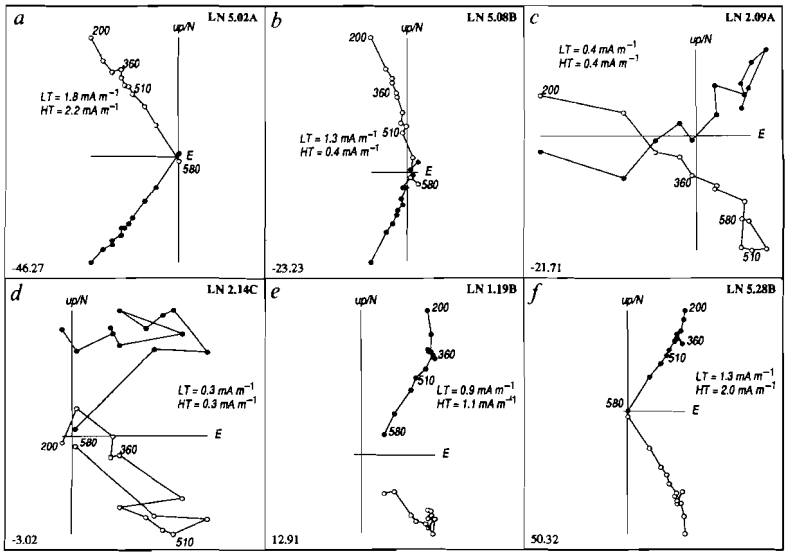

FIG. 1 Representative thermal demagnetization diagrams of the lower Nunivak (LN) reversal. Stratigraphic level in cm (left-hand axis) refers to the stratigraphic column of Fig. 2. Solid (open) symbols are horizontal (vertical) projections. Temperature steps below 200 °C are not shown in order to enhance the details at seen higher temperatures. Intensities of LT and HT are given in each diagram. Temperature steps are 200, 250, 300, 330, 360, 390, 420, 450, 480, 510, 540, 560, 580 °C. *a*, the HT and LT components are both clearly reversed and include the 35° rotation; *b* shows the same, except that 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, whereas the LT component is still clearly reversed—both components are 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 north direction, but the inclination is still very shallow; *f*, both components have approximately the same normal polarity direction, including the familiar 35° rotation, and the intensities have largely recovered to pre-reversal values.

reversed and the LT component is normal, the latter is easily interpreted to represent a (partial) overprint by secondary (sub-recent) remanences. But if the LT component is reversed and the HT component is normal, as observed in the lower Nunivak reversal record, then the LT component is not a secondary overprint. Apparently, (just) after the R-N geomagnetic field transition has occurred, the HT component records, at a certain depth below the sediment/water interface, the existing (N) geomagnetic field, whereas at that depth the LT component has already locked into the previously existing (R) geomagnetic field. Thus, there is a lag in magnetization between the two components. In the lower Nunivak record, this lag is ~35 cm, corresponding to ~7 kyr.

There appears to be a relation between lithology and (timing of) acquisition of the two components. This is illustrated by the results of the next R-N reversal, the lower Cochiti (Fig. 3). In the upper part of the record, the grey layer shows a simultaneous reversal of both LT and HT components at a stratigraphic level of 85 cm, implying no time lag. In the beige layer, between levels 35 cm and 25 cm, the HT component is normal and the LT component has already locked the prereversal (R) geomagnetic field indicating a lag of 10 cm. From 25 cm down to 15 cm the results are ambiguous: the HT component has been interpreted as reversed, but normal polarities are already found at the highest temperatures (similar to the lower Nunivak example of Fig. 1*b*). The lag may therefore be slightly larger, by as much as 10 cm. Thus, it appears that the HT component lags the LT component of 35 cm (lower Nunivak) and 10-20 cm (lower Cochiti) in the beige and white layers, where in the grey layer there is no apparent lag.

The lower Cochiti record (Fig. 3) could be interpreted as a reversal with a complex geomagnetic field behaviour: either as a R-N transition followed by a R 'rebound' or as a R-N transition preceded by a N 'excursion'. But we find it difficult to believe that the normal polarities in the beige and white layer between levels 35 and 65 cm are due to geomagnetic field changes, whereas those between levels 25 and 35 cm are clearly due to a lag in magnetization. We therefore conclude that the normal magnetizations of both the HT component (between levels 15-25 and 65 cm) and the LT component (between levels 35 and 65 cm) are caused by a delayed NRM acquisition. The absolute timing of the delay is difficult to assess, because the

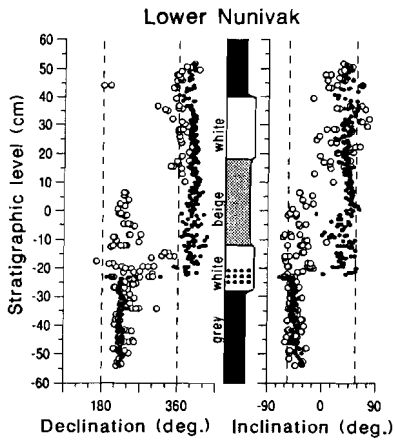

FIG. 2 Declinations and inclinations during the lower Nunivak reversal. Open (closed) symbols denote LT (HT) directions. The LT component reverses polarity ~35 cm higher than the HT component and shows more scatter. The weathering colours (carbonate contents[4]) of each sedimentary cycle are grey (70%), white (80%), beige (60%) and white (80%), whereas fresh colours show more gradual changes from dark blue to light blue. Dots in the lower white layer refer to brown oxidation spots. The 0-cm level was (arbitrarily) put at a clearly recognizable (sedimentary) level.

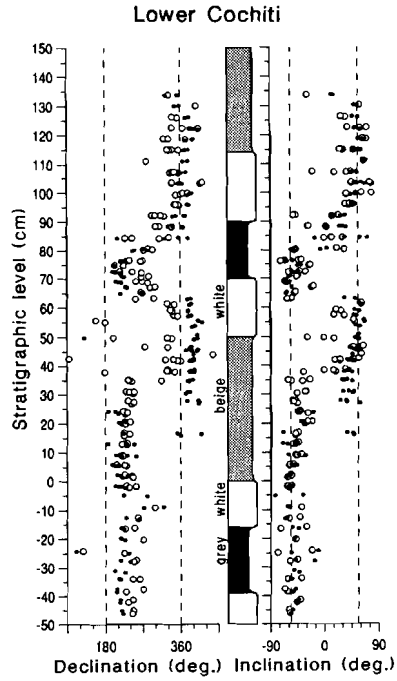

FIG. 3 Declination and inclinations during the lower Cochiti reversal. See also Fig. 2 legend. The record shows three 'polarity reversals', respectively R-N, N-R and N-R, in both the HT and LT directions.

actual reversal may take place anywhere upwards of the level where both components have consistently the same polarity (the 10-cm level in the lower Nunivak; the 95-cm level in the lower Cochiti). The stratigraphic distance between this level and the lowest level that shows a delayed acquisition, provides a minimum estimate of the absolute depth of the lag. In the lower Cochiti for instance, this amounts to an absolute lag of at least 70-80 (60) cm for the HT (LT) component, corresponding to a delay of 14-16 (12) kyr.

The origin of the LT and HT components is likely to be related to physical and chemical processes below the sediment/water interface, resulting in diagenetic alteration and authigenic formation of magnetic minerals[1]. The HT component is carried by fine-grained single-domain magnetite[6] which may be of biogenic origin (see refs 12-14). Indications of biogenic magnetite were found in late Miocene marls from Crete[12] which show exactly the same demagnetization characteristics[11] as the marls studied here. The sharp decay of the HT component between 540 and 580 °C and the corresponding narrow peak in the blocking temperature spectrum may be characteristic of ultra-fine-grained single-domain magnetite in a narrow grain size and possibly of biogenic magnetite. The origin of the LT component is uncertain. The strong decay of this component between 200 and 330-360 °C may indicate that either pyrrhotite or greigite may be a carrier of the remanence. Indications of pyrrhotite have been found in similar marine marls from Calabria[8]. Microbially mediated sulphate reduction to pyrite below the iron-reduction interval in suboxic (to anoxic) sediments may have produced intermediate authigenic forms like greigite[15,16]. On the other hand, the observed decay between 200 and 330-360 °C may also be an indication of maghemite[17], which may arise from (syn-sedimentary) low-temperature oxidation of magnetite[18,19,7] under oxic conditions. Rock-magnetic and geochemical studies in our laboratory are presently aimed at determining the mag-

netomineralogy and palaeoredox indicators.

Marine sediments have been used extensively for detailed geomagnetic reversal studies, many of which are from Mediterranean marls[8,10,20] having demagnetization characteristics very similar to the Sicilian marls. Our study suggests that, considering the still largely unknown mechanism and especially the timing of remanence acquisition, it is premature to derive characteristics of the geodynamo during reversals from such records. In the records considered here, interpretation of the HT component in terms of geomagnetic field behaviour would lead to an extremely rapid R-N reversal, taking place within 200-400 kyr. In the case of the lower Cochiti, the transition would be followed by a 'rebound' to reversed polarities. Geomagnetic rebounds are probably a real feature of the transition field because they are also observed in volcanic records[21]. But we show that rebounds recorded in (suboxic-anoxic) marine marls may well be due to sedimentary artefacts. In addition, although (oxic) deep-sea marine sediments seem to provide more reliable records of the geomagnetic field, the often cyclic nature of their rock-magnetic and geochemical properties[22-25] warrants a cautionary interpretation of transitional remanence behaviour in these sediments as well. □

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Chapter 2

Longitudinal confinement of geomagnetic reversal paths as a possible sedimentary artefact

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THE positions of the virtual geomagnetic pole (VGP) during a large number of reversals of the Earth's magnetic field seem to show a remarkable confinement to longitudes over the Americas or to antipodal longitudes[1,2], although it has been argued that this confinement is statistically insufficiently constrained[3]. It has also been pointed out[4] that the same bands of longitude appear 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, suggesting a causal relationship. Here we show that longitudinal confinement of VGPs can arise from the smoothing of non-antipodal stable directions (just) before and after a geomagnetic reversal, because of the filtering effect of the remanence acquisition process in sediments. The origin of this non-antipodality is still uncertain and must remain speculative until more reversal records become available that include sufficiently long pre- and post-transitional intervals.

Many new palaeomagnetic reversal records have been reported in recent years, mostly from sedimentary sequences. These records have served as a test for reversal models. For instance, the low-order zonal harmonics model of Williams and Fuller[5] predicts that VGP paths should be confined either to the site longitude (near sided) or to its antipode (far sided). Results[1,2] suggesting a confinement independent of site longitude either to a path over (approximately) the Americas or to antipodal longitudes over Southeast Asia and Australia raised considerable excitement in the palaeomagnetic and geomagnetic community[6], mainly because of the suggested relationship with other geophysical observations[4]. But there is reason for considerable caution in interpreting palaeomagnetic records from relatively slowly deposited sediments. Evidence from volcanic records indicates different transitional field behaviour including wide longitudinal scatter (see for example refs 7-10), although it has been suggested that particular VGP clusters recur[11]. More important, many aspects of the process of remanence acquisition and retention in sediments, such as authigenesis and (early) diagenesis, are still largely unknown. Recently it was shown that the palaeomagnetic components in two sedimentary reversal records have a delayed remanence acquisition and probably do not reflect the true geomagnetic field during a transition[12]. In addition, it was pointed out earlier[13] that smoothing due to remanence acquisition in a zone long with respect to reversal duration produces a record in which intensity changes take longer than directional changes, as is usually observed in sediments. This smoothing may further produce the typical confined VGP paths found in sedimentary records, particularly in the case of non-antipodal stable directions (just) before and after a reversal. Non-antipodality in declination typically yields VGP paths 90° W or 90° E of the observation site[3,13,14], whereas non-antipodality in inclination (inclination anomaly[15-18]) leads to a near-sided or far-sided path[5].

To test the effect of sediment smoothing, we have analysed reversals recorded in late Miocene marine marls from Crete[14] and in the Pliocene Trubi marls from Calabria[19] and Sicily (refs 12, 19, 20; and unpublished data). All records have typical sedimentation rates of 5 ± 1 cm kyr^{-1}. These records were chosen because the established magnetostratigraphies for Crete[21], Calabria[22] and Sicily[23] allow the determination of non-transitional stable directions. These are the mean directions of the entire underlying and overlying polarity zones with durations that range from ~60 to more than 400 kyr (refs 21-24); they explicitly exclude directions close (less than ~50 cm) to a transition. To determine whether there is a more pronounced non-antipodality of stable directions close to the actual transition, caused for example by geomagnetic instabilities that are related to the reversal[25-27], we have in addition determined the mean near-

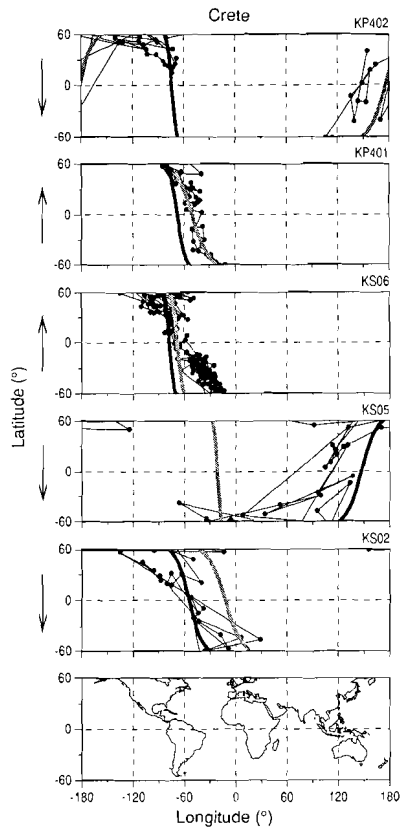

FIG. 1 Intermediate VGPs (those between latitudes 60° S and 60° N) of upper Miocene reversal records from Crete[14]. KS02 and KS05 are two older reversals, KS06 and $KP_4$01 represent the same reversal recorded in two different sections, $KP_4$02 is the youngest reversal (codes follow ref. 14). The arrow indicates the sense of the reversal; the lowermost frame presents the continents for reference. The solid line shows the VGP path obtained by filtering the mean directions of the entire under- or overlying polarity zones[21] (non-transitional directions), the shaded line represents filtering of near-transitional directions as determined from the reversal records (see also Table 1). Filtering was done by assuming an instantaneous reversal from the pre-transitional to the post-transitional stable directions (both of unit intensity) using a moving rectangular window that 'sees' an increasing portion (0-100%) of the usually non-antipodal post-transitional direction; a triangular or gaussian window gives essentially the same results[13]. The resulting intermediate directions have been used to calculate the corresponding intermediate VGPs.

transitional stable directions, where available and determined from the reversal records themselves. We have checked the (non-)antipodality of both non-transitional and near-transitional mean directions by using the reversal test[28] (Table 1), and found that in almost all cases these directions are significantly non-antipodal. We refer to the resulting filtered VGP paths (see Fig. 1) as the non-transitional model and the near-transitional model, respectively.

Caution must be used in interpreting near-transitional directions. Because of the general lack of reversal records that include sufficiently long pre- and post-transitional sampled intervals, some intervals used are less than typical post-depositional remanent magnetization (PDRM) lock-in depths[29,30] (15–20 cm), and this may introduce an unwanted bias. But even if a longer interval is used, authigenic magnetite formation may cause a chemical remanent magnetization (CRM) to be acquired down to depths of 80–120 cm (refs 12, 20). The substantial CRM acquisition depth may well bias the near-transitional directions as an average of stable polarity and (true?) transitional directions. In addition, this process may be complicated by the fact that CRM acquisition probably occurs only (or mainly) in specific depth intervals that depend on cyclically changing redox conditions[20], leading to a discontinuous smoothing filter much more complex than applied here.

For the Cretan records, it seems that the observed VGPs generally follow a path close to that determined by the non-transitional model (Fig. 1). The near-transitional model shows less agreement with the observed VGPs for the older two records (KS02 and KS05). In KS05, VGPs are initially close to the non-transitional model, but in a later stage seem to 'hesitate' between the latter and the near-transitional path. For the high-resolution records KS06 and especially $KP_4$01, however, the agreement of the near-transitional model with the observed VGPs is better. Furthermore, for the younger reversal, $KP_4$02, there is now considerably better agreement: the observed VGPs initially follow the non-transitional path over North America, but then 'jump' to the near-transitional path over Australia. This switching between antipodal VGP paths is a pattern of behaviour often seen in sedimentary records.

For the Calabrian records, the results of filtering generally agree well with the observed VGPs (Fig. 2). Although the lack of intermediate VGPs for the lower Sidufjall makes any comparison unreliable, the better resolution of the lower and upper Nunivak again provides a very good correlation between observed and filtered VGPs, either non-transitional or near-transitional. In the lower Nunivak, both observed and filtered VGPs show a near-sided path, whereas the upper Nunivak shows the 'more familiar' path over the Americas.

From the Sicilian records (Fig. 3), the lower Mammoth VGP path[19] is poorly established, but the observed VGPs still agree with both models and pass over Southeast Asia. The non-transitional model predicts the same Asian path for the upper Mammoth[19], but the observed path lies west of the Americas and is in good agreement with the near-transitional model. Our most recent and very-high-resolution records from Sicily are those of the Thvera and Nunivak subchrons (ref. 12; and unpublished data). Because of earlier observations of 'excursions' due to delayed remanence acquisition[12], care was taken to sample long and detailed records, and near-transitional directions are accurarely established. The agreement with the near-transitional model is particularly good.

The analysis of these central Mediterranean records shows that there is good agreement with the non-transitional model and an even better correlation with the near-transitional model. We have analysed in total 23 reversal records, 19 of which show intermediate VGPs in good or excellent agreement; one record is undetermined because of a lack of intermediate directions (LSC, Fig. 2), and three give VGPs that are antipodal with respect to the model. Hence, more than 80% of the records are successfully modelled by smoothing of non-antipodal directions.

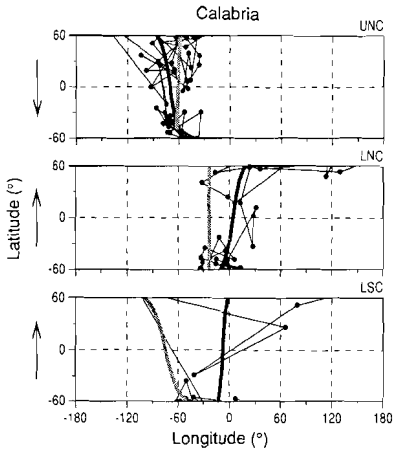

FIG. 2 Intermediate VGPs of lower Pliocene reversal records from Calabria[19]. LS, US indicate lower, upper Sidufjall; LN, UN represent lower, upper Nunivak. Post-fix C refers to Calabria. See also Fig. 1.

Furthermore, depending on the degree of smoothing, a clockwise swing in declination (an American VGP path for a reversed-normal (R–N) transition) may appear as a counterclockwise swing in the record[31] (a Southeast Asian/Australian VGP path). This could provide a possible explanation why some records show VGPs antipodal to the expected path. In addition, a differential degree of smoothing may well explain why several records such as the upper Sidufjall in Calabria[19], the lower Kaena[19] and upper Kaena[20] from Sicily, show two antipodal bands of VGPs within one and the same transition. Such behaviour is not easily explained by models involving a (non-reversing) equatorial-symmetric dipole[1] or nondipole[2] component, although it may be compatible with a reversing equatorial dipole[11]. Our analysis raises several important questions: what is the extent of smoothing and/or why are non- and near-transitional directions not antipodal? But first we discuss the simplest possible cause, an unremoved overprint.

A combined analysis of the palaeomagnetic[23] and carbonate[32] data from the Pliocene sediments on Sicily suggests that a higher carbonate content causes increased porosity, leading to increased weathering and thus to a larger secondary (normal) overprint. Rock magnetic studies showed that such an overprint may persist even to high (demagnetization) temperatures[33,34]. A similar constant unremoved overprint was reported from the Deccan traps basalts, but this may be a bias from early palaeomagnetic data[35]. It affects 'primary' (synsedimentary or near-depositional) normal and reversed directions differently, depending on tectonic tilting and/or rotation. For instance, no tilting or rotation, plus an inclination error of the primary component, will lead to non-antipodality in inclination only, whereas superposition of a secondary overprint on a clockwise rotated primary component (35° in Sicily; 15° in Calabria) leads to non-antipodality in declination with a westerly offset and hence to VGPs over America. Although this mechanism may largely explain the (high-carbonate) lower Pliocene data, it does not explain the strong easterly non-antipodality of the (low-carbonate) upper Pliocene mean directions, with a clockwise rotation of 25–35°, that show no or only a small overprint[36]. This suggests that at least here the observed non-antipodality may have a different origin.

TABLE 1 Results of reversal test

Reversal	Sense		Interval (cm) pre	post	N1	R1	N2	R2	γ	γ^*	ctmd
Crete											
$KP_1 03$	(R–N)	non	650	600	10	9.940	8	7.940	16.8	8.3	negative
		near	22	34	5	4.988	6	5.932	12.7	8.9	negative
$KP_1 02$	(N–R)	non	850	650	12	11.900	10	9.940	11.8	5.4	negative
		near	78	226	12	11.875	7	6.910	20.7	7.6	negative
$KP_4 02$	(N–R)	non	900	350	15	14.770	4	3.990	14.1	6.4	negative
		near	32	25	6	5.900	6	5.604	8.9	19.9	Class C
$KP_1 01$	(R–N)	non	1,900	850	28	27.750	12	11.900	15.8	4.8	negative
		near	87	91	25	24.810	10	9.854	29.7	6.3	negative
$KP_4 01$	(R–N)	non	3,000	900	22	21.770	15	14.770	14.3	5.6	negative
		near	39	67	10	9.838	11	10.775	33.3	9.1	negative
KS06	(R–N)	non	1,600	700	14	13.760	9	9.930	14.6	7.1	negative
		near	95	110	33	32.786	34	32.634	21.6	5.4	negative
KS05	(N–R)	non	300	1,600	5	4.970	14	13.760	2.3	7.7	Class B
		near	74	57	6	5.776	7	6.504	20.5	21.9	Indeterm.
KS02	(R–N)	non	400	700	7	6.962	8	7.980	9.8	5.3	negative
		near	26	17	11	10.769	11	9.950	10.4	16.6	Class C
Calabria											
UNC	(N–R)	non	1,000	1,000	29	28.811	16	15.128	10.4	9.4	negative
		near	8	6	19	18.261	19	18.814	37.3	7.5	negative
LNC	(R–N)	non	1,250	1,000	20	19.888	29	28.811	4.1	3.2	negative
		near	60	44	26	25.970	20	19.260	10.4	7.1	negative
USC	(N–R)	non	350	1,250	14	13.782	20	19.888	7.6	5.8	negative
		near	23	36	17	16.346	60	59.426	14.5	7.5	negative
LSC	(R–N)	non	550	350	3	2.991	14	13.782	7.6	5.8	negative
		near	43	35	16	15.387	73	71.594	13.6	8.2	negative
UTC	(N–R)	non	950	550	12	11.696	3	2.991	3.1	9.9	Class C
		near	27	30	30	28.074	52	51.380	13.9	7.5	negative
Sicily, upper Pliocene											
UKS	(R–N)	non	350	2,600	12	11.950	32	31.750	9.3	3.6	negative
		near	9	21	25	24.035	52	51.199	14.7	6.3	negative
LKS	(N–R)	non	350	350	8	7.860	12	11.950	9.2	8.2	negative
		near	26	11	15	14.674	22	20.975	5.6	9.3	Class B
UMS	(R–N)	non	450	350	9	8.922	8	7.860	8.8	9.2	Class B
		near	15	10	15	14.942	16	15.602	15.8	6.6	negative
LMS	(N–R)	non	1,150	450	19	18.923	9	8.922	7.0	5.7	negative
		near	15	15	28	27.783	23	22.418	11.7	5.4	negative
Sicily, lower Pliocene											
GGS	(R–N)	non	3,300	1,150	49	48.280	19	18.920	5.0	3.2	negative
		near	11	85	13	12.904	92	8.630	21.5	4.3	negative
UCS	(N–R)	non	450	3,300	8	7.950	49	48.280	12.3	5.9	negative
		near									
LCS	(R–N)	non	900	450	17	16.780	8	7.950	9.3	6.0	negative
		near									
UNS	(N–R)	non	650	900	12	11.930	17	16.780	10.1	5.2	negative
		near	8	25	37	35.381	94	92.511	37.3	5.6	negative
LNS	(R–N)	non	850	650	12	11.930	13	12.790	9.9	6.2	negative
		near	30	71	75	74.550	233	226.233	9.5	2.1	negative
LSS	(R–N)	non	500	400	7	6.930	4	3.980	12.5	8.6	negative
		near	15	52	11	10.748	27	26.185	27.6	8.6	negative
UTS	(R–N)	non	1,200	500	7	6.930	11	10.860	10.3	8.0	negative
		near	6	21	20	19.517	68	67.494	53.6	5.5	negative
LTS	(N–R)	non	500	1,200	11	10.860	4	3.960	7.8	10.7	Class C
		near	30	31	66	65.146	62	59.848	36.7	3.9	negative

Results of the reversal test[28] of both non-transitional (non) and near-transitional (near) stable mean directions, testing for a common true mean direction (ctmd) between the mean directions of two distributions (of N_1 and N_2 samples). The test is positive if the angle between the two means (γ) does not exceed the critical angle (γ^*); the latter depends on N_1, N_2 and the respective unit vector sums $\mathbf{R}_1$, $\mathbf{R}_2$ (refs 28, 37) of the two distributions. The classification (A, B, C, indeterminate) depends on the value of γ^*, with breakpoint at 5°, 10° and 20°. The test is negative if $\gamma > \gamma^*$. The sense of the reversal (N–R or R–N) is given. Intervals used to determine near- and non-transitional directions before (pre) and after (post) the transition are given in centimetres. Most cases show a negative test, indicating statistically significant non-antipodality; only in a few cases is an indeterminate or a B/C classification found. Abbreviations are as in Figs 1–3; $KP_1 01$, $KP_4 01$ and KS06 represent the same reversal from different sections[14], as do $KP_1 02$ and $KP_4 02$, and the same reversal may be sampled both in Calabria and Sicily (for example UTC and UTS). Near-transitional directions are taken as stable (that is, non-intermediate and approximately near-axial dipole field) directions from the actual reversal records, overlapping only slightly or not at all with the 'non-transitional' intervals. In some records, the near-transitional interval is of the order of typical PDRM lock-in depths[29,30] and the filtering results may thus be biased.

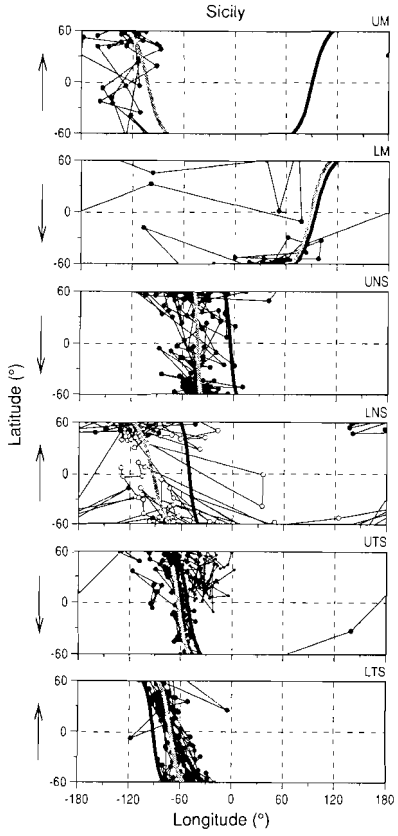

FIG. 3 Intermediate VGPs of Pliocene reversal records from Sicily (refs 19, 20 and unpublished data). LT, UT indicate lower, upper Thvera; LN, UN as in Fig. 2; LM, UM indicate lower, upper Mammoth. Post-fix S refers to Sicily. See also Fig. 1. Open symbols in LNS refer to the low-temperature component[12], as the high-temperature component (closed symbols) shows almost no intermediate directions.

Non-antipodality in inclination is well established[15-18], and sediment smoothing of this inclination anomaly may have been the cause of the many observed near- and far-sided VGP paths that prompted the Williams and Fuller model[5]. A declination anomaly is less well established[16]. From a recent analysis of outer core flux patterns[27], however, a qualitative model for the time-averaged geomagnetic field was derived that shows that none of the time-averaged normal declinations are significantly different from zero, but reversed declinations are, in some latitudinal regions in the Atlantic hemisphere. Filtering of these declinations (in our case defined as non-transitional) would lead to a predominantly westerly VGP path in the Atlantic hemisphere, in agreement with recent observations[1,2,4] and with most of our own observations (Figs 1-3). Furthermore, even the significant CRM acquisition depths cannot explain the non-antipodality of polarity zone mean directions that represent the average of usually much longer intervals (3-30 m; Table 1).

The good fit of the near-transitional model may have a similar (non-antipodality) origin, but may also be related to the fact that it 'sees' a considerable part of the transition. Although the wide longitudinal scatter observed in volcanic records[3,7-10] seems to provide no reason for the observed confinement, it has been argued that particular VGP clusters (in South America and Australia) may recur[11]. Indeed, smoothing of such a temporal feature would bias the resulting magnetizations[11] so that they would often agree with our observations. Although similar VGP clusters (near South America) are seen in some of our 'continuous' sedimentary records (for example UNC, Fig. 2; UNS, Fig. 3), the complex and 'discontinuous' CRM filter prevents a firm assessment of their validity.

Apparently, the information that we may derive from sedimentary records (which are probably biased, possibly continuous, but often insufficiently long) and from volcanic records (discontinuous, possibly unbiased, but too few) is still far from complete[38]. □

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Chapter 3

A paleomagnetic and geochemical record of the upper Cochiti reversal and two subsequent precessional cycles from Southern Sicily (Italy)

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A detailed paleomagnetic and geochemical study of the upper Cochiti (N-R) reversal recorded in marine marls on southern Sicily shows two consecutive and very rapid transitions (R-N and N-R) that coincide with distinct lithological boundaries. A 'Fe-migration model' is presented in which magnetite is formed under different diagenetic conditions. During sulphate reduction originally formed ('primary') magnetite is preserved. This primary magnetite has recorded the pre-transitional normal polarity. After burial, migration of ferrous iron occurs and subsequent oxidation produces 'secondary magnetite' that records the post-transitional reversed polarity and produces the two apparent transitions. In addition, low (LT) and high (HT) temperature components record the reversal in a slightly different way, the HT component being acquired with a delay. Probably, both components reside in magnetite but have a different origin and grain-size distribution, giving different blocking temperature spectra. The actual upper Cochiti reversal occurs at a level with an estimated astronomically calibrated age of 4.165 Ma.

1. Introduction

In recent years, an increasing number of detailed records of geomagnetic polarity reversals have been published and they have provided important information on the characteristics of polarity reversals. The majority of these records have been obtained from sedimentary sequences because of their abundance relative to lava sequences and their advantages of appropriate time control and continuous registration of the geomagnetic signal. However, there is reason for considerable caution in interpreting reversal records from relatively slowly deposited sediments. Not only because of evidence from volcanic records which often show quite different transitional field behaviour, but also because many aspects of the mechanism of remanence acquisition in sediments are still poorly known. Indeed, recently it could be shown that a high temperature magnetite component - in two examples of marine sedimentary reversal records from Sicily - has a delayed remanence acquisition with respect to a low temperature component and probably does not reflect the true geomagnetic field during the reversal [1]. The magnetic minerals are unlikely to be detrital because the changes in direction take place at distinct lithological boundaries in the sediment. Hence, the mechanism for the delay as suggested by Tucker [2;3] is invalid in the sediments under study. Van Hoof and Langereis [1] suggested that the magnetic minerals carrying the components were authigenically formed under different and cyclically fluctuating paleoredox conditions. Most likely, early diagenetic processes play an important role in the complex remanence acquisition mechanism.

To contribute to our understanding of this mechanism, additional and detailed rock magnetic and geochemical investigations are clearly needed.

To this end, the study of reversal records is particularly useful because the (rapid) change of the geomagnetic field to an opposite polarity provides the opportunity to get temporal constraints on the delay of the remanence acquisition. Our current research concerns the reversals records - from the lower Thvera to the Gauss/Matuyama boundary - that are located in the Rossello composite section, which consists of the marine marls from the Trubi and the Narbone formation [4;5]. These sediments show a distinct cyclicity in both $CaCO_3$-content and weathering (induration) profile, a cyclicity which is related to the precessional cycle of the Earth's orbit [4;6].

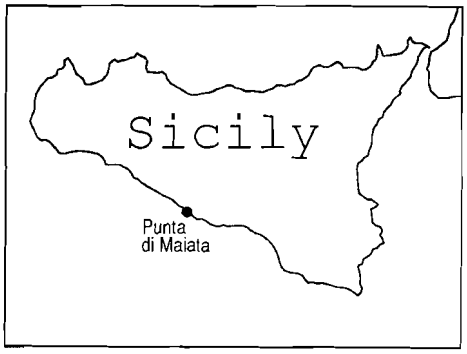

Figure 1. Location of the Punta di Maiata section in Southern Sicily (Italy).

In this paper, we present the paleomagnetic, rock magnetic and geochemical results of the upper Cochiti (UC) polarity reversal, sampled in the Punta di Maiata section on Southern Sicily (fig. 1). In addition, we have sampled two subsequent precessional cycles in order to investigate the magnetic and geochemical changes due to a cyclically changing environment outside the transition. Finally, since the remanence is not simply of detrital origin, we introduce a model that accounts for the acquisition of the remanence during early diagenesis. In this model, the changes in the directions of the remanence are thought to be caused by changes in paleoredox conditions in the sediment, as derived from the geochemical and rock magnetic data.

2. Geological setting and sampling

Punta di Maiata is a small prominent cape in a series of cliffs along the south coast of Sicily (fig. 1). The Punta di Maiata section forms the middle part of the Rossello composite section [4;5] and consists of marine marls of the Pliocene Trubi formation; the bedding plane has a strike and dip of 354° N and 13° E. The average sedimentation rate in this section can accurately be determined and proves to be remarkably constant: 4.5 to 5.5 cm/kyr. The Trubi marls consist mainly of carbonates (60 to 80% $CaCO_3$) and a mixture of clay minerals, and they show a pronounced rhythmic bedding which is characteristic for this formation on Sicily (fig. 2). Small scale sedimentary cycles are quadripartite and show a distinct grey-white$_{(1)}$-beige-white$_{(2)}$ colour layering [4]. The average thickness of these cycles - in which the grey and beige marls represent the less indurated, $CaCO_3$-poor beds - is approximately 1 metre. The (midpoints of) individual grey layers have been correlated to individual minima of the precession index [7], providing a high resolution astronomical polarity time scale [8]. Larger scale sedimentary cycles can be distinguished on the basis of the cyclic recurrence of relatively thick and/or indurated marly intervals in the succession [4;6]; they can be correlated to the eccentricity cycle with periods of 100 and 400 kyr [8]. Although the weathering profile shows quite sharp changes in colour and induration, the changes in the fresh, unweathered sediment are much more gradual. Often, the bottom part of the white$_{(1)}$ layers shows brown oxidation spots, even in the freshest possible marls. These spots appear to be important with respect to both paleomagnetic and geochemical properties, as will be discussed later.

The samples were taken on the basis of the magnetostratigraphy of the Punta di Maiata section [5]. The section contains the upper Sidufjall reversal boundary, the Nunivak and Cochiti subchronozones and the Gilbert-Gauss reversal boundary. The stratigraphic interval in which the original upper Cochiti reversal was found, was sampled in detail over a stratigraphic length of 120 cm. The two additional quadruplets were sampled over an interval of 220 cm. The record comprises the complete cycles 50, 51 and 52 (fig. 2); [8]. The 0-level was defined at the boundary between the grey layer and the first white layer of cycle 50 (fig. 2). Hence, the entire stratigraphic interval sampled ranges from -40 to 80 cm for the upper Cochiti reversal record, and from 80 to 300 cm for the two additional quadruplets. Due to the topography of the chosen sampling locality it was not possible to take samples below -40 cm. Furthermore, the lowermost part (-40 to 0 cm) consisted of fresh and clearly unweathered sediment, but - despite our efforts to remove the weathered surface - in the middle and upper part the sampling of slightly weathered intervals could not always be avoided. Especially in the white layers - with the exception of white$_{(2)}$ of cycle 49 - the sediment kept its weathering colour, whereas in the beige layers the original light-blue colour of

the marls could mostly be reached.

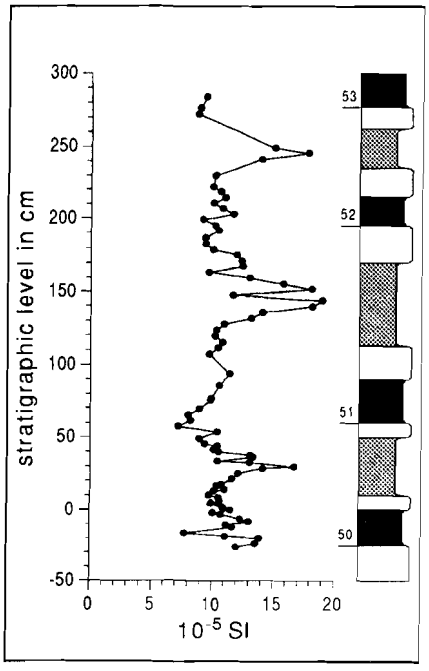

Figure 2. Susceptibility and lithology of the upper Cochiti (UC) record. Small scale sedimentary cycles are quadripartite and consist each of grey(black) - white(white) - beige(shaded) - white(white) coloured marls. Cycle numbers are according to Hilgen (1987; 1991) and are denoted just left of the grey layer. Clear maxima of susceptibility are found in the beige layers, probably due to the higher clay content and the corresponding higher paramagnetic contribution.

Sampling was done by taking oriented cores of 25 mm diameter more or less parallel to the bedding plane at very close intervals (<1 cm), from a freshly cut near-vertical plane. In the laboratory, cores were cut into standard specimens of 22 mm height. The stratigraphic position of each specimen was accurately determined by taking into account the drilling orientation, strike and dip of the bedding plane and width of the saw cut.

For geochemical studies, the samples were cleaned by carefully scraping about 1 mm from the surface, followed by grinding and homogenising the sample in an agate mortar. For determination of major and trace element concentrations, 250 mg split of each sample was digested in a $HClO_4/HF/HNO_3$ mixture. After digestion, the final solutions were made up in 50 ml 1N HCl. Analyses of the total element concentrations of Al, Ti, Mn, Ca, S, Zr and Fe were done with an ARL34000 ICP emission spectrometer. Carbonate content was calculated from the total Ca concentrations, with a correction of 2% for non-carbonate Ca: $CaCO_3 = (Ca_{total} - 2)* 2.5$. The accuracy and precision of the analysis was monitored by replicate analyses of international and laboratory standards and was found to be better than 2% for the major elements Al, Fe, Al, Mn, Ca and better than 3% for the elements S, Zr, Ti.

3. Rock magnetic and chemical properties

The cyclicity in the Trubi marls is related to the precessional cycle which causes changes in climate - by variations in seasonal contrasts - and thus in the concentration and nature of terrigeneous (clay and other minerals) and biogenic (carbonate) material supplied to the sea bed. The rock magnetism of the Trubi marls in general have been extensively studied by van Velzen and Zijderveld [9;10]. They found magnetites to be the most important carrier of the natural remanent magnetisation (NRM); small amounts of hematite may be present but it is unlikely that this mineral contributes to the remanence. In addition, some goethite is found in these sediments [9;10;11]. In order to distinguish magnetic changes caused by geomagnetic variations from those caused by lithological variations and early diagenetic processes, a number of geochemical and rock magnetic properties of the sediment have been determined. The rock magnetic properties concern the initial susceptibility (X_0), remanent saturation magnetisation (J_{rs}), the remanent coercive force (H_{cr}) and their interparametric ratios.

3.1 Rock magnetic results

X_0 is strongly concentration dependent, and it also depends on the grain size and type of magnetic mineral. The X_0 of the three precessional cycles (fig. 2) is not significantly higher in grey than in white, but it shows prominent maxima in the beige layers.

IRM acquisition curves of a number of samples from different lithologies in the UC reversal were determined (fig. 3a,b). Samples from the white and grey layers are almost (92 - 96%) saturated after application of a 300 mT field, and are fully saturated after application of a 400-500 mT field. Samples from the beige layers are only saturated after application of the highest fields (1.5 to 2.0 T) because of the presence of some additional high coercivity mineral, like goethite/hematite. Subsequent thermal demagnetisation of the J_{rs}

(fig. 3c) does support the presence of some goethite in the beige layer: the decay of the saturation remanence at 100-120°C - the maximum unblocking temperature of goethite - is slightly larger for the beige layers than for the grey and white layers. More significant, however, is the decrease of the intensity at 135 - 155 °C, which is probably caused by the reduction of stress in superficially maghemitised SD magnetite grains [10]. The observed maximum blocking temperatures support the presence of magnetite as the dominating magnetic mineral.

fine grained magnetite [12;13]. The H_{cr} values are significantly lower for stratigraphic levels between -30 and -15 cm, i.e. in white$_{(2)}$ and the lower part of the grey layer, and highest from -10 to 0 and from 40 to 60 cm. H_{cr} values tend to increase if some weathering has occurred, which may explain the higher values between 40 and 60 cm, but this cannot explain the high values in the upper part of the grey layer, which showed very fresh sediment during sampling. Further, the relatively low H_{cr} values in the beige layer give no clear indication for a high coercivity mineral, as could be deduced

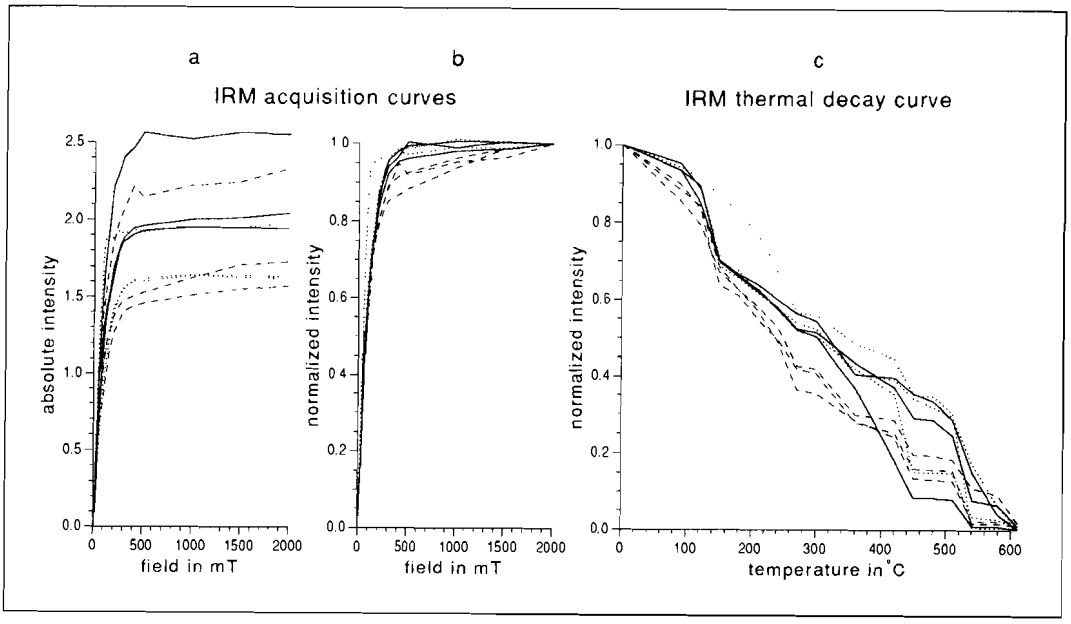

Figure 3. IRM acquisition curves and susbsequent thermal demagnetization of the saturation IRM for different lithologies from the UC reversal record. Grey, white and beige layers are denoted by solid, dotted and dashed lines, respectively. Grey and white layers reach (almost) saturation at 200-300 mT, whereas beige layers are not even totally saturated in the highest fields.

The J_{rs} decreases with increasing grain size but - like the susceptibility - is not a good grain size indicator, but depends mainly on the concentration and type of the magnetic material. The J_{rs}/X_o ratio, however, may eliminate - if only one magnetic mineral is dominant - the concentration dependence of J_{rs} and X_o. In the case of magnetite the ratio increases with decreasing grain size. The J_{rs}/X_o values in the grey and white layers (15-25 kA/m; fig. 4) support magnetite but the ratio is lower in the beige layer (10-15 kA/m).

The remanent coercive force H_{cr} is largely independent of the concentration of magnetic material and can therefore be used to discriminate between different magnetic minerals. H_{cr} values are generally 50-70 mT which is somewhat high for

from the IRM acquisition curves (fig. 3). Apparently, the contribution of magnetite (85-95% in fig. 3) strongly dominates the remanent coercive force.

The $H_{cr}/(J_{rs}/X_o)$ ratio may give useful information on both magnetic mineralogy and grain size. The values for this ratio (generally 2 - 2.5, but higher in beige: >3) are somewhat high for fine grained magnetites. The deflection of the $H_{cr}/(J_{rs}/X_o)$ and the J_{rs}/X_o ratios from the values for fine grained magnetites is probably not due to a significant change in magnetic minerals, but rather to a change in the concentration of clay minerals. The paramagnetic contribution of the clay minerals in X_o is some tens of percents with maxima in the beige layers [11]. A correction for

this paramagnetic contribution of the $H_{cr}/(J_{rs}/X_0)$ and the J_{rs}/X_0 values will change these ratios into values more - but not entirely - compatible with fine grained magnetites. Likely, changes in magnetomineralogy occur that are related to the different (beige) lithology, but these changes remain largely unnoticed due to the paramagnetic dominance.

The grey layers show some sulphur spikes of 0.2 - 0.5% (fig. 5), and some sulphur is found in the lowermost part of the record. Everywhere else, the concentration of sulphur is below the detection limit. Although some weathering has occurred, these sulphur enrichments are probably relict Fe-sulphides, as is indicated by coinciding iron peaks; Fe-sulphides are authigenically formed during sulphate reduction shortly after deposition [14]. The highest sulphur content is found in the grey layer of cycle 50 (0.5%). This suggests that within this cycle diagenetic circumstances were the most reducing during the formation of the grey layer, rather than during the formation of the white and beige layers. The same may apply to the grey layers of other cycles. The increased subaerial weathering in the middle and upper part of the record, as was observed during sampling, however, has very likely resulted in the oxidation of most sulphides in the grey layers of cycles 51 and 52, causing the lower observed sulphur concentrations.

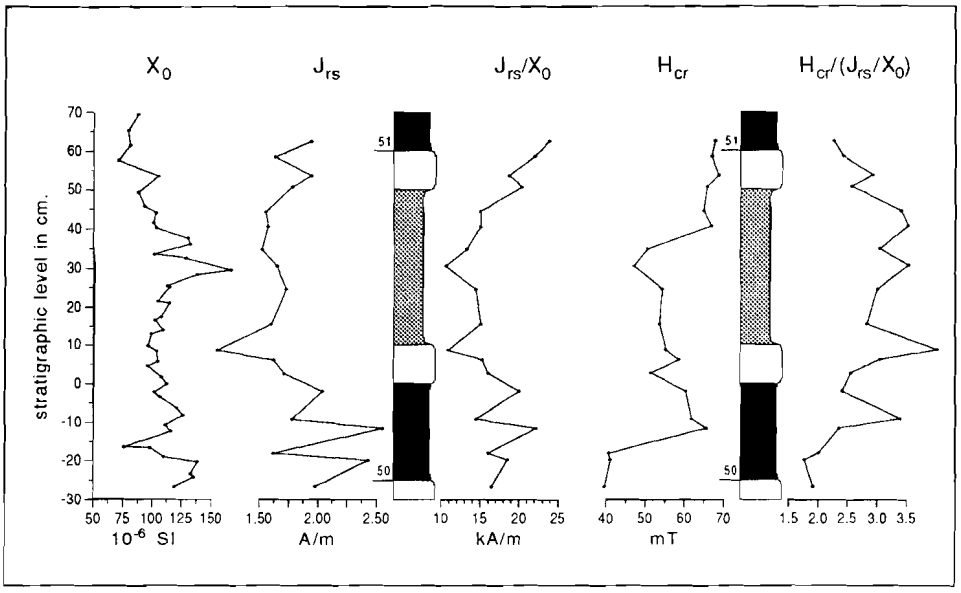

Figure 4. Rock magnetic properties for the UC reversal record, involving susceptibility X_0, saturation remanence Jrs, their ratio X_0/J_{rs}, the remanent coercive force H_{cr}, and their interparametric ration $H_{cr}/(J_{rs}/X_0)$. See text for further discussion.

3.2 Geochemical results

Relatively high concentrations of carbonate (65-70%) are found in the white layers, intermediate values in the grey layers (60-65%) and low values in the beige layers (55%) (figure 5). The white$_{(1)}$ beds have slightly higher carbonate values than the white$_{(2)}$ layers.

Mainly due to the high Ca-carbonate content of the sediment, the concentrations of all major elements show a negative relation with Ca concentrations (closure effect) (figure 5). We have assumed a more or less constant non-carbonate input, as is suggested by rather constant Zr/Al and Ti/Al ratios - although the Ti/Al ratio seems to be slightly higher in the beige layers (fig. 5). Therefore, all concentrations were divided by the Al-content of the samples, in order to compensate for carbonate dilution.

Manganese is a good indicator for paleoredox conditions [15]. Manganese concentrations are generally low and display values of 600-800 ppm in the grey and beige layers that have a low to moderate carbonate content, but may go up to 1600 ppm in the carbonate-rich white$_{(1)}$ layers. The general distribution of Mn seems closely rela-

ted to the carbonate content and may well be determined by manganese adsorption and subsequent overgrowth on calcite surfaces [16,17,18]. This correlation is clearly demonstrated by the relative constant Mn/Ca ratio, but the white$_{(1)}$ layers display a large enrichment of Mn over Ca. The higher Mn/Ca ratio in the white$_{(1)}$ layers compared with the white$_{(2)}$ layers (fig. 5) suggests that more dissolved manganese was available during or after the formation of the white $_{(1)}$ layers.

be explained by the higher concentrations of Fe and, possibly, higher concentrations of Ti (figs. 4, 5). The higher Fe-content - even when corrected for Al - may be caused by an increase of (Fe-containing) clay minerals, rather than to a higher concentration of (ferri)magnetic minerals [20]. Similarly, the higher Ti-concentration may reside in Ti-magnetites (or Ti-maghemites) - also causing a higher susceptibility - although from the rock magnetic properties no clear evidence for Ti-

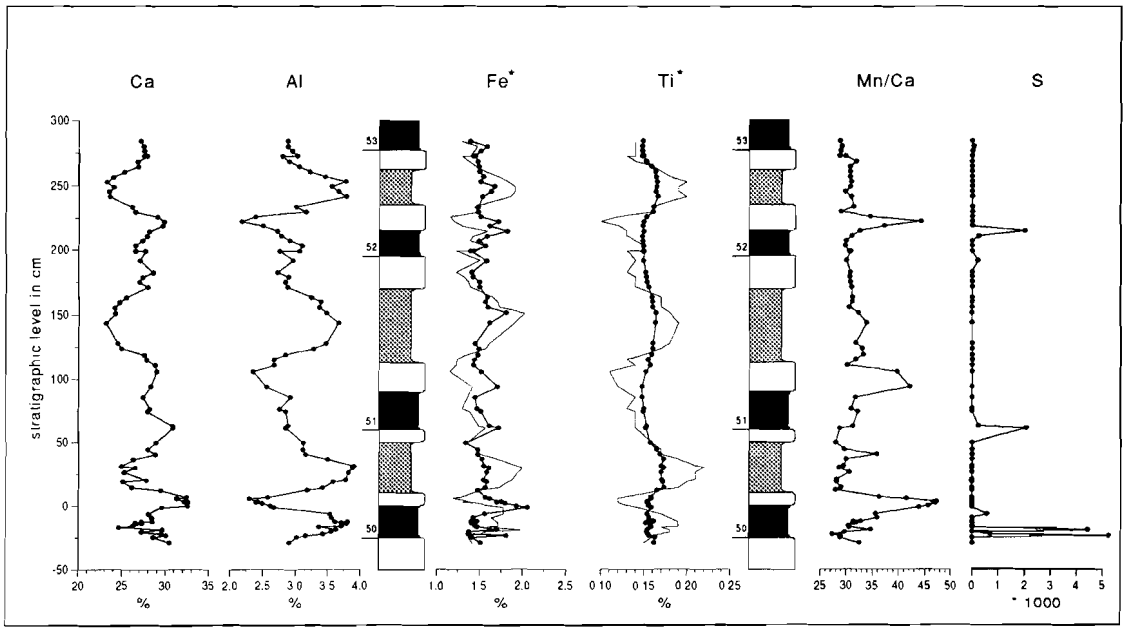

Figure 5. Abundances of the most relevant elements Ca, Al, Fe (thin line), Fe (Fe/Al, thick line), Ti (thin line), Ti* (Ti/Al, thick line). Mn/Ca and S. The high Ca-carbonate content causes the concentrations of all major elements to show a negative relation with Ca concentrations (closure effect). We have assumed a more or less constant non-carbonate input, as is suggested by rather constant Zr/Al and Ti/Al ratios - although the Ti/Al ratio seems to be slightly higher in the beige layers. The grey layers show some sulphur (S) spikes of 0.2 - 0.5%, everywhere else the concentration of sulphur is below the detection limit.*

The organic carbon content has not been measured in these samples. In a series of quadruplets from the same Punta Maiata section, values were found to vary between 0.05% and 0.25% [19]. The organic carbon content is highest in the grey, intermediate in the white and lowest in the beige layers. At younger stratigraphic levels in the Trubi formation, the grey layers are laminated and they are often developed into diatomitic or sapropelitic layers [4], typical for a higher organic carbon content.

The differences in lithology - and thus in chemical composition and magnetic mineral content of each layer - are reflected in the geochemical and rock magnetic parameters. Initial susceptibility values are higher in the beige layers, which can

magnetites is found. Any existing Ti-magnetites may be masked, however, and not contribute to the remanence: large grained Ti-magnetite was found in magnetic concentrates of the Calabrian (southern Italy) Trubi formation and was shown not to carry any noticeable remanence [21]. In addition, other mineral phases, such as rutile and ilmenite, dominate the Ti concentration.

This enrichment in iron-rich minerals strongly influences the rock magnetic properties which are directly related to the concentration of magnetic minerals, like X_0 and J_{rs}. It should further be realised that these properties - although showing a linear relation with concentration - are also influenced by changes in grain size and crystallinity of the magnetic minerals.

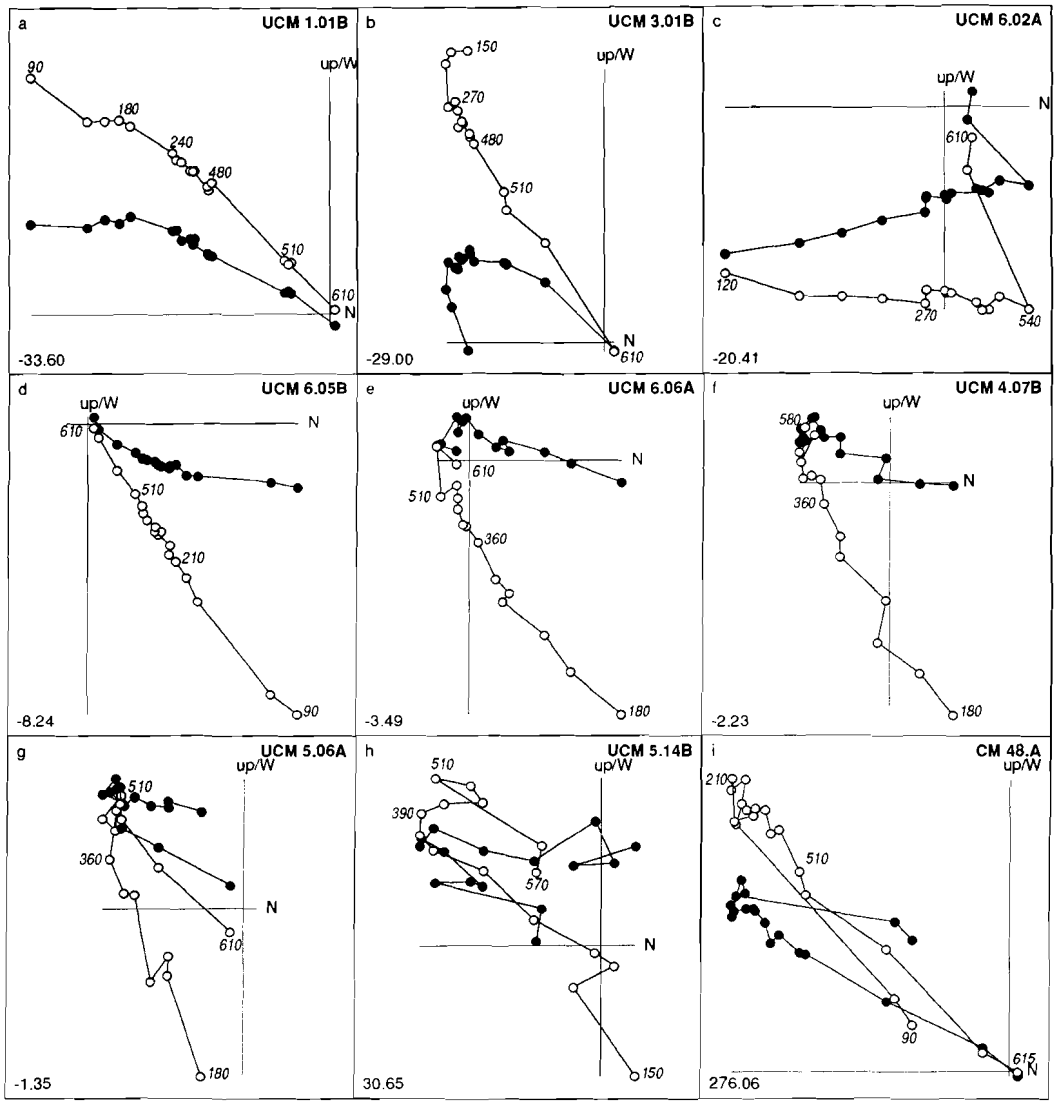

Figure 6. Thermal (th) demagnetization diagrams of 9 representative samples from the UC reversal and the two additional quadruplets. Open (closed) symbols denote projection on the vertical (horizontal) plane. The stratigraphic level in cm is plotted down-left; numbers refer to temperatures in °C; tc means that tilt correction has been applied.

4. NRM components

4.1 Demagnetisation

Thermal demagnetisation of the natural remanent magnetisation (fig. 6) generally shows the presence of three components, as usually found in these Trubi marls [5;11]. Apart from a small laboratory induced component removed at 90-100°C having an orientation related to storage, there is often a first component that has a present-day field direction and is removed at 200-250 °C. It is clearly of secondary origin and must be due to weathering, since this component does not occur in the

lowermost and least weathered part of the record (fig. 6a).

A characteristic remanence is generally removed at higher temperatures and consists of two components. A low temperature (LT) component is removed between 250°C and 360 - 450 °C. In some samples, the direction of the remanence starts to fluctuate at temperatures higher than 390 °C (fig. 6h). These fluctuations are due to viscous behaviour, likely caused by the production of superparamagnetic magnetite grains during heating through oxidation of pyrite [10]. Although in most of the record sulphur is below the detection limit, trace amounts of pyrite are probably sufficient to produce a (magnetically) detectable amount of magnetite. A high temperature (HT) phase is usually removed at 580 °C and occasionally at somewhat higher temperatures (610 °C). The HT component is most probably carried by single domain magnetite [9]. The LT and HT components have both - where possible - been determined for every sample. Poorly determined directions as a result of very low intensities and/or scatter have not been plotted in figure 8.

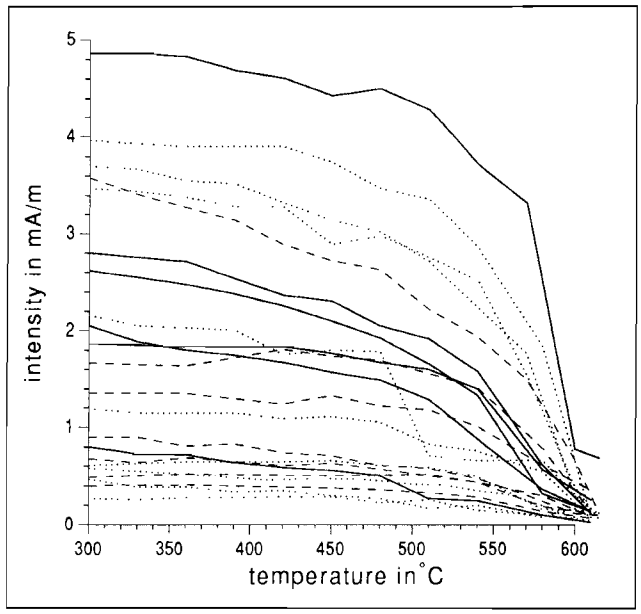

Figure 7. The NRM decay curves - starting at 300 °C - of beige (dashed), white (dotted) and grey (solid) marl samples. Intensities are generally lower for the beige marls than for the grey and white marls. Maximum blocking temperatures are indicative of (cation deficient) magnetite.

The demagnetisation diagrams further show the same clockwise rotation as observed in the other Trubi sections on Sicily [5;22]. The characteristic remanence directions are marked by consistently shallower inclinations than the geocentric axial dipole field for southern Sicily, likely caused by sediment compaction [23;24].

Partial blocking temperature spectra (>300 °C) for the different lithologies have been determined throughout the section (fig. 7). They show a similar trend for all lithologies in that the NRM decreases most rapidly between 510 and 600 °C. The decay curves clearly indicate the presence of magnetite which is probably slightly cation deficient considering the maximum unblocking temperatures that are higher than 580 °C [21;25;26]. In general, the beige layers show appreciably lower NRM intensities, in agreement with the rock magnetic and geochemical data.

4.2 The reversal record

The directions of the LT and HT components for the upper Cochiti record are shown in figure 8. For the earlier magnetostratigraphic study of the Punta di Maiata section [5], most samples were taken in the grey and beige layers. The reversal, therefore, was thought to occur between the grey and beige layer of cycle 50. It now appears, however, to be a more complex reversal showing reversed directions in the white$_{(2)}$ layer of cycle 49. The mean directions for the reversed and normal intervals are

	N	Dec	Inc	α95	Rsum
reversed: below -27.5 cm	7	214.1	-39.2	9.8	6.847
normal: -24.1 to -4.4 cm	31	29.7	54.8	3.9	30.591
reversed: above 26.3 cm	78	213.8	-41.3	2.3	76.382

where N is the number of samples, dec, inc are declination, inclination, α_{95} represents the cone of confidence at the 95% level, and Rsum is the vectorial sum of N unit vectors. The mean declinations are essentially the same as the general 35° clockwise rotation of the Caltanisetta basin [22].

The first (R - N) transition is characterised by an instantaneous polarity change of the HT component at -25 cm, at the boundary of white$_{(2)}$ to grey, with no (reliable) intermediate directions. The LT component shows a similarly quick change in declination, whereas its inclination seems to reverse more gradually. Obviously, the LT reversal takes place at a somewhat higher level (-25 to -15 cm). The HT polarities remain normal in the interval from -25 to -5 cm, and the average

declination is only slightly less than the generally observed 35°, but inclinations are a bit steeper than usual.

observed in the upper Kaena reversal record [20].

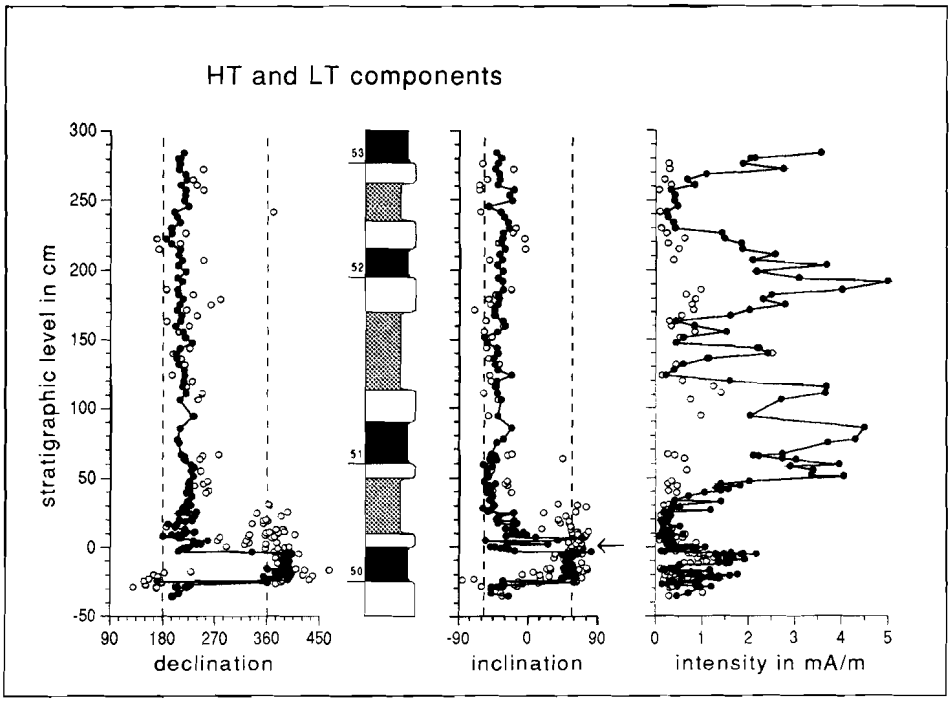

Figure 8. LT (circles) and HT (dots) components of the UC reversal record and two subsequent precessional cycles, and their respective intensities. Diamonds denote directions that could not reliably be interpreted due to very low intensities, random scatter and/or significant viscous behaviour at higher temperatures. Arrow indicates where the maximum in Mn was found.

The second transition (N - R) is at level -5 cm, just below the boundary of grey to white$_{(1)}$. The HT component again changes very rapidly from normal to reversed polarity with no intermediate directions. After this sharp transition, the HT component shows a 'normal excursion', especially in inclination. Alternatively, one could interpret the sudden change to reversed polarity as a short 'reversed excursion' (between -5 and 5 cm), only just before a more gradual change from normal to reversed, say from 5 to 15 cm. Remarkably, this 'reversed excursion coincides exactly with the peak in Fe* and Mn/Ca (fig. 5); the latter ratio is a good paleoredox indicator as mentioned above.

The LT component changes polarity at distinctly higher levels, approximately between 30 and 35 cm, and coinciding with a relatively small but significant peak in Mn/Ca (fig. 5). The relatively small directional changes in the HT component above 26 cm are compatible with secular variation changes smoothed by sedimentary remanence acquisition, and are similar to those

5. Discussion

The initial location of the reversal was based on the magnetostratigraphy of the Punta di Maiata section [5], for which samples were taken mainly from the grey and beige layers. Hence, the reversed polarities in white$_{(2)}$ of cycle 49 were then not observed. Since the beige layer of cycle 49 shows normal polarities [5], this implies that there must be yet another (N-R) transition, recorded somewhere near the transition of beige to white$_{(2)}$ in cycle 49.

It is clear that the recorded transitions are not caused by a rebound or excursion related to the UC reversal, but are more likely caused by early diagenetic conditions that vary with the lithology. Therefore, interpretation in terms of timing and duration of the two recorded transitions (while a third transition must occur below) is not meaningful. Early diagenetic changes cause one and the same reversal to be recorded several times, *via* a

delayed acquisition mechanism [1]. In addition, the coincidence of two extremely rapid reversals that occur exactly at distinct lithological or geochemical boundaries - apparently related to paleoredox boundaries - makes any interpretation in terms of geomagnetic field behaviour improbable. The assertion that sediments have a resolution determined by the most rapid changes recorded [27] is therefore not generally valid, even though the sediment may have no visible (lithological) boundaries. We thus emphasise the need for additional and detailed lithological (and rock magnetic, geochemical) information in reversal records. Because of the very rapid changes of the magnetic remanence no intermediate directions are recorded and therefore we refrain from presenting VGP paths. It is more useful to discuss the paleoredox conditions and corresponding diagenetic changes that may have caused the observations.

5.1 Diagenetic changes during deposition

The alternating sequence of beige (carbonate poor with iron (hydro)oxides), white (carbonate rich) and grey layers (richer in organic matter than beige and white, and containing sulphides, fig. 5) suggests that the diagenetic conditions changed from oxic *via* suboxic to anoxic for the respective layers [28;29]. The differences in diagenetic stages are caused by different amounts of reactive organic matter in each layer, i.e. a higher amount of organic matter will lead to an enhanced consumption of oxidants used for the degradation of organic matter. After oxygen is depleted, nitrate, manganese-oxides, iron hydroxides and finally sulphate will be used respectively as electron-acceptors [29;30].

Bacterially mediated magnetite formation has reported to occur under oxic and suboxic conditions [31], whereas anoxic conditions inhibit bacterially mediated magnetite formation due to substrate competition with sulphate reducing bacteria. This implies that during the deposition of the white and beige layers magnetite formation continued, while during the deposition of the grey layer magnetite formation was inhibited in the anoxic part of the sediment. The unlaminated appearance of the grey layer suggest that conditions were not fully anoxic during deposition. Therefore, sulphate reduction occurred not at the top of the sediment. Above the zone of sulphate reduction, iron hydroxides, manganese oxides, nitrate and even oxygen may have been used as oxidants for the decomposition of organic matter and thus suboxic conditions may have developed during the deposition of the grey layer, causing the formation of magnetite (at time t=1 in figure 9).

During burial, magnetite dissolution may occur in the sulphate reduction zone [32]. But because of the higher reactivity towards HS^- from respectively dissolved iron, amorphous and poorly crystalline iron - like lepidocrocite, goethite and hematite - magnetite may persist in this zone. Magnetite is only attacked after all other iron-oxide phases are "titrated" with HS^- [33;34]. In addition, magnetite is resistant to microbial iron reduction [35;36]. Evidently, the IRM acquisition, as well as the IRM and NRM thermal decay curves imply the presence of magnetite and thus its preservation during sulphate reduction (figs. 3, 7).

Magnetic minerals that could form during sulphate reduction are greigite and pyrrhotite [37]. However, greigite and pyrrhotite would normally not survive oxidation, for example because of sub aerial weathering. The formation of greigite and pyrrhotite has been reported to occur mainly in a low sulphate environment, and thus they are mainly found in fresh and brackish water sediments [38]. Moreover, greigite and pyrrhotite are not found in recent anoxic basins [39, 1981; Morse and Cornwell, 1987). Conclusively, chemical and rock magnetic results indicate that magnetite is the main carrier of the magnetic signal and that it is not significantly dissolved during sulphate reduction in these sediments.

5.2 Processes after deposition

Shortly after deposition of a grey anoxic layer (at time t=2 in figure 9), the organic carbon flux diminishes and sulphate reduction ceases. As a result, the dissolution of magnetite by HS^- will also stop. This situation will cause a surplus of oxygen over the reductive species in the subsequently deposited organic-poor beige and white layers. Reductive species in these layers (organic matter and possibly sulphides) will be oxidised and the oxidation front will move down relative to the sediment-water interface. This formation of a progressive oxidation front is similar to that described in the "burn-down" model developed by Wilson et al. [40] and is best marked by the redox-sensitive elements Fe and Mn. Diffusion of dissolved Fe and Mn from the reduced zone towards the oxygen boundary can cause the formation of iron and manganese enrichments - amorphous Fe and Mn (hydr)oxides - at the top of the reduced layer [41-43]. The iron enrichments found above the grey layers in the bottom part of the white$_{(1)}$ layers (Fig. 5) are clearly the result of this process. The oxidation-front can penetrate only slowly into the grey layer because this contains high levels of

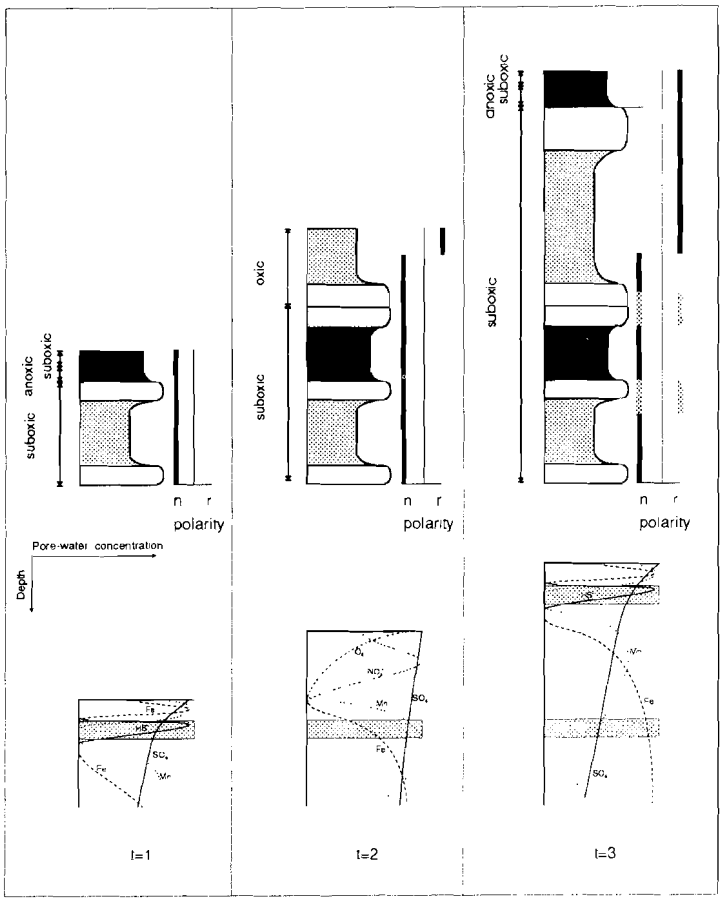

Figure 9. The 'Fe-migration model' with the pore water concentrations vs. depth at the time of deposition: during deposition of the anoxic grey layer 'primary' magnetite is formed at t = 1, and during post-oxic diagenesis upon burial by the white and beige layers 'secondary' magnetite is formed by migration of Fe^{2+} (at t = 2 and t = 3). Primary magnetite causes the geomagnetic field to be recorded during or very shortly after the initial formation of this magnetite, by growing through its critical blocking diameter and causing acquisition of a CRM. Upon burial of the grey layer, secondary magnetite is formed on either side of the grey layer (in the white layers), in the grey layers this depends on its redox circumstances. If at that moment a polarity reversal occurs, the secondary magnetites will record the new polarity. The presence of already formed primary magnetite - having recorded the old polarity - competes with the secondary magnetite (dashed polarities). If secondary magnetite predominates, the resulting direction will reflect the new polarity, in our case reversed. Lithology as in figure 2.

organic carbon and of sulphides. In addition, the oxidation of pyrite by still available ferric iron in the grey layer will increase the upward flux of Fe^{2+}. This occurs following the reaction [44;45]:

$$FeS_2 + 14Fe^{3+} + 8H_2O \rightarrow 15Fe^{2+} + 2SO_4^{2-} + 16H^+ \quad (1)$$

This additional supply of reduced species will hamper the descent of the oxidation front and will enhance the iron enrichment at the top of the grey layer and in the superimposed $white_{(1)}$ layer.

5.3 Processes after burial

After deposition of the white and the beige layer on top of the grey layer, a new cycle begins with the deposition of a new $white_{(2)}$ and a grey layer (at time t=3 in figure 9). An increased flux of organic matter during the formation of the new grey layer will merely stop the downward diffusion of oxygen. Subsequently, below the newly deposited grey layer oxygen will become rapidly exhausted. However, the low content of organic matter in the beige and white layers between the

new and the buried grey layer will not support sulphate reduction. The diagenetic stage between sulphate and Mn reduction coinciding with a low reactive organic carbon content is called post-oxic following the classification of Berner [39] and can be compared with suboxic conditions [30;40].

In the buried grey layer pyrite oxidation probably continues, although oxygen is exhausted. The remaining ferric iron in this buried grey layer (iron hydroxides) enhances the oxidation of pyrite, providing a continuous source of Fe^{2+}.

Karlin [46] found that magnetite is formed under suboxic conditions between the nitrate and iron reduction zone. Bacterially mediated magnetite formation can occur under such conditions [36;47]. Dissimilatory iron-reducing micro-organisms can form magnetite provided that reactive ferric iron is present [36]. Therefore, magnetite formation will be especially enhanced at the very reactive iron enrichment zone formed during burn-down just above the buried grey layer (the 'reversed excursion'; see section 4.2), but may also occur to a lesser extent just below the grey layer. The previously formed iron hydroxides are an ideal substrate for dissimilatory micro-organisms to produce magnetite (Fig. 9, t = 3).

Within the buried grey layer itself, sulphate reduction and subsequent pyrite formation have diminished the amount of available reactive iron (see previous section). Obviously, magnetite formation is not supported in the buried grey layer. Magnetite can precipitate directly in the surrounding white and beige layers following the reaction:

$$2Fe^{3+} + Fe^{2+} + 4H_2O \rightarrow Fe_3O_4 + 8H^+ \qquad \text{(II)}$$

or bacterially mediated following:

$$CH_2O + 24Fe(OH)_3 \rightarrow 2CO_2 + 54H_2O + 8Fe_3O_4 \qquad \text{(III)}$$

Magnetite formation takes place whenever the organic matter content and the dissolved oxygen concentration are low and Fe^{2+} and reactive iron-oxides are available.

The preservation of magnetite is also enhanced by non-steady state (organic-rich versus organic-poor) diagenesis. As mentioned above, an increasing flux of organic carbon - i.e. deposition of a new grey layer, - causes an upward shift of the redox front. Such a shift can occur more rapidly than iron and manganese oxides at the redox front may dissolve. Therefore, these oxides can survive [46]. Similarly, Finney et al. [15] found a correlation between the preservation of manganese oxide peaks and an increased organic carbon flux. However, persisting suboxic conditions will eventually dissolve all manganese-oxides [43;48; 49;50]. Not only the coprecipitation of Mn with calcite but also the adsorption/recrystalisation of Mn on calcite can induce the formation of Mn enriched crusts [16;17]. Previously formed manganese-hydroxides can consequently be preserved during sub-oxic diagenesis. The peaks in the Mn/Ca observed in the white$_{(1)}$ layers above the grey layers are the result of this process. The presence of Mn-enrichments in carbonate-rich rocks can indicate a change in environmental regime and thus the formation of magnetite during diagenesis. Straightforward evidence for the preservation of magnetite is given by the presence of the iron enrichments at the top of the grey layers just below the Mn enrichments as is shown by the Fe/Al ratio's (Fig. 5). The interval of -15 to +15 cm is a classical example of such sequence of these redox-mobile elements, indicating their mobilisation in the grey layer followed by upward flux to the overlying white/beige layer [43]. The preservation of this diagenetic Fe-layer and magnetite indicates that the redox conditions after the formation of this layer did not reach the iron-reduction stage.

5.4 Implications and conclusions

From our discussion above it follows that magnetite can be formed in different stages of diagenesis: during deposition primary magnetite is formed (Fig. 9, t = 1) and during suboxic diagenesis secondary magnetite (Fig. 9; t = 3). Primary magnetite causes the geomagnetic field to be recorded during or very shortly after the initial formation of this magnetite, by growing through its critical blocking diameter and causing acquisition of a chemical remanent magnetisation (CRM). This process is very similar to that described earlier by Channell et al. [51] in Italian sections where secondary hematite acquired a delayed remanence due to an authigenic CRM at some depth below the interface. The depth at which this occurs - presumably in the iron reduction zone - may vary with lithology but is probably comparable with post-depositional remanent magnetisation (PDRM) lock-in depths (15-20 cm). In fact, magnetite grains acquiring a CRM will probably be subject to typical PDRM lock-in depth intervals where CRM carrying grains are mechanically fixed, mainly through compactional loading [52]. This would imply a minimum lock-in depth of approximately 15 cm. A maximum lock-in depth, however, depends on the depth of CRM acquisition which may be well below maximum PDRM lock-in depths. Indeed, our recent work shows possible depths as much as 80-120 cm [1;20].

Upon burial of the grey layer, secondary magnetite is formed on either side of the grey layer (in

the white layers), depending on the amount of reactive iron. If at that moment a polarity reversal occurs, it will record the new polarity. The presence of already formed primary magnetite - having recorded the old polarity - competes with the secondary magnetite. If secondary magnetite predominates, the resulting direction will reflect the new polarity, in our case reversed. At the same time, intensities are much lower (fig. 8) in the white layers as a consequence of the vectorial sum of both recorded opposite directions/polarities. Thus besides a lower geomagnetic field intensity during a transition, the mechanism of opposite directions may also cause lowered NRM intensities. Further, there is an additional complication represented by the presence of both a LT and HT component. As observed earlier [1], there is a difference in acquisition between the two components, the HT component showing a delay with respect to the LT component. In other words, the LT component has already locked the old polarity (normal, e.g. between 0 and 25-30 cm), while the minerals carrying the HT component are still being formed and have not yet grown through their critical blocking diameter. However, also the LT component itself shows a delayed CRM acquisition, as evidenced by the reversed polarities in the lowermost part of the record (fig. 8). Thus, the LT component may record the ambient polarity before the HT component, but it is also subject to a 'secondary' formation. Obviously, this yields some information on the magnetic minerals carrying the LT component. Contrary to our earlier suggestions [1] we find it unlikely that it is carried by pyrrhotite or some other magnetic iron-sulphide. Firstly, the temperature trajectory that determines the LT component has much higher blocking temperatures than the Curie temperature of pyrrhotite (±320 °C) [53]; sometimes it is only totally removed at temperatures of 510-540 °C (fig. 6). Secondly, detailed rock magnetic studies of the Trubi marls from Sicily have shown no evidence for pyrrhotite [9;10]. Also our present geochemical data and discussion above virtually exclude the existence of pyrrhotite. Therefore, we assume that also the LT component is residing in magnetite. The different - and varying - blocking temperature spectra of the LT and HT components probably reflect different grain-size spectra, that in turn reflect their (authigenic) origin and timing of formation. The earlier lock-in of the LT component argues for a larger grain-size than that of the HT component, i.e. the LT magnetite has grown through its critical blocking diameter before the HT magnetite. An alternative explanation for the lower blocking temperatures of the LT component - grain-sizes near the threshold of single domain and superparamagnetic grains - is unlikely since it would imply partial dissolution of the earlier locked-in LT magnetite and the simultaneous growth/formation of secondary (HT) magnetite. Some unexplained details in the record, such as the occurrence of reversed LT directions in the bottom part of the grey layer leads us to suspect that the actual processes are even more complicated. Clearly, additional rock magnetic studies attacking this particular problem are needed.

The sequence of events suggested by our current 'Fe-migration model' - primary magnetite preserved in the grey layer and formation of secondary magnetite especially in the white$_{(1)}$ layer - supplies us with a quite accurate location of the actual reversal. The grey layer of cycle 51 consistently shows reversed polarities, suggesting that at least here the geomagnetic transition has been completed. We retain as our best and most consistent estimate the level where both LT and HT directions consistently show a reversed polarity, i.e. approximately at the 35 cm level.

In principle, this level gives a more accurate age than the one previously established 4.18 Ma [8] on the basis of the initial magnetostratigraphy [5]. The 35 cm level gives an age of 4.165 Ma, taking into account a lag of 3-4 kyr between precessional forcing and climate response [54].

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PART 2

RECORDS OF POLARITY TRANSITIONS

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Chapter 4

The upper and lower Thvera sedimentary geomagnetic reversal records from southern Sicily

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ABSTRACT

Detailed paleomagnetic records of the upper and lower Thvera polarity transitions have been determined from Pliocene marine marls in southern Sicily. The dominant magnetic mineral is fine grained magnetite. The two transitions have VGP paths following a great circle passing South America and the west coast of North America. There are strong indications that the VGP paths result from smoothing of the non-antipodal stable directions before and after the transitions. The upper Thvera transitional record is preceded by two excursions and followed by a third one. The first excursion is a sedimentary artefact caused by post-depositional migration of magnetic minerals and the third one by is caused by weathering of the sediment. The upper Thvera transition from Sicily is compared with the record from Calabria, about 250 km away. The two records show similarities as well as differences: both transitions have identical VGP paths, but in Calabria the transition is recorded lower in the sediment, the first and third excursions have different characters and the second one is not present. Apparently the registration of transitions of the Earth's magnetic field in sediments is strongly influenced by smoothing and diagenetic processes after deposition.

1. Introduction

The Earth's magnetic field has reversed its polarity many times in the geological past, but the mechanism which causes this feature still remains largely unresolved. An aid in solving this problem is the detailed paleomagnetic study of the behaviour of the Earth's magnetic field during these transitions. Detailed records of geomagnetic transitions have been obtained from different sources, the most important of which are those recorded in lava sequences and those in sedimentary sequences. Although sedimentary sequences concern appropriate time control and continuous registration of the geomagnetic signal, the recording mechanism in sediments is still badly understood and may easily lead to sedimentary artefacts [3–6]. However, Laj et al. [7] and Tric et al. [8] concluded, from studies of the VGP paths of several transitions (showing a wide distribution in both time and place) that the VGP paths are predominantly confined to a longitudinal band over the Americas or its antipode, and that there may have been a strong dipolar field during transitions. However, Valet et al. [9] pointed out that the transitional fields cannot be dipolar and that the longitudinal confinement found by Laj et al. [7] and Tric et al. [8] is statistically insufficiently constrained. Mary and Courtillot [10] found that, in many reversals, the reversing fields decrease and subsequently increase along the dipole direction on which a random noise is superposed. In addition, Langereis et al. [6] have shown that the longitudinal confinement of the VGP paths from sedimentary records can also be explained by a process of smoothing of the stable non-antipodal directions before and after the transitions. Clearly, many more transitional records from both lavas and sediments are required before there is a consensus on this matter.

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One important aspect when making a contribution to the number of known and (highly) detailed records is to ensure that the reversal(s) are well identified. This makes it possible to compare the same reversal from different geographical locations, and from different sedimentary (or volcanic) environments. In this paper we present the detailed reversal records of the lower and upper boundaries of the Thvera subchron from marine marls in southern Sicily. The results of five successive transition records from the Lower Pliocene in Calabria, in the Gilbert Chron, which range from the upper Thvera to the upper Nunivak, have already been reported by Linssen [2]. For the present work the upper Thvera transitional record from Sicily, together with the record of the lower Thvera, were sampled in considerably more detail and both are presented here.

2. Geological setting and sampling

The Thvera subchronozone in the Gilbert Chronozone was identified by a detailed magnetostratigraphic study of the Eraclea Minoa section in the Caltanisetta basin of Southern Sicily (Fig. 1) [11]. The Eraclea Minoa section forms the basal part of the Rossello composite section [12] and it consists of marine marls of the Pliocene Trubi Formation (Fig. 2); the bedding plane at the sampling locality has a strike and a dip of 267°W and 14°N. The average sedimentation rate can be accurately determined on the basis of the astronomically calibrated polarity time scale (APTS) [13] and is 5.0 cm/kyr in the lower Thvera section and 4.4 cm/kyr in the upper Thvera section. The marls of the Trubi formation mainly consist of carbonates (60–80% $CaCO_3$) and a mixture of clay minerals [14]. Hilgen [15] recognized a long succession of small-scale sedimentary cycles—the so-called quadruplets. The weathering profile of these quadruplets form a repetition of grey, white, beige and white coloured beds (Fig. 2), which were deposited during cyclic sedimentation periods of approximately 19–23 ka. These sedimentary cycles are clearly related to the precessional cycle of the Earth's orbit [16]. Although the weathering profile shows quite sharp changes in colour and induration, the changes in fresh, unweathered sediment are much more gradual. Considerable effort was taken to remove the weathered surface (up to 1 m) in order to expose the fresh (dark grey, light and dark blue) sediment. This method proved successful for almost the entire intervals sampled, except for the top part of the upper Thvera interval where the uppermost white layer remained visibly and strongly weathered. In the lower Thvera interval the colour layering was not as clear as in the higher part of the section. Only a greyish and a beige layer could be distinguished, while between these two layers there was a somewhat lighter coloured (white?) layer.

Sampling was carried out by taking oriented cores, 25 mm in diameter, more or less parallel to the bedding plane at very close intervals (~ 1 cm), from a freshly cut, near-vertical plane. Each core was divided into specimens 22 mm long. The

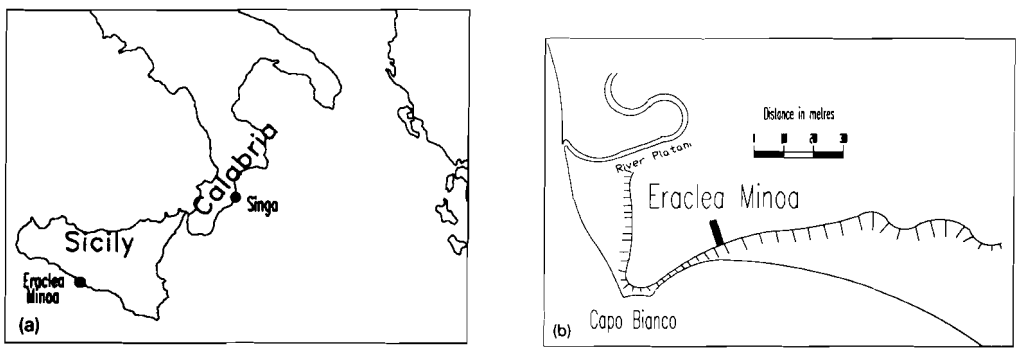
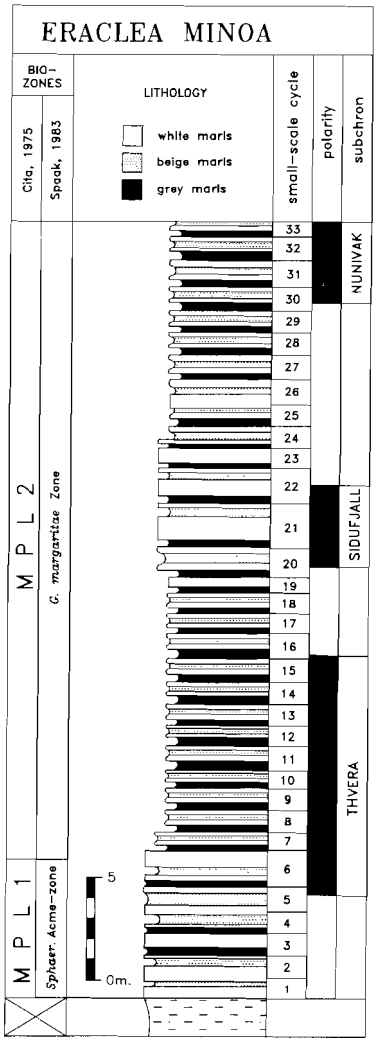


Fig. 1. (a) Map of Sicily and Calabria. The transitions described in this paper were sampled at the Eraclea Minoa Section. The upper Thvera transition will be compared with the same transition sampled in the Singa section in Calabria, some 250 km away from Eraclea Minoa. (b) Location of the Eraclea Minoa section in Sicily (Italy).

stratigraphic position of each specimen was accurately determined by taking into account drilling orientation, bedding plane and width of the saw cut. The resolution of these records is of the order of a few millimetres; hence it is better than 100 years. Variations in the parameter within a sample will, therefore, in principle only smooth the high frequency signal of the secular variation [17].



3. Rock magnetism

The acquisition of a stable remanence may take place at a certain depth below the sediment–water interface, resulting in a time lag between the deposition of the sediment and the acquisition of the remanence. Usually this is referred to as a post-depositional detrital remanent magnetization (pDRM) and the corresponding time lag is directly related to the lock-in depth, which, in turn, depends for a major part on sediment compaction. A typical value of the lock-in depth for relatively slowly deposited sediments (1–8 cm/kyr) is approximately 16 cm [18]. In the case of authigenic (biogenic) formation of magnetic minerals, however, a chemical remanent magnetization (CRM) may be acquired at depths or in depth intervals that depend on redox conditions rather than mainly on compaction. The "lock-in depth" or depth lag may then be considerably larger [5]. Post-depositional diagenetic processes controlled by changing redox conditions, may, in addition, lead to the migration of Fe ions, which can cause a CRM at a depth of more than 1 m below the sediment–water interface [1,16]. Rock magnetic parameters may give an indication of the change in the character and concentration of the magnetic minerals. Therefore, we determined the initial susceptibility (χ_0) and the (remanent) saturation magnetization (J_{rs}, J_s) and the (remanent) coercivity (H_{cr}, H_c).

The acquisition of an isothermal remanent magnetization (IRM) for a number of samples from different lithologies up to a maximum DC field of 2 T (Fig. 3a) reveals that there are deflec-

Fig. 2. Lithostratigraphy and magnetostratigraphy of the Trubi formation at the Eraclea Minoa section [12,15]. The lower Thvera transition was identified in small-scale cycle 5 and was sampled in detail over an interval of 1 m. The upper Thvera transition was identified in small-scale cycle 16 and was samples over 1.75 m. In this paper the zero levels of the two transition records are arbitrary, and were chosen at pronounced layer-parallel sedimentary lines. For the lower Thvera this zero level was chosen at a line in the beige layer of cycle 5. The zero level of the upper Thvera record is at the top of the gray layer of cycle 15. The white marls have a higher carbonate content, while the grey and beige marls are relatively poor in carbonate [16], probably due to increased continental run-off (grey) or increased African wind-blown input (beige) [14].

ion points at fields of 100–200 mT, indicating that a low coercivity mineral has reached the saturation remanence. The increase in the IRM in higher fields, of mainly the white lithology, records, in addition, the presence of a high coercivity mineral. The saturation IRM (J_{rs}) at 2 T was thermally demagnetized. The relative intensity decrease at 120°C as a percentage of the initial J_{rs} of the upper Thvera section is shown in Fig. 3b. The samples with a relatively large amount of highly coercive minerals during IRM acquisition are from the same sedimentary interval (mainly white marls) as samples with the strongest decay at 120°C. This suggests that the highly coercive, low unblocking temperature mineral is goethite, which is a typical product of weathering. A strong indication of the presence of goethite is the brown and strongly oxidized layer at level 135 of the upper Thvera section. At this level the decrease in the J_{rs} intensity between room temperature and 120°C is at a maximum of 40% (Fig. 3b). From the NRM demagnetization results, any remanence carried by goethite is not evident.

J_{rs} seems to show a lithological dependence. Correction for the amount of goethite is made by subtracting the part of the J_{rs} demagnetized at 120°C from the total J_{rs} (Fig. 4). This correction lowers the J_{rs} values somewhat but the lithology dependence remains. There are maxima in the beige layers, and minima in the lower part of the grey layers and in the white layers above the grey layers. The maxima are up to 3.5 A/m for the upper Thvera section and 1.5 A/m for the lower.

As observed in the entire section [cf., 1], the initial susceptibility (χ_0) is strongly dependent on the lithology; the beige layers consistently show maxima (here the maximum is 250×10^{-6} SI), while minimum values are found at the transition from white to grey.

The ratio J_{rs}/χ_0 is largely independent of concentration (provided that the dominant magnetic mineral is magnetite) and it may give an indication of the grain size. The ratio shows typical

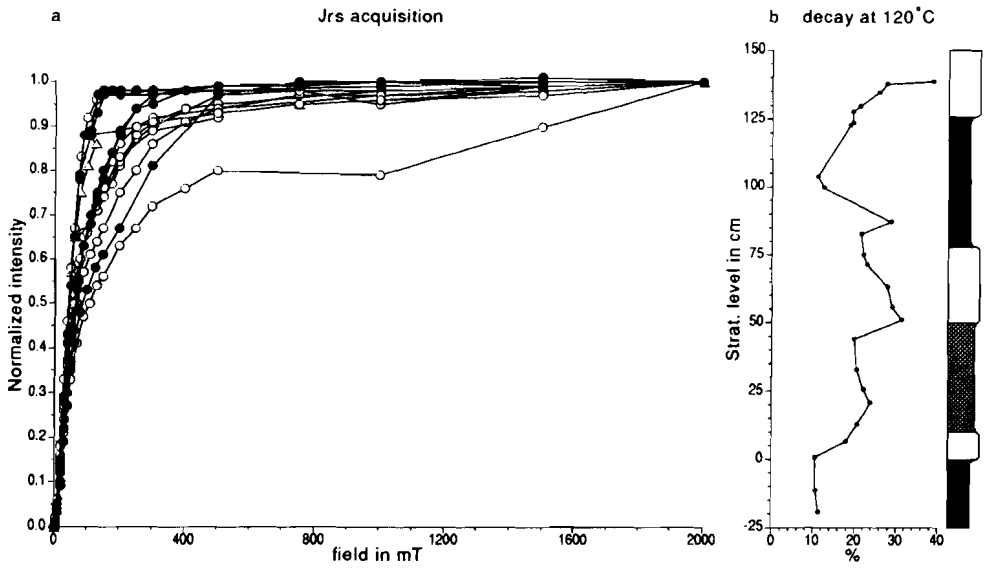


Fig. 3. (a) Normalized IRM acquisition curves of samples from different lithologies, up to a maximum field of 2T. Most samples from the grey (black points) and beige (triangles) lithologies and one from the white (circles) lithology are saturated at fields lower than 400 mT. The rest of the white samples and one grey sample are saturated in higher fields. (b) During thermal demagnetization of samples from the upper Thvera section with a saturation IRM (J_{rs}) the relative decrease in intensity up to 120°C is determined as a function of stratigraphic level. The samples having the strongest decay got their saturation remanence at higher fields, which indicates the presence of goethite, especially at level 135, which was a brown coloured part of the lithology.

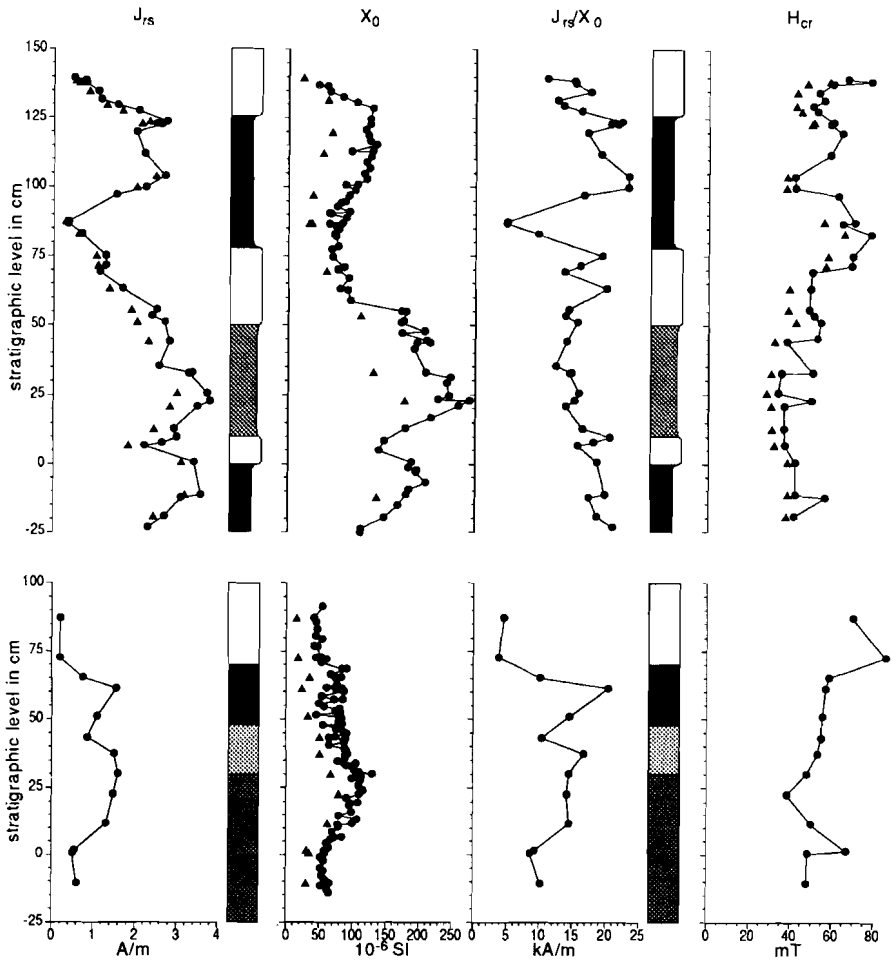


Fig. 4. Variation in the magnetic parameters saturation IRM (J_{rs}), initial susceptibility (χ_0), the ratio J_{rs}/χ_0 and remanent coercive force (H_{cr}) as functions of the stratigraphical levels and lithology. The upper part of the figure is the upper Thvera section; the lower is the lower Thvera section. The legend of the lithological column of the upper Thvera section is the same as in Fig. 2. In the lower Thvera section the differences between layers were hard to distinguish. The colours in this column are, from bottom to top: beige, whitish, greyish and white. Triangles in the J_{rs} diagrams denote the intensities of J_{rs} at 120°C. At this temperature it is assumed that only goethite fraction has been demagnetized. Triangles in the χ_0 diagrams are the corrections for the high field susceptibility. In the H_{cr} diagram the triangles denote the high coercive mineral correction (goethite), see also text. J_{rs} and χ_0 have maxima in the beige parts of the lithology. The ratios of both parameters lie in the range of fine grained magnetite. The remanent coercive forces H_{cr} lie in the range of fine grained magnetite and maghemite.

values of 15–20 kA/m, except for a clear minimum in the bottom part of grey. Fine-grained magnetites have values larger than 20 kA/m [19] and they are presumably of single domain size [20,21]. The clay fraction of the lithology has a strong paramagnetic contribution and will increase the bulk susceptibility, χ_0. Hysteresis loop experiments quantify the amount of the paramagnetic contribution (Fig. 5). Correction of χ_0 for the paramagnetic susceptibility increases the ratio

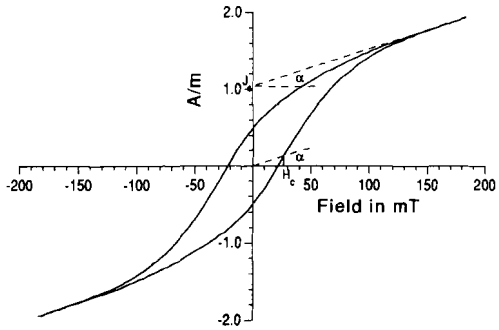


Fig. 5. Hysteresis loop. The linear trend at higher fields (dashed lines) is due to the paramagnetic susceptibility of the clay minerals. It can be derived from the curve as the tangent (of α). Horizontal axis = inducing field; vertical axis = induced remanence; J_s = saturation remanence determined 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.

J_{rs}/χ_0 to values typical of fine grained magnetites.

The remanent coercivity, H_{cr}, in natural sediments is independent of the concentration of magnetic material and is not influenced by paramagnetic clay minerals. Assuming that there is only one mineral present, H_{cr} is measured by a stepwise increase in the DC field in a direction opposite to J_{rs}, where H_{cr} is the DC field strength required to decrease J_{rs} to zero. However, in a mixture of a low coercivity mineral (magnetite) and a high coercivity mineral (goethite), a correction for the goethite must be made. The J_{rs} of the mixture, gained at a 2 T field, will consist of the IRM's of both minerals. While determining the H_{cr} of the low coercivity mineral occurring in this mixture, the DC field opposite to J_{rs} will not decrease the IRM of goethite. Therefore, the DC field must be increased until the IRM intensity of goethite is reached, instead of the zero intensity for a sample containing only one mineral. In order to obtain an approximation of the goethite fraction, we assume that the IRM intensity of goethite is the part of J_{rs} that is demagnetized between room temperature and 120°C. Typical H_{cr} values of fine grained magnetites are 40–60 mT [20,22,23], which are the values observed in the upper Thvera section (Fig. 4). After the goethite correction the data show a better fit to the fine-grained magnetite values. In the lower Thvera section the H_{cr} data are somewhat high for fine-grained magnetite values between levels 0 and 75, but here no goethite correction was made.

In their rock magnetic study of the Trubi marls from Eraclea Minoa [20], using the ratios H_{cr}/H_c and J_{rs}/J_s, Van Velzen and Zijderveld concluded that the magnetic minerals were dominated by fine grained (SD) magnetites. A small discrepancy between their data and those from Dunlop [23] was explained by the presence of some goethite, and possibly some super paramagnetic magnetite. Our H_{cr}/H_c and J_{rs}/J_s data are the same as those of Van Velzen and Zijderveld [20] (Fig. 6).

On the basis of the rock magnetic properties we conclude that, in spite of a presumed dependence of the rock magnetic parameters on lithology, the dominant magnetic mineral is fine grained magnetite throughout the sedimentary

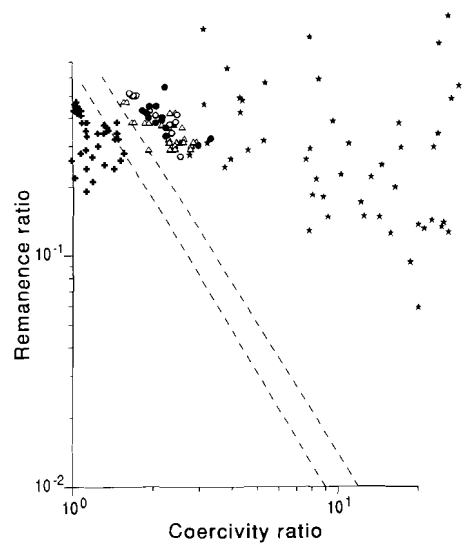


Fig. 6. Double logarithmic plot of coercivity ratio H_{cr}/H_c versus remanence ratio J_{rs}/J_s after [20]. Literature data for magnetite of known grain sizes fall on a single trend indicated by two dashed lines [23]. Stars = data from goethite; crosses = data from pyrrhotite [33]; circles = data from this study, which fit the values from the Trubi sediments (triangles) by van Velzen and Zijderveld [20] very well. These authors concluded that fine SD magnetites were the most important magnetic minerals in the sediment.

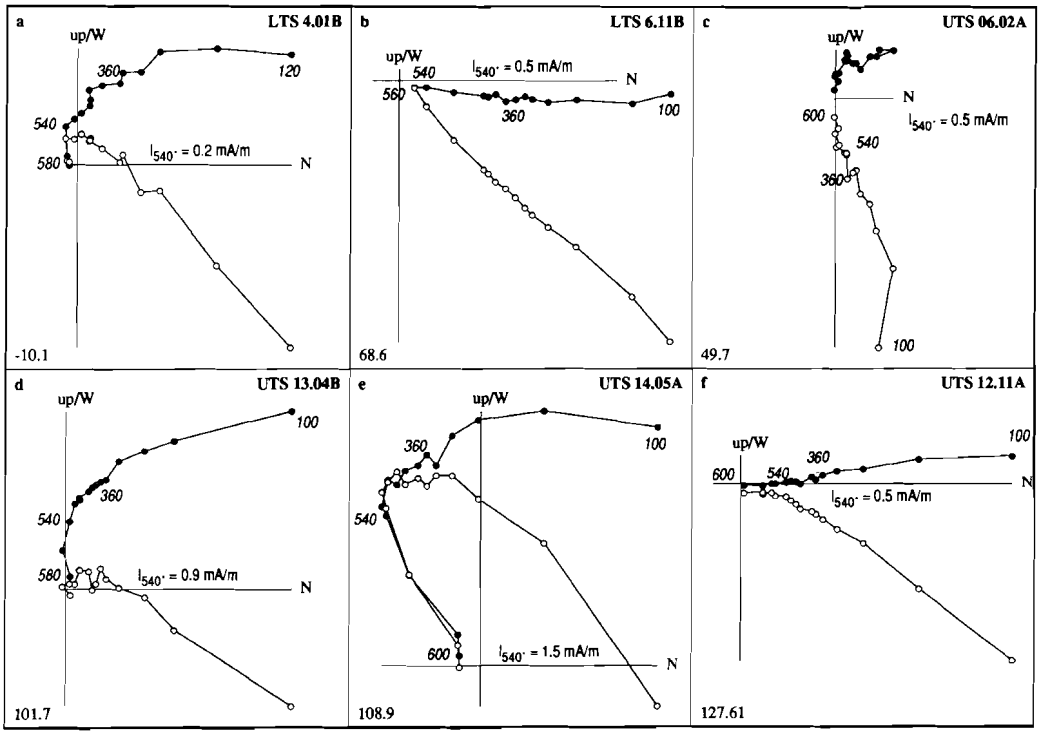


Fig. 7. Thermal demagnetization diagrams. (a) and (b) Lower Thvera. (c)–(f) Upper Thvera. (c) and (d) Intermediate directions. Up to temperatures of 250°C the directions are normal. Between 300 and 480°C the decay is small. The maximum unblocking range is from 580°C to 600°C, indicating that magnetite is the carrier of the ChRM.

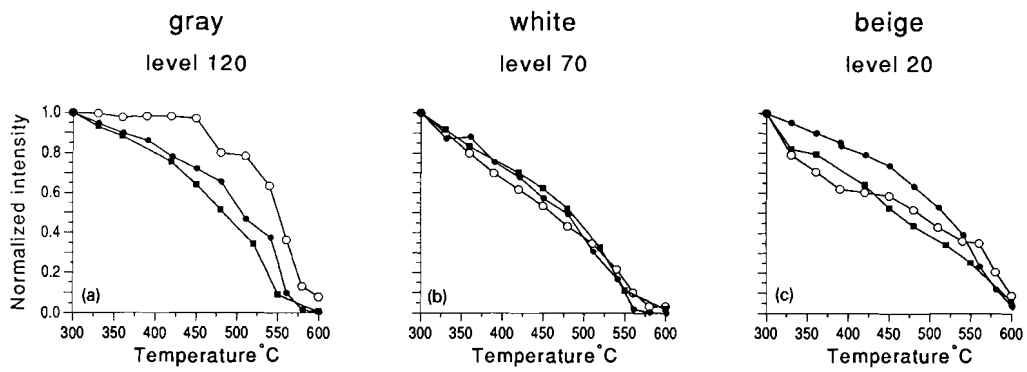


Fig. 8. Normalized thermal demagnetization decay curves of ChRM (circles), IRM(2T) (squares) and ARM (dots) for (a) level 120 (gray); (b) level 70 (white); and (c) level 20 (beige) from the upper Thvera section. Intensities are normalized with respect to 300°C, after removal of the secondary component. Generally, at temperatures above 500°C the relative decrease in the ARM intensities is more similar to ChRM behaviour than that of IRM. Therefore, the ARM is considered to be representative of the ChRM.

interval. A small amount of goethite and possibly some super paramagnetic material is present.

4. NRM components

In almost all paleomagnetic studies of the Pliocene Trubi marls the NRM generally shows a secondary magnetization (removed at temperatures below 300–330°C) but, more importantly, two other prominent magnetization components: a low temperature (LT) component, removed between 360 and 480–510°C, and a high temperature (HT) component, removed between 480–510°C and 600°C [4,17]. The LT and HT components were obvious because in some parts of the records the two components were completely anti-parallel; in addition, these two components have also been observed in other sediments [24]. In the case of the Thvera transitions from the Trubi sediments on Sicily the LT component is less obvious (Fig. 7); however, for reasons of consistency the LT component has nevertheless been determined as the component removed in the temperature trajectory 360–480°C.

The directions of all components are determined by fitting a least squares line [25], usually through five or more demagnetization steps. The secondary magnetization has a typical present day field direction; it probably resides in MD magnetite [20]. In the Trubi marls, this secondary magnetization is removed at 200–250°C, but in the Thvera records a substantial part persists to higher temperatures resulting in LT components intermediate between the secondary and HT components (Fig. 10). During demagnetization of the LT component fluctuations in the remanence are sometimes observed, especially during a corresponding "plateau" in the decay curves (Fig. 8). Finally, the HT component, or characteristic remanent magnetization (ChRM), is removed at

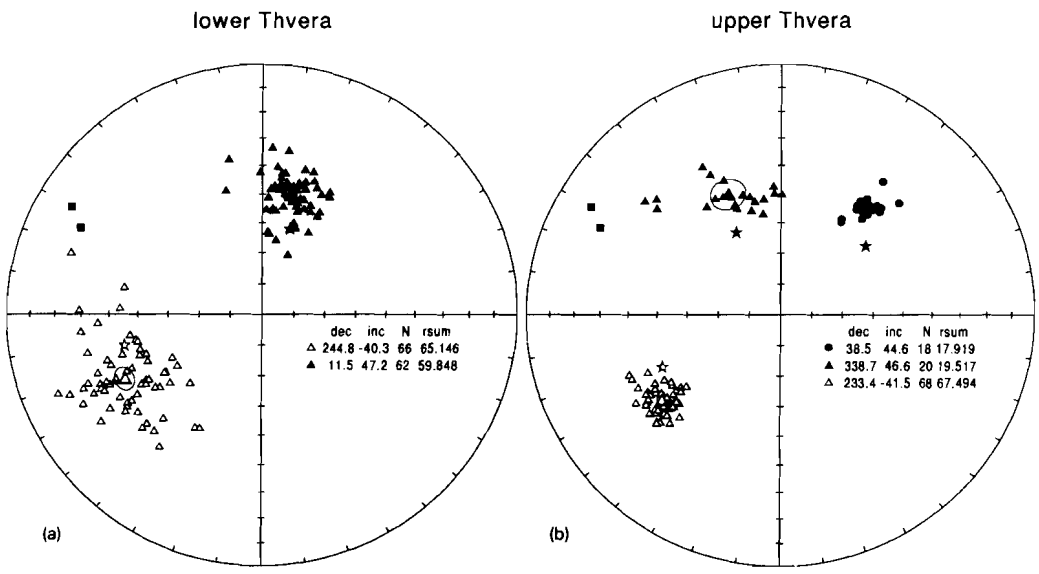


Fig. 9. Stable HT directions just before and after the transitions (triangles). Reversed directions of the lower Thvera transition were determined from the interval −15–16 cm and normal direction from 57 to 90 cm. Normal directions of the upper Thvera transition were determined from the interval 85–95 cm and reversed directions from the interval 108–125 cm. The mean directions of these stable, near-transitional directions before and after the transitions show a clear offset. Circles = calculated stable normal directions between −25 and 0 cm, preceeding two excursions in the lowermost part of the upper Thvera transition; asterisks = mean directions before tectonic correction; squares = mean directions of LT component before and after tectonic correction, averaged over trajectories where the HT component is reversed. Before tectonic correction the LT component is steeper and more westerly. The normal LT component is almost parallel to the normal HT component.

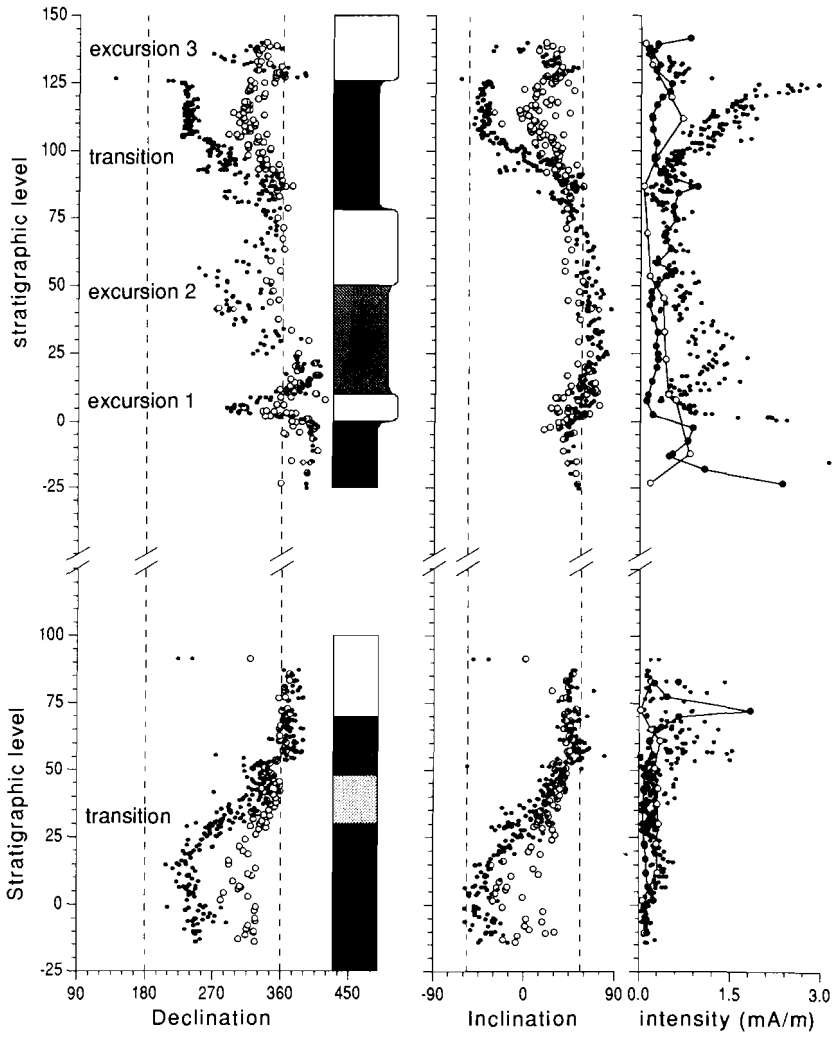


Fig. 10. Records of the declination, inclination and intensity obtained after thermal demagnetization of the upper and lower Thvera transitions. For legend of sedimentary column see Fig. 4. Circles = the LT component; dots = the HT component; dashed lines = declination and inclination (57.5°) of the geocentric axial dipole for the present latitude of the locality. The lower Thvera transition is gradual. The upper Thvera record has, besides the actual transition at level 95 cm, three intervals with major directional changes: between level 0 and 15, where the declination swings to almost 270° west, with a minor steepening of the inclination (excursion 1); between levels 25 and 90, in which the declination again swings to 270° west while the inclination steepens slowly (excursion 2) and then becomes more shallow again, back to its initial value at level −5. The declination does not return back to its initial values (35°E). Excursion 3 is at level 125, where the ChRM directions jumps from reversed to normal directions coinciding with a (sharp) lithological change, followed by a slower change back to reversed directions. Small dots = the record of ChRM intensities at 510°C in mA/m; circles = ARM(510°C); dots connected by a line = the smoothed record of the ratio ChRM/ARM(510°C), both are in arbitrary units. Data were smoothed by a 5 cm spatial window. ChRM data larger than 3 mA/m below the zero level of the upper Thvera record were not plotted in order to enlarge horizontal scale. The changes in the ratio are mainly caused by changes in the ARM(510°C).

temperatures higher than 450°C, but the most rapid decay is observed only at temperatures higher than 510°C. It is also at these highest temperatures that this component shows a (more or less) linear decrease towards the origin.

The (mean) normal and reversed directions before and after the transitions are not anti-podal but show a clear offset (Fig. 9). An overlap in the blocking temperature spectrum of the secondary component with the spectrum of the HT component is probably quite persistent up to the highest temperatures [21] and may therefore introduce an offset in HT directions: the reversed HT component is expected to have a southwest declination and the normal component a northeast declintion due to the rotation of the basin and a relatively shallow inclination caused by compaction. A secondary component will offset the reversed declination to west and decrease the inclination, conversely the normal declination will be offset to the north and the inclination increased, as is seen in Fig. 9. Therefore, the offset may be caused by the geomagnetic field but is more likely due to the secondary overprint. The offset in the normal declinations of the lowermost part of the upper Thvera transition has an average clockwise rotation of rotation of 38.5°, which is more easterly than expected for the secondary overprint. As was expected, the inclinations in the normal directions are steeper than in the reversed directions. The stable directions before and after the two transitions show negative reversal tests [26]. Uncorrected for the bedding plane, the mean normal HT directions deflect further from the north and inclinations are steeper, while the mean reversed HT directions tend to more westerly directions and inclinations also steepen (Fig. 9). The mean "reversed" LT component shows, before tectonic correction, a more westerly declination and a steeper inclination.

Scheepers and Langereis [27] also find differences between normal and reversed directions (30° and 40°, respectively) for the entire Thvera subchronozone and the overlying reversed subchronozone. They suggest that this is probably related to the higher carbonate content (and thus higher porosity) of the basal part (i.e., the Eraclea Minoa section) of the Rossello composite section [12,16]. A higher porosity may result in increased weathering of SD magnetite, which causes a secondary component that persists up to the highest temperatures [21]. The non-transitional ChRM directions show an inclination error of 7–16°(Fig. 9) and it is stronger in the reversed directions. Hence, there is probably a bias in inclination due to a secondary component. The most important cause of the inclination error, however, is most likely due to compaction and its magnitude is clearly related to the carbonate content [27], a result which is known from earlier studies [28,29].

5. The transition records

The registration of the lower Thvera transition record in the HT component (Fig. 10) shows a smooth reversal from reversed to normal directions, starting at level 15 cm. At the end of the transition between levels 50 and 55 cm, a small "acceleration" to normal declinations (including the rotation) can be seen in the lower part of the (supposedly) grey layer. The directional changes of the entire transition take place in 40 cm. This interval would represent some 8.0 kyr using a sedimentation rate of 5.0 cm/kyr.

The upper Thvera record, on the other hand, appears to be complicated and essentially four main features can be recognized (Fig. 10): two excursions in the lower part, the transition itself and one more excursion in the upper part. Excursion 1 (between 0 and 15 cm) shows a large swing of, in total, more than 100° in declination to the west, and even some 70° within 2.5 cm. This excursion is rapid and is consistently recorded in the declination. There is no corresponding swing in inclination, only a slight steepening can be seen. After the return to stable declinations (again including the 35° rotation) there is a second swing in declination: excursion 2 in the interval between 25 and 70 cm. The inclinations remain steep. The return to normal directions does not include the 35° rotation. The third feature is the actual transition. This takes place between levels 80 and 105 cm (within the grey layer), although at levels 80–85 cm some rapid changes can be seen. Using a sedimentation rate of 4.4 cm/kyr, this interval represents some 8.0 kyr. During the transition at level 93 there are some rapid changes in declination as well as inclination.

The stable, post-transitional reversed directions are followed by excursion 3. This excursion coincides with the boundary between the grey and white lithology. Moreover, the main part of the excursion is in a brown coloured part of the white layer, the part of the lithology where in the rock magnetic section the maximum in goethite was found. This makes it a priori very unlikely that excursion 3 was caused by directional changes of the geomagnetic field.

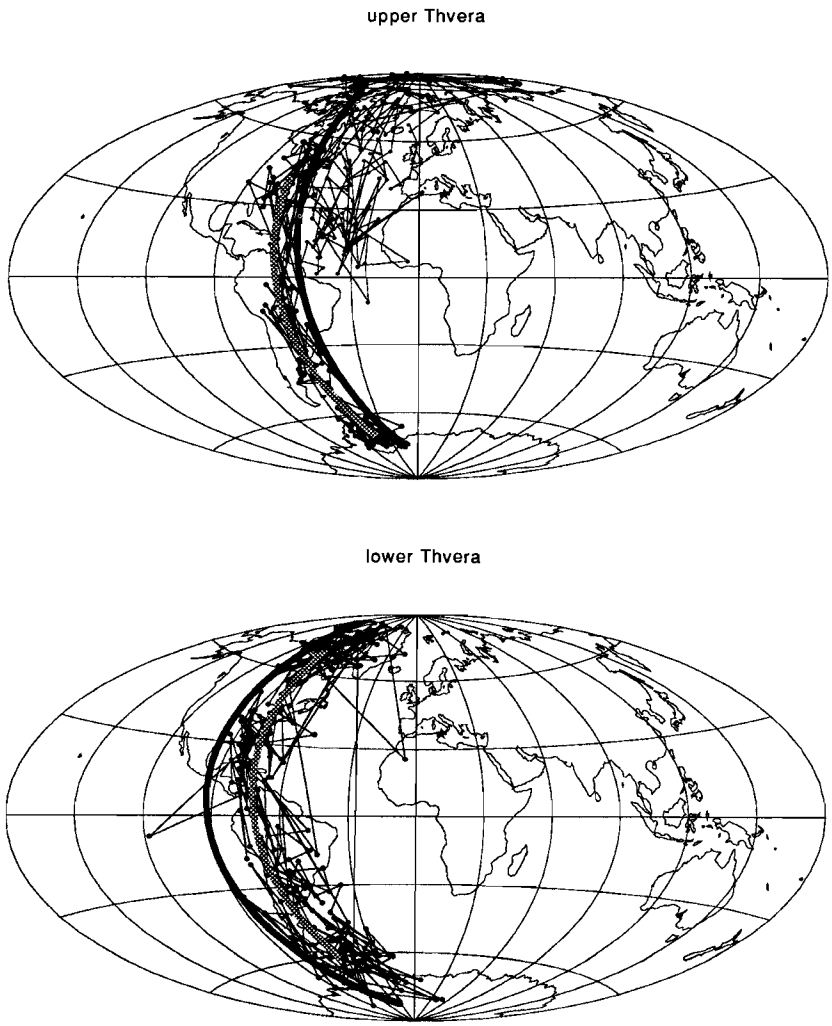


Fig. 11. Aitoff projection VGP paths of the transitions show a strong confinement to a great circle over North and South America. The excursions preceding the upper Thvera transition (smaller symbols) have their VGP's north of the equator in the Atlantic. Star = location of the site; solid line = the VGP path obtained by filtering the mean directions of under/overlying polarity zones resulting from magnetostratigraphy [12]; shaded line = filtering of near-transitional directions determined from the reversal records. The near-transitional directions are the same as in Fig. 9, except for the normal directions preceding the upper Thvera transition. There the mean direction of the ChRM's between levels 85 and 92 cm was used.

5.1 Intensity

The observed changes in magnetic mineralogy over the transition records do not allow the intensities of the ChRM (510°C) to be used as a measure of the relative paleointensity of the geomagnetic field during the transition. For a magnetite-dominated magnetic mineralogy like the Trubi sediments [21], King et al. [30] proposed a method which uses the ARM intensity as the normalizing factor for the abundance of magnetite carrying the remanence. Since the ChRM was determined at temperatures higher than 510°C, we used the ARM intensity at 510°C as a normalizing factor, although the NRM and ChRM demagnetization curves are somewhat different (Fig. 8). The number of data points of the ChRM exceeds the number of ARM points, therefore a linear interpolation was used between the subsequent ARM data points. The ratios ChRM (510°C)/ARM(510°C) were calculated by first smoothing the ChRM intensity record with a 5 cm wide rectangular moving window and normalizing it with the interpolated ARM data points.

The normalized intensity of the lower Thvera record shows a clear maximum at level 75 cm (Fig. 10). There are small local maxima at levels 2 and 57 cm, respectively. The upper Thvera record starts with very high values of 23 mA/m (not shown in Fig. 10 to enhance variations in the lower intensities), and it has a minimum of 0.15 at level 8 cm. This minimum coincides with excursion 1. The ChRM/ARM ratio increases gradually to 1 at level 85, which is at the onset of the actual transition. The intensity minimum between 100 and 115 cm is followed by a local maximum at level 125. Above level 125 there is another decrease in intensity. This decrease occurs at excursion 3, which is most likely caused by a lithological change at that level.

6. Discussion

6.1 The Sicilian lower and upper Thvera records

The average sedimentation rate during the Thvera subchronozone has been established at 4.4 (upper Thvera) to 5.0 (lower Thvera) cm/kyr [11,13,16]. The sedimentation rate during the deposition of the grey and beige marls could be somewhat higher due to increased continental run-off and a constant carbonate flux [14], or somewhat lower due to increased carbonate production in the white marls while the flux of the non-carbonate fraction is constant [31]. The difference in ages between the midpoints of two subsequent grey layers [13] is the time during which one quadruplet is deposited. By considering this deposition time, the thickness and carbonate content of each individual layer, we can compute the average sedimentation rate of each individual layer. This results (with the assumption of a constant carbonate production with a varying non-carbonate fraction, or a constant non-carbonate fraction with a varying carbonate production) in a variation in sedimentation rate from 3.9 (resp. 3.5) cm/kyr in the white (beige) layers to 5.3 (resp. 6.2) cm/kyr in the beige (white) layer and from 3.5 cm/kyr in the beige layers to 6.2 cm/kyr in the white layer. These values will not significantly alter any conclusions about the duration of the record, so sedimentation rates of 5.0 and 4.4 cm/kyr are assumed for the lower and upper Thvera records, respectively.

The VGPs were calculated after applying a 35° correction for the clockwise rotation of the location. The VGP paths of the two transitions are very strongly confined to meridians over both the Americas (Fig. 11). The VGP paths of the two excursions preceding the upper Thvera transition lie in the northern Atlantic. Tric et al. [8] showed that VGP paths of two-thirds of recently obtained transitions behave very similarly to those of the lower and upper Thvera. Laj et al. [7] 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, suggesting a causal relationship. However, Rochette [4] showed that smoothing of the non-antipodal directions before and after a transition will also result in VGP paths with a strong longitudinal confinement. By smoothing the mean stable directions before and after late Miocene and Pliocene transitions sampled in the Mediterranean, Langereis et al. [6] found synthetic VGP paths confined to the Americas which were identical to the observed VGP paths.

Mean stable directions were determined (Fig. 9) by averaging the directions over a sedimentary

interval just before or after the transitions where the changes in directions are due to noise. The stable directions of the polarity zones before, during and after the Thvera subchronozone from the magnetostratigraphic study of the Eraclea Minoa by Hilgen and Langereis [11] are shown in Table 1, together with the near transitional directions. The mean directions of the previous reversed polarity zone and the Thvera subchronozone show a marginally positive reversal test (class C) [26], while the mean directions of the subsequent polarity zone and the Thvera subchronozone, as well as the near transitional directions, have negative reversal tests. Synthetic VGP paths were calculated by smoothing the mean stable magnetostratigraphic directions (Table 1) before and after each transition, using the method of Rochette [4] (thick lines, Fig. 11). The lower Thvera synthetic VGP path is some 30° away from the observed VGP path of the lower Thvera transition. The upper Thvera synthetic VGP path has the same band of longitude as the observed upper Thvera VGP path.

Using the same procedure, the synthetic VGP paths were also calculated by smoothing the mean stable directions just before and after the transitions (Fig. 9). The coincidence of the observed VGP paths with these synthetic VGP paths is even more striking (shaded lines, Fig. 11). The smoothing may well be due to the filtering mechanism of remanence acquisition in these sediments. The coincidence of the transitional data and the smoothed non-transitional directions is also apparent in several other transitional records from the Sicilian Trubi marls [6]. This strongly indicates that the filtering mechanism of the acquisition of the sediment obscures the real geomagnetic transitional directions.

Hoffman [32] found, in transitional records from lava sequences (which are not or hardly smoothed), long-lived VGP positions clustering at spots on the globe that coincide with the preferential longitudinal bands over the Americas or its antipode found earlier [7,8]. These recurring clusters are still highly hypothetical and they may only reflect short periods of active volcanism during transitions. On the other hand, if these clusters do represent a stage of a dipolar transitional configuration, the VGP will be independent of the sampling site on the globe. Filtering of this record by a sedimentary NRM acquisition will result in a VGP path that includes the spot of these long-lived VGP positions and (if the filter width is large relative to the time of the transition) the longitudinal bands that contain these spots. Therefore, depending on the relative filter width of the sediment, some information about the transitional path may be registered by sediments and the synthetic VGP path will be confined to the Americas or its antipode.

Since the demagnetization curves of ARM and ChRM are slightly different, calculation of the relative paleointensity may not be meaningful. Indeed, the absolute maxima in the relative paleointensity records are due to high ChRM intensities as well as extremely low ARM(510°) intensities. A high geomagnetic field intensity cannot change the magnetic parameters such as the ARM(510°), so the paleointensity records must be considered unreliable. Similarly, since the demagnetization curves of J_{rs} and ChRM are different (Fig. 8), J_{rs} is also unsuitable as a normalizing factor. Another commonly used method for determining paleointensities is the ChRM/χ_0 ratio. The χ_0 records of both transitions (Fig. 4) show the same tendency as the ARM(510°) records,

TABLE 1

Mean declinations (dec) and inclinations (inc) of subchronozones before, during and after the Thvera suchcronozone

Subchrononzone	dec	inc	N	$r$sum	Length (cm)	Angle	Critical angle	Results
4.47–4.57	187.1	−41.4	7	6.93	500			
						10.3	8.0	neg
Thvera	−4.6	47.4	11	10.86	1200			
						7.8	10.7	C
4.77–4.86	186.5	−50.3	4	3.96	500			

Figures in first column are the ages of the boundaries of the subchronozones. N = number of samples; length = length of subchronozone; $r$sum, angle, critical angle and results (negative or class C) are parameters from the reversal test [26].

and the ChRM/χ_0 ratio will, therefore, result in the same maxima as those shown in Fig. 10.

6.2 Comparison of the upper Thvera records from Sicily and Calabria

We have compared the Sicilian and Calabrian records from the same upper Thvera transition in the Mediterranean region, because both records should be identical up to very high order coefficients. The upper Thvera transition record was earlier reported as one of five successive transitions [2]. These transitions were sampled at the Singa section, some 250 km from Eraclea Minoa (Fig. 1) in the Calabrian Trubi sediments. The grey–white–beige–white sequence in the Sicilian Trubi is equivalent to a grey–white sequence in the Calabrian Trubi [15]. Therefore, the vertical scales have been stretched linearly so that the boundaries from grey to white at levels 0 and 125 cm in the Sicilian record are almost coincident with the same boundaries in the Calabrian record (Fig. 12). Although the stratigraphic resolution of the Calabrian upper Thvera record is in the order of a few centimetres (a much lower resolution than obtained in Sicily) the excursions and transition should be identical. It appears, however, that the first two excursions from the Sicilian record are either different (excursion 1) or absent (excursion 2) in the Calabrian record. Steep inclinations after excursion 1 occur in both records. The excursions cannot be caused by recent overprints because a recent overprint would have normal directions, whereas excursion 1 in the Calabrian record is even fully reversed; also showing, in addition to the Sicilian record, negative inclination values.

On the other hand, one could argue that the (tendency to) normal directions between excursion 1 and the transition are caused by a recent overprint. This seems rather unlikely because the rock magnetic parameters do not indicate secondary minerals in this part of the record. Recently, Van Hoof and Langereis [5] showed that the HT (and LT) component can acquire their remanence at a considerable depth below the sediment–water interface and that this depth lag was not constant throughout the lithology. It was suggested that the magnetic minerals carrying the components were authigenically formed under

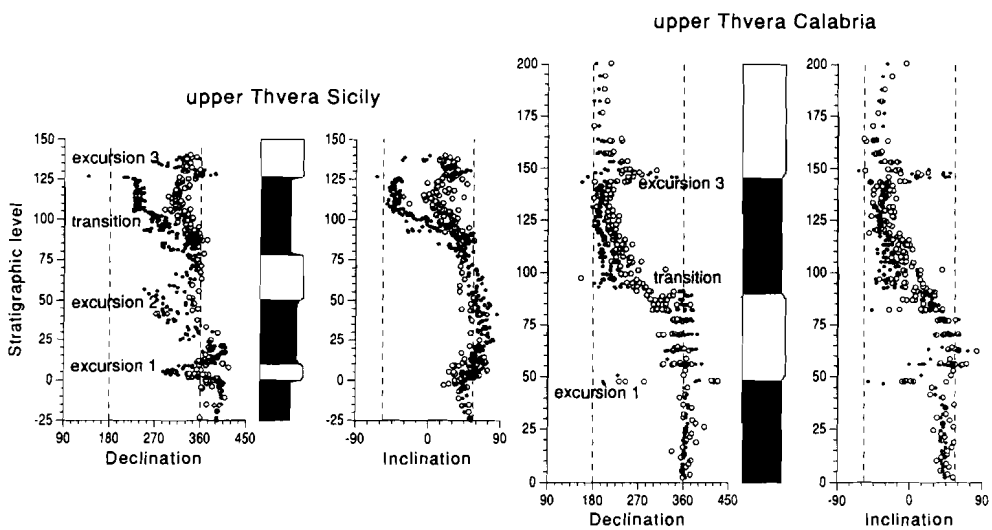


Fig. 12. Comparison of the upper Thvera record from the Sicilian Trubi Formation and the one from the Calabrian Trubi Formation [2]. The vertical scale is changed calibrate the same lithological boundaries from grey to white. In the Calabrian sampling excursion 1 is present as full reversed directions, excursion 2 is absent, the transition takes place lower and faster in the sediment and transition 3 is similar to that in the Sicilian record.

different and cyclically fluctuating paleoredox conditions. This mechanism does explain excursion 1 if it is assumed that, in the Calabrian record, the white layer has totally acquired the post-transitional, reversed direction, while the Sicilian record has only partially acquired the post-transitional directions in excursions 1 and 2.

Van Hoof et al. [1] have suggested that, due to the changing paleoredox conditions, a migration of Fe^{2+} and Mn into mainly the lower white (i.e., on top of grey and below beige) layers will take place, leading to the formation of secondary magnetite. Therefore, a maximum delay in the lower white layer is most likely. The transition itself in the Calabrian record is somewhat lower in the sediment and somewhat "faster". In spite of some small differences in the transitions, the VGP paths of both upper Thvera records are confined within the very same great circle over North and South America. Langereis et al. [6] indicate that the VGP path of the Calabrian record may also be attributed to a smoothing process of the sediment, similar to the VGP paths of the transitions in this paper.

Excursion 3 is also recorded in the Calabrian Trubi as fully normal directions; it is more or less identical in both records but, in both cases, we attribute this feature to a very recent overprint caused by weathering. Contrary to the Sicilian record, the LT component in the Calabrian sediment has clearly recorded the transition. This indicates that this component in the Sicilian record (if present at all), has an different origin from than in the Calabrian record.

7. Conclusion

We have identified magnetites as the most important carrier of the HT and LT components in records of the lower and upper Thvera reversal boundaries recorded in the Trubi marls of Sicily. The lower Thvera transition is recorded as a smooth change from reversed to normal directions. The actual upper Thvera transition is preceded by two "excursions" and followed by another. This last "excursion" is most probably caused by a recent overprint due to weathering. The two excursions preceding the transitions are considered to be sedimentary artefacts due to early diagenetic processes. The directions of the excursions are inferred to be caused by the post-transitional reversed geomagnetic field. The observed HT (and LT) components in this sediment interval most probably acquired post-transitional directions due to the formation of secondary magnetite, while outside the excursion the sediment had already acquired the pre-transitional directions. The upper Thvera transition, itself, is a smooth change from normal to reversed directions.

The relative paleointensity record was obtained by normalizing the ChRM(510°C) with ARM(510°C). The minima and maxima are not only due to changes in ChRM(510°C) but also to the (lithology dependent) ARM(510°C), indicating that the normalizing procedure (with ARM(510°C)) is probably not suitable for determining paleomagnetic field changes in intensity. In addition, if the ChRM is the vector sum of normal and reversed directions caused by smoothing, then the intensity is low while the geomagnetic field intensity during the normal and reversed directions may be strong. The VGP paths of the upper and lower Thvera are most probably the result of smoothing of the stable directions before and after the transition [3,5]. In addition, it is unlikely that the complex behaviour of the upper Thvera record is a registration of the transitional geomagnetic field because the directional changes do not completely match directional changes of the Calabrian upper Thvera record only about 250 km away.

For a study of the long-term geomagnetic behaviour (magnetostratigraphy), these sediments are very suitable, since they match the geomagnetic polarity time scale very well [cf., 11]. Any study of polarity transitions however, necessitates an extremely good understanding of remanence acquisition in sediments and requires ample rock magnetic and geochemical studies.

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Chapter 5

A comparison of the lower and upper Sidufjall geomagnetic transition records from Southern Sicily with the records from Calabria (Italy)

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Abstract. The paleomagnetic records of the upper and lower Sidufjall polarity transitions from marine marls on southern Sicily are compared with the same transitions recorded in similar marls from Calabria (*Linssen*, 1988; 1991), some 250 km east from the location on Sicily. The morphology of the same transitions should be identical up to the highest degree, since on global scale these two sites are at the same location. The morphology of the transition records, presented as declination-inclination-intensity plots, VGP paths and in Zijderveld diagrams are completely different. In addition, comparison of the mean directions in the subchronozones result in significant variations. For magnetostratigraphic studies the sediment is very suitable, for detailed studies of the geomagnetic field however, there are still too much unknown local sedimentary and geochemical factors that influence the directions of the remanence. The observations cannot be explained by the diagenetic magnetite model proposed by *van Hoof et al.* (1993).

1. Introduction

In studies to the behaviour of the geodynamo during polarity transitions, sediments are commonly used as a recorder of the geomagnetic field. Although its significancy is discussed (c.f. *Valet et al.*, 1992; *Laj et al.*, 1992), the observation that the Virtual Geomagnetic Pole (VGP) paths of many sedimentary transition records are confined to a great circle over North and south America or its antipode (*Laj et al.*, 1991) excited the world of the geomagnetists. *Constable* (1992) showed that the VGPs of non-transitional data from lavas have a non-zonal bias according to the path over the Americas or its antipode due to a equatorial dipole. A reversal, starting and ending with stable VGPs in these preferential longitudinal bands of the VGP paths, will follow this band, whether the transition is completely the result of smoothing of the geomagnetic signal by the sediment as suggested by *Rochette* (1990), or the result of an axial dipole that decays and recovers in opposite direction while the equatorial dipole remains unchanged. *Langereis et al.* (1992) simulated the smoothing effect by filtering the average stable directions before and after the transitional records in Mediterranean sediments with a moving window and they found a good correlation with the observations. *Weeks et al.*, (1992) argue that filtering of the geomagnetic signal during the acquisition of the remanence in sediments indeed may take place but that it is unlikely that the transitional geomagnetic field is completely obscured by smoothing. They feel therefore that in the remanence of the sediments not only the non-zonal bias of the stable VGPs is observed but also intermediate geomagnetic directions and the long-lived clusters of intermediate directions during transitions as suggested by *Hoffman* (1991).

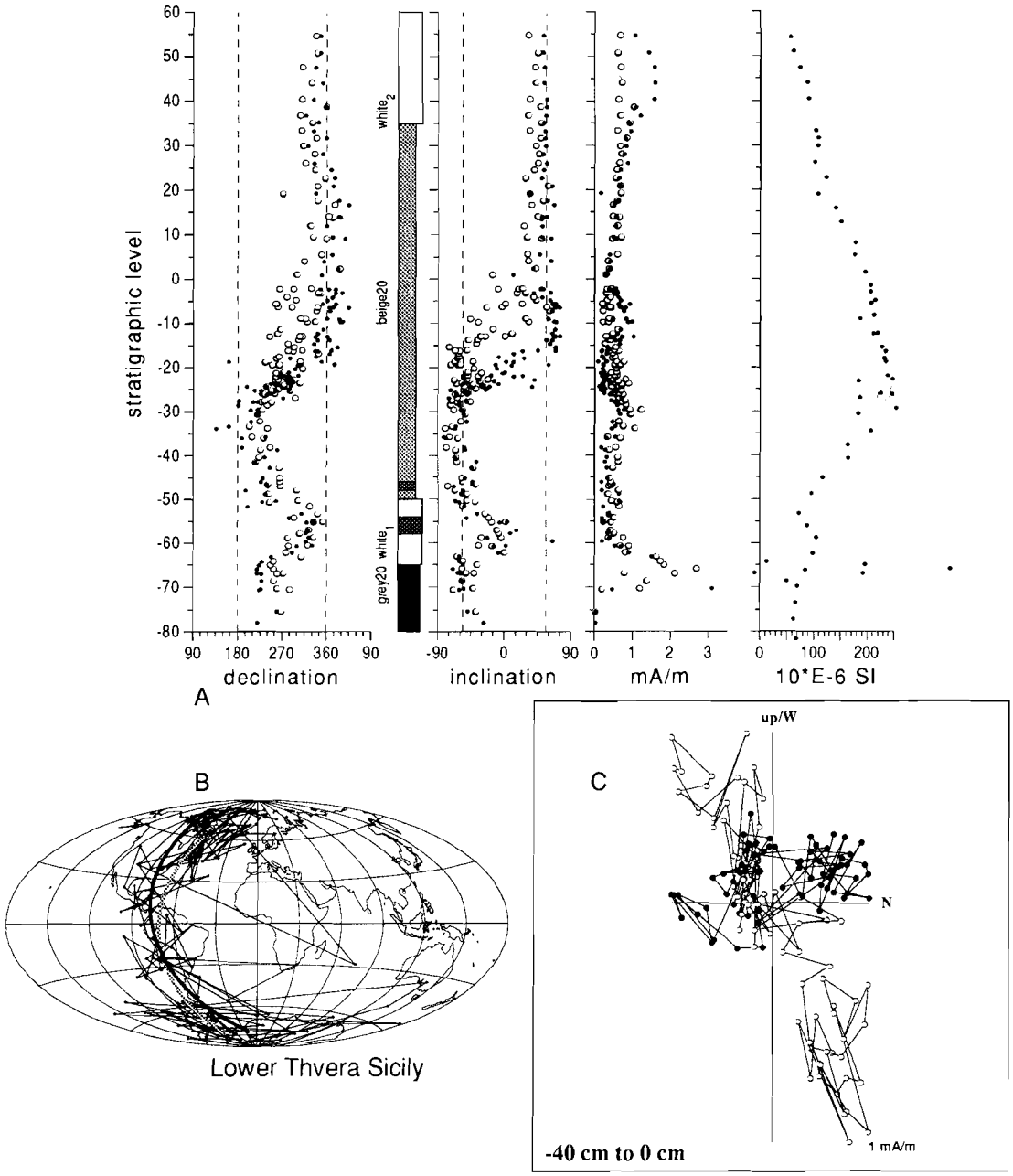

figure 1. A). Record of the declination, inclination, NRM intensity obtained after thermal demagnetization at 510° C, as well as the initial susceptibility of the lower Sidufjall record from Sicily. The lithological repetition of grey/white/beige/white is presented in the stratigraphic column as black/white/shaded/white. The dark grey levels in the white$_{(1)}$ and beige 20 are brown coloured levels, due to oxidation spots. These oxidation spots are presumably ironhydroxides, formed shortly after deposition under (sub)oxic circumstances. Circles denote the LT component, solid dots the HT component. Dashed lines indicate declination and inclination (57.5°) of the geocentric axial dipole for the present latitude of the locality. B). Virtual Geomagnetic Pole (VGP) paths of the data. The smoothed mean non-transitional (near-transitional) directions are denoted in black (grey) lines. C). Zijderveld diagram presentation of the transitional data after Mary and Courtillot (1992). Solid circles are projections on the horizontal plane; open circles are projections on the vertical north/south plane.

The way in which the sediment acquires its remanence is largely unknown. Not only a certain degree of smoothing might occur (*Hoffman and Slade*, 1986; *Rochette*, 1990) as shown above, but also changes in timing of the formation of the remanence bearing minerals, changes in bioturbation, small changes in mineralogy during diagenesis (c.f. *Karlin*, 1990; *van Hoof and Langereis*, 1991; *van Hoof et al*, 1993). This may result in variation of the remanence lock-in depth or a variation in degree to which the recording of the geomagnetic field is smoothed. These changes may take place down to one metre or more below the sediment water interface (*van Hoof and Langereis*, 1991). On a scale, as used in magnetostratigraphy, one metre is usually less than the spatial resolution, but in the study of high-frequency changes of the geomagnetic field these modifications are fatal. Therefore, it is crucial to investigate whether variations in directions and intensities of the remanence are due to the paleomagnetic field or to other causes. One way of doing this is comparing the same polarity transition that has been recorded on - in global sense - the same location. These records should be identical and any difference in direction and intensity must be caused by changes other than those of the geomagnetic field.

In this context, the data of the upper and lower Sidufjall transitions, recorded in marine marls of the Trubi sediments on southern Sicily are compared with the same transitions recorded in quite similar sediments from Calabria, which were presented earlier by *Linssen* (1991).

Geology

The Trubi marls from Sicily and Calabria, in which the geomagnetic transitions were recorded, have been described and correlated in lithostratigraphic, magnetostratigraphic and cyclostratigraphic studies (cf. *Hilgen*, 1991; and references therein). The sediments show a layering with an on average 21-22 kyr cyclicity, correlated with orbital forcing due to the precession of the Earth's rotational axis. The cycles range from number 1 at the Mio-Pliocene boundary up to number 119 at the Gauss/Matuyama boundary (*Hilgen*, 1991). On Sicily, a single cycle is expressed in a colour layering with a grey/white$_{(1)}$/beige/white$_{(2)}$, whereas in Calabria the white$_{(1)}$/beige/white$_{(2)}$ layering is represented by one single white layer. In the Trubi sediments on Sicily the initial susceptibility (χ_0) is a good indicator for the cyclicity in these sediments, since this parameter shows a clear maxima in the beige layers (figs. 1,3; van Hoof et al., 1992), while there is no indication for this relation in the Calabrian Trubi (figs 2,4). The stratigraphical distance between the centres of grey 20 and grey 23 is some 4.7 metres in the Sicilian section. The time difference between the two levels is 105 kyr and the sedimentation rate is therefore 4.5 cm/kyr (Hilgen, 1991). In the Calabrian section the stratigraphical distance between the two layers is some 5.1 m and the sedimentation rate is 4.9 cm/kyr. The lower (upper) Sidufjall polarity transition was determined in small scale cycle 20 (22). Due to a relatively high carbonate content in cycles 22 and 23, the distinction between the individual colour layers is difficult to make. The white$_{(1)}$/beige part of cycle 22 is exposed as one beige-like layer, the white$_{(1)}$/beige part of cycle 23 as a white layer. The magnetomineralogy of the Sicilian sediment and the Calabrian sediment is similar; the remanence of the two sections is carried by magnetites (van Velzen and Zijderveld, 1990; 1992) and the transitions are recorded by two components, a low temperature (LT) component, demagnetized at temperatures of 480 - 510°C, and a high temperature (HT) component demagnetized at 580° - 600° C (van Hoof and Langereis, 1991). These components have earlier been described in detail (van Hoof et al., 1993)

Lower Sidufjall

Lower Sidufjall record from Sicily (LSS)

Up to level -65 the directions of the LSS record are clearly reversed (fig. 1). The lowermost data have very low intensities, immediately followed by extremely high intensities of 7 mA/m between levels -70 and -65 cm. During the decrease in intensities just above level -65 cm - in the part of the white layer with brown spots (level -62 cm) - an excursion from the initial reversed directions to Northwest/shallow directions is observed. In beige, at level -45 cm the directions are again reversed and between levels -30 and -15 cm (-25 and 5 cm) the HT (LT) component changes from reversed (R) to

normal (N) via west/shallow directions. During the transitions of both components the intensities show minima. The LT component reverses polarity higher in the sediment column than the HT component, while the excursion in the white$_{(1)}$ 20 layer is recorded at the same level for both components. This implies that the lock-in depth of the HT component was equal to the lock-in depth of the LT component at the level of the excursion but it was larger at the level of the transition (see also van Hoof and Langereis, 1991).

The virtual geomagnetic pole (VGP) path, calculated after a 35° correction for the clockwise rotation of the sedimentary basin is strongly confined to the east coast of North America and to South America. For the smoothing procedure, the near-transitional directions before (after) the LSS record were determined by averaging the data below level -63 cm (above level 2.3 cm). The non-transitional data are calculated by averaging data from the magnetostratigraphic sub-zones in the near-by Eraclea Minoa section (Langereis and Hilgen, 1991). The synthetic VGP paths that result from smoothing the stable directions before and after the LSS transition show a striking coincidence with the observed VGP path (fig. 1B).

In a Zijderveld demagnetization diagram, (see Mary and Courtillot, 1992), the confinement of the VGP path over the Americas is corresponds to westerly declinations and shallow inclinations. Only the directions between levels -40 and 0 cm have been plotted to enhance the low intensities, the west/shallow directions are observed, although the signal is rather noisy (fig 1C).

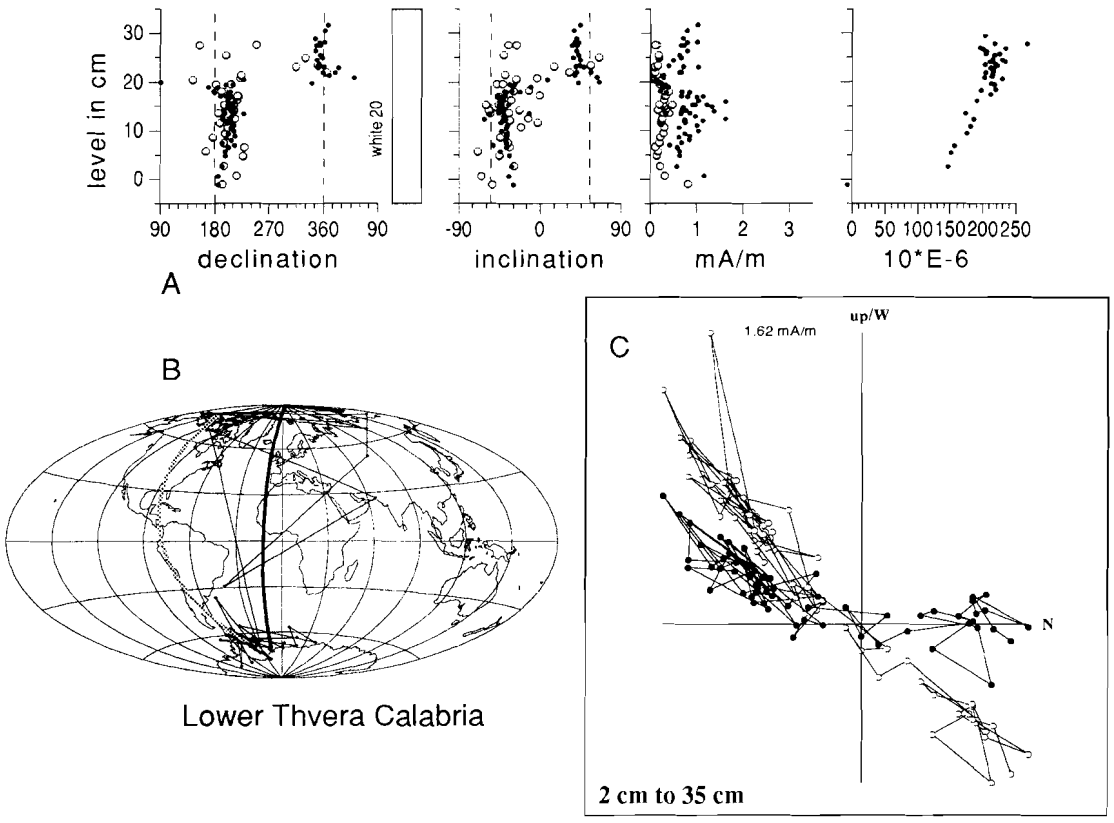

figure 2. Lower Sidufjall record from Calabria (Linssen, 1988; 1991). In the lithological column, a white layer is equivalent to the white/beige/white alternation of the Sicilian sediment. See also caption to figure 1.

Lower Sidufjall record from Calabria (LSC) (Linssen, 1988; 1991)

The LSC transition was sampled over an interval of approximately 40 cm, completely within a white layer. This interval may correlate with any interval from the white$_{(1)}$-beige-white$_{(2)}$ sequence from Sicily (fig. 2). After the maximum of 4 mA/m at level 0, the HT component has in general intensities of approximately 1 mA/m with a local minimum at level 10 cm and a pronounced minimum at level 20 where the transition is recorded. The intensities of the LT component are considerable lower than the HT component, with the exception of level 20 cm where they are equally low. Up to this level, the pattern of changes in intensity is similar to that of the HT component. Above level 20 cm, the LT intensities do not increase, contrary to the HT intensities. The transition recorded by the HT component takes place between levels 18 and 23 cm with very few intermediate points, of which one declination is clearly east. The transition is not clearly recorded by the LT component, since above level 20 cm the directions are scattered.

A rotational correction of the sedimentary basin has not been applied to the directions of the records from Calabria. The stable non-transitional directions were determined from data of the subchronozones in Zijderveld et al. (1986). Smoothing of these directions show a strongly near-sided, zonal VGP path. The smoothed near-transitional directions, resulting from averaging data below level 17 cm, respectively above level 23 cm pass North and South America. The synthetic VGP paths do not fit the very few observed intermediate VGP data points although path of the smoothed non-transitional directions show a hardly significant tendency to the points.

All datapoints of the LSC record, with the exception of the lowermost two have been used in the Zijderveld diagram (fig 2). The offset of the declination of the reversed directions is not observed in the normal directions and the intermediate directions appear to be nothing more than some scatter around the origin of the projection. The transition is recorded as a reversed direction with an offset, that decays through the origin and recovers in a normal direction without an offset.

Comparison of the lower Sidufjall records

The LSS transition has been sampled over a much longer trajectory than the LSC. It is therefore likely that the directional changes in the white layer at level -55 cm and the intensity maximum at level -65 in the LSS record was not sampled in the LSC transition. The transition recorded by the HT component takes place over some 15 cm in the LSS record, and over some 5 cm in the LSC record. In the LSS record 15 cm represents 3 kyr, in the LSC record 5 cm represents less than 1 kyr, a difference that is difficult to be explained by the geomagnetic field. In addition, the maximum in intensity in the LSC record at level 0, 20 cm below the level with intermediate directions is probably not due to the Earth's magnetic field since this maximum is not observed in the LSS record.

Although the LSC record has less intermediate data than the LSS record it is clear that the VGP paths are completely different. The synthetic VGP paths from smoothing the mean near-transitional directions of the LSC and LSS records coincide, but the synthetic VGP path from smoothing the mean non-transitional directions do not.

In Zijderveld projection, the LSS record has west shallow intermediate directions and the intermediate directions of the LSC record seem to be nothing more than noise. The equatorial dipole that would have caused the west/shallow directions in the LSS record is completely absent in the LSC record. This is an indication that the morphology of the transition records is not caused by the transitional geomagnetic field.

Upper Sidufjall

Upper Sidufjall record from Sicily (USS)

The USS record (fig. 3) starts with extremely low intensities and as a consequence the directions are scattered, although they generally have positive inclinations. The scatter in the HT component is stronger than in the LT component, which may indicate that the present day field has a stronger influence on the LT component than on the HT component. Between -60 and -40 cm there is a change to more stable reversed directions via west/shallow directions, but where the intensities are very low (from levels 10 to 40 cm) the scatter in directions is increased.

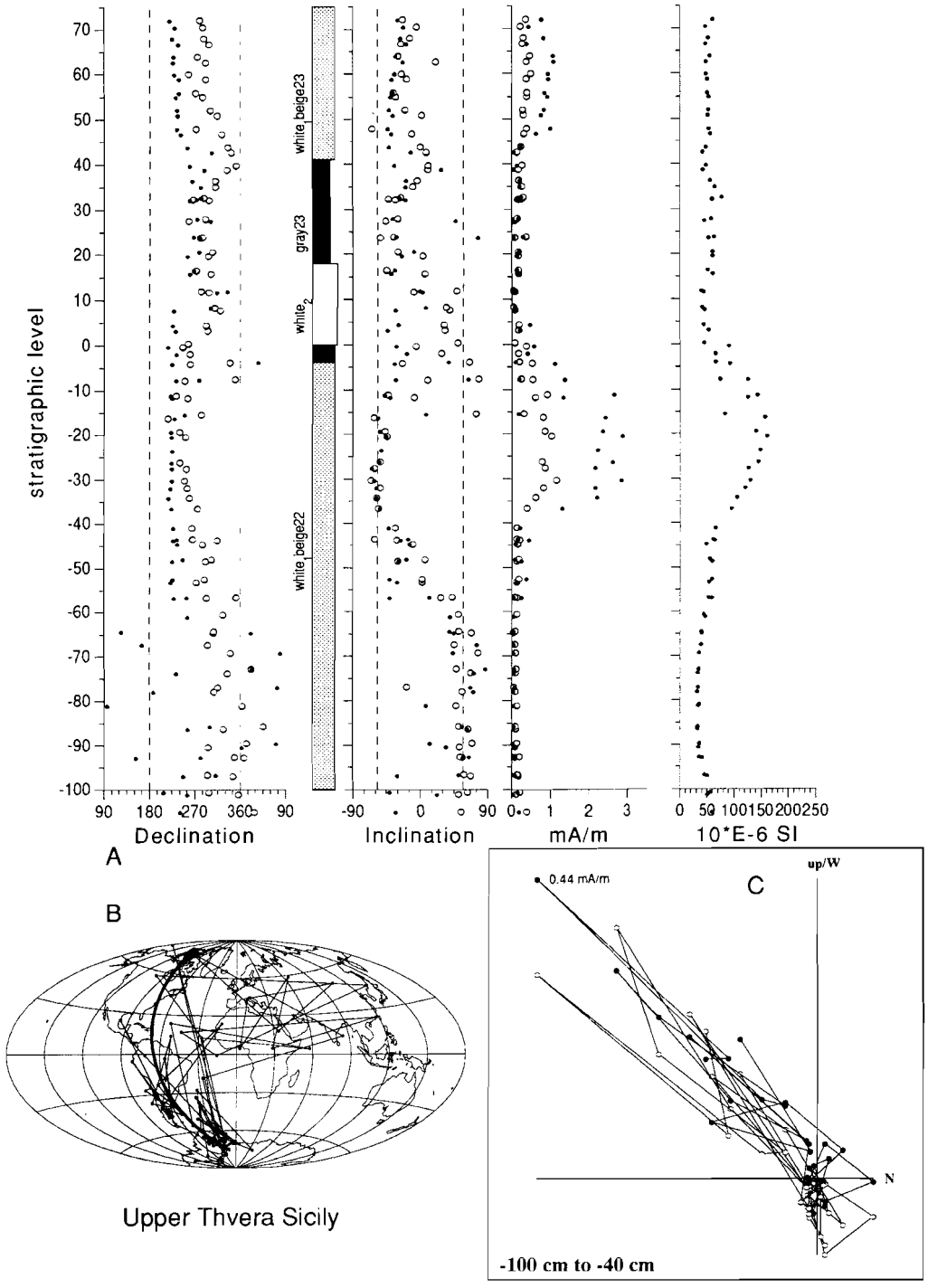

figure 3. Upper Sidufjall record from Sicily. In the field both white$_{(1)}$/beige layers are expressed as white layers, therefore no distinction between white$_{(1)}$ and beige has been made in the stratigraphical column. See also caption to figure 1.

The VGP path (fig 3b) starts with low-latitude near-sided locations and a cluster over the South-pole, followed by a cluster in South America. This second cluster is a representation of the low intensity directions in figure 3a between levels 0 and 40 cm. No near-transitional synthetic VGP path has been calculated since the USS record has no clearly normal directions before the transition. The synthetic VGP path based on smoothing of the mean non-transitional directions is confined to North and South America and passes the clusters on the South-Pole and South America.

The VGPs from the data below level -60 cm in figure 3 are located in North Africa, Asia and Europe. From Zijderveld projections - where only the data up to level -40 cm have been plotted to enhance the low intensities - it is clear that these VGPs represent mainly noise and that the reversed directions show high frequency variations, (i.e. variations between the subsequent data) mainly a resulting from variations in intensity.

Upper Sidufjall record from Calabria (USC) (Linssen, 1988; 1991)

In the USC record (*Linssen*, 1991; fig 4) the directions are north/shallow below level 40 cm and after a decrease in intensity at level 40 declinations go to south. Above level 55 the directions are clearly normal, while the intensities remain low. Between levels 80 and 115 the transition is recorded with swings from normal to reversed directions between levels 80 and 85 cm, reversed to normal at the white/grey lithology change and finally a normal to reversed directional change at level 115 cm. From this level the intensities increase to a maximum of 5 mA/m at level 160.

The intermediate VGPs of the USC record are scattered all over the globe. The smoothed non-transitional VGP is, like in the LSC record, almost zonal, whereas the near-transitional path passes over India.

In the Zijderveld projections the data between levels 40 and 115 cm have been used. Apart from the noise, it appears that the directions are mainly north/shallow and south/shallow up. High frequency variations are seen both in directions and intensity, suggesting a very noisy signal.

Comparison of the upper Sidufjall records

The USS record is sampled from the bottom of the white$_{(1)}$/beige layer of cycle 22, over white$_{(2)}$22, grey23 and ends in white$_{(1)}$/beige of cycle 23. In terms of small scale cycles the stratigraphical trajectory of the USC record is less than the USS record. It starts in white22 and its uppermost part is in grey23. The intensity maximum observed in grey23 of the USC record is not observed in the grey23 layer of the USS record. The intensity maximum in the USS record is observed in beige22, and it coincides with the maximum in χ_0, but both maxima are not observed in the white layer of the USC record.

The directional changes in the USS record take place below the maximum in susceptibility, i.e. in the lower part of the white$_{(1)}$/beige-/white$_{(2)}$ of cycle 22. The main directional changes in the USC record take place in the middle and upper part of the white 22 layer and the lowermost part of the grey 23 layer. The equivalent of the white$_{(1)}$/beige/white$_{(2)}$ cycle in the Sicilian Trubi is the white layer in the Calabrian Trubi, so the directional changes do not coincide stratigraphically. Moreover, the character of the changes are completely different. The USC record is completed after a short directional excursion to reversed directions, while in the USS record this excursion is not clearly observed.

Although the VGPs of both records are strongly scattered, the characteristics are quite different. The intermediate VGPs of the USS records are near-sided on the northern hemisphere, in the USC record they are scattered over the globe. Like in the both lower Sidufjall records the smoothed non-transitional VGP paths from the upper Sidufjall records are clearly different, in addition the smoothed near-transitional directions have completely different paths.

The high frequency variations in the USS, as shown in the Zijderveld projections are strongly related to intensity variations, whereas the variations in the USC are due to both intensity and directional changes.

Like in the two lower Sidufjall records it is obvious that in both upper Sidufjall records the transitional geomagnetic field plays a minor role in the directions of the records of the transition.

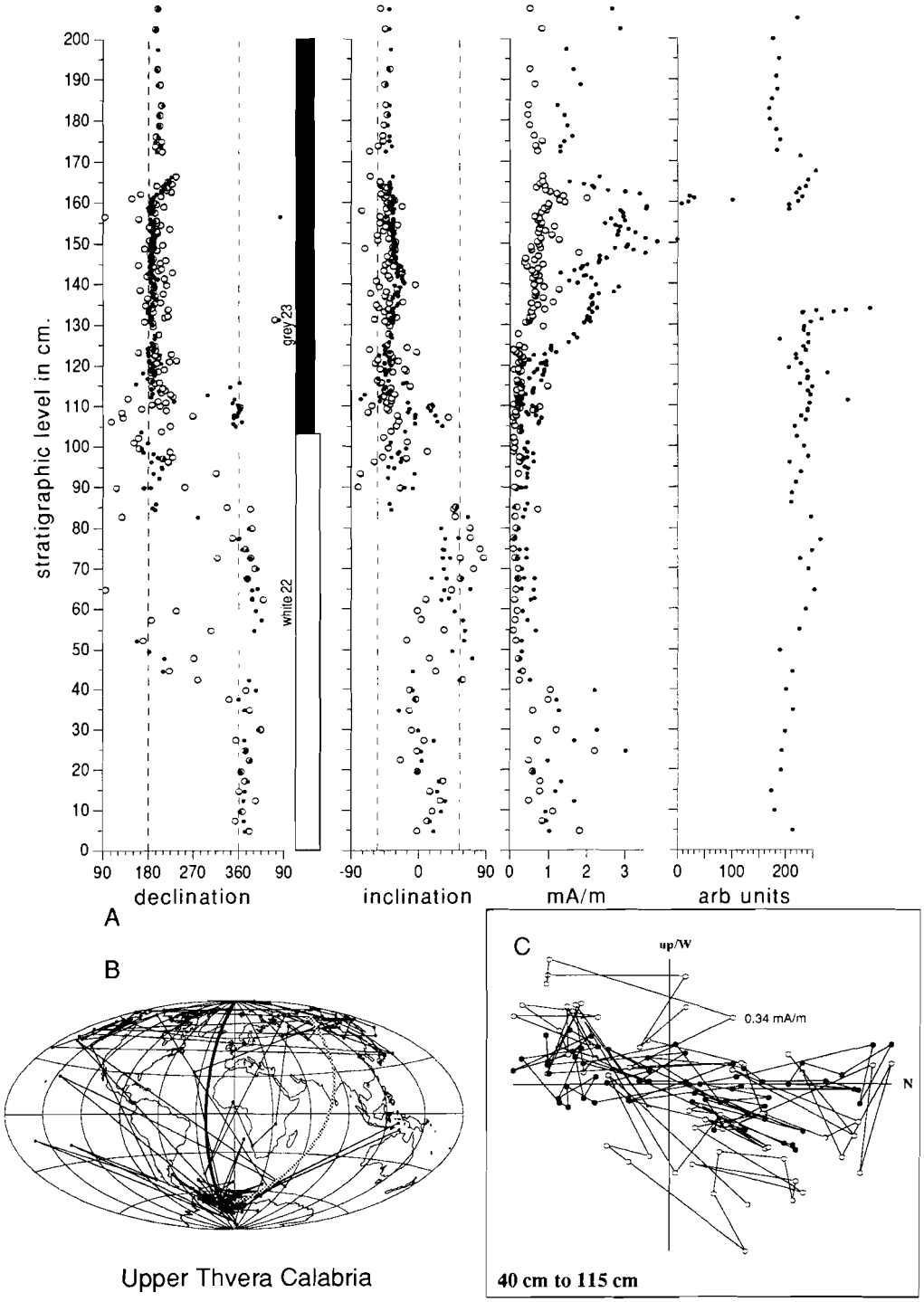

figure 4. Upper Sidufjall from Calabria (Linssen, 1988; 1991). See also caption to figure 1.

Discussion

The two polarity transitions, sampled at two sites separated by a distance of some 250 km, were sampled over different intervals and slightly different lithologies. In spite of these differences, the directional behaviour of the parallel sections should be identical if the remanence exactly records the geomagnetic field. However, they are totally different and the transition of the upper Sidufjall from Calabria is recorded in a different sedimentary cycle. It is therefore obvious that the directional changes recorded in the sediments are not true registrations of the transitional geomagnetic field but sedimentary artefacts. In addition, the cluster of intermediate VGP positions at South America observed in the USS record (fig 4) is not a cluster of long-lived intermediate directions as suggested by Hoffman (1992) but its is more likely that they are due to a decrease in the NRM intensity and as a consequence more influenced by a normal overprint

magnetites is superposed over the remanence of the primary magnetites. The superposition of the secondary magnetites may introduce intermediate directions or even complete polarity changes in the sediment. In the grey layers the environment is always anoxic due to a high organic carbon content. This prevents the formation of secondary magnetites in these layers and therefore these layers the remanence will show the true geomagnetic field at the time of formation of the primary magnetites. If the remanences of two subsequent grey layers have the same post transitional directions then the secondary magnetites are also post transitional and the layers inbetween these two layers should therefore have also a post transitional direction. If the remanences of two subsequent grey layers show different polarities of the same pre-transitional directions, then the primary and secondary magnetites in the lithology in-between may have respectively pre-transtional and post-transitional directions. The remanence of these layers may therefore show reversed or normal polarities as well as intermediate directions, excursions and transitions, depending on

Non transitional directions								Near transitional directions							
Age (Ma)	N	D(tc)	I(tc)	rsum	α95	χ	χ*		N	D(tc)	I(tc)	rsum	α95	χ	χ*
Eraclea Minoa															
4.24-4.40	13	223.2	-44.9	12.79	5.53	12.3	8.1	LSS R	11	229.5	-56.9	10.700	7.3		
Sidufjall	4	25.8	45.3	3.980	7.530			LSS N	27	358.0	56.4	26.185	5	27.6	8.6
4.47-4.57	7	222.1	-41.4	6.93	6.65	12.5	8.6	USS N		no data					
								USS R	33	228.3	-40.5	32.088	4.28		
Singa															
4.24-4.40	3	190.4	-40.1	2.991	8.290	10.2	8.3	LSC R	39	203.0	-41.1	38.368	2.98		
Sidufjall	14	7.4	50.1	13.782	5.190			LSC N	13	354.8	40.7	12.935	3.05	21.1	4.2
4.47-4.57	20	191.7	-43.1	19.888	2.510	7.6	5.8	USC N	17	22.0	43.0	16.346	7.36		
								USC R	64	187.6	-34.2	63.363	1.79	14.2	7.6

table 1: Mean non- and near-transitional directions. In the first column are ages of the subchronozones,
N: sumber of samples, D(tc): mean declination after tectonic correction, I(tc): mean inclination after tectonic correction
χ,χ: angle, respectively critical angle between pro and post transitional directions*

Van Hoof et al., (1992) proposed a geochemically constrained model that could account for the observed morphology of a transition record also from the Trubi sediments on Sicily. In this diagenetic magnetite model, immediately after the deposition of the sediment, primary magnetites have recorded the geomagnetic field. In oxic to sub-oxic environments, secondary magnetites may be formed during several thousands of years after deposition. The remanence of the secondary

the relative intensities and directions of the remanence of the primary and secondary magnetites. In addition, differences between the observations in the records at the two sites may be explained by the dependency on the local redox circumstances. Summarizing, the diagenetic magnetite model predicts that close to a transition the lithology between two subsequent grey layers may show all kinds of directional changes as long as the layers do not have the same post transitional direction. The grey layers

before and after the lower Sidufjall transition have pre, respectively post transitional directions (fig 1; Zijderveld et al, 1986; Langereis and Hilgen, 1991). Therefore the differences between the LSS and LSC records may be caused by differences in local redox circumstances and they do therefore not contradict the model. On the other hand, the restriction of the model that the grey layers are reliable recorders of the geomagnetic field needs some adjustments because the directional swing, observed in the lowermost part of grey 23 in the USC record, is not present in grey 23 of the USS record.

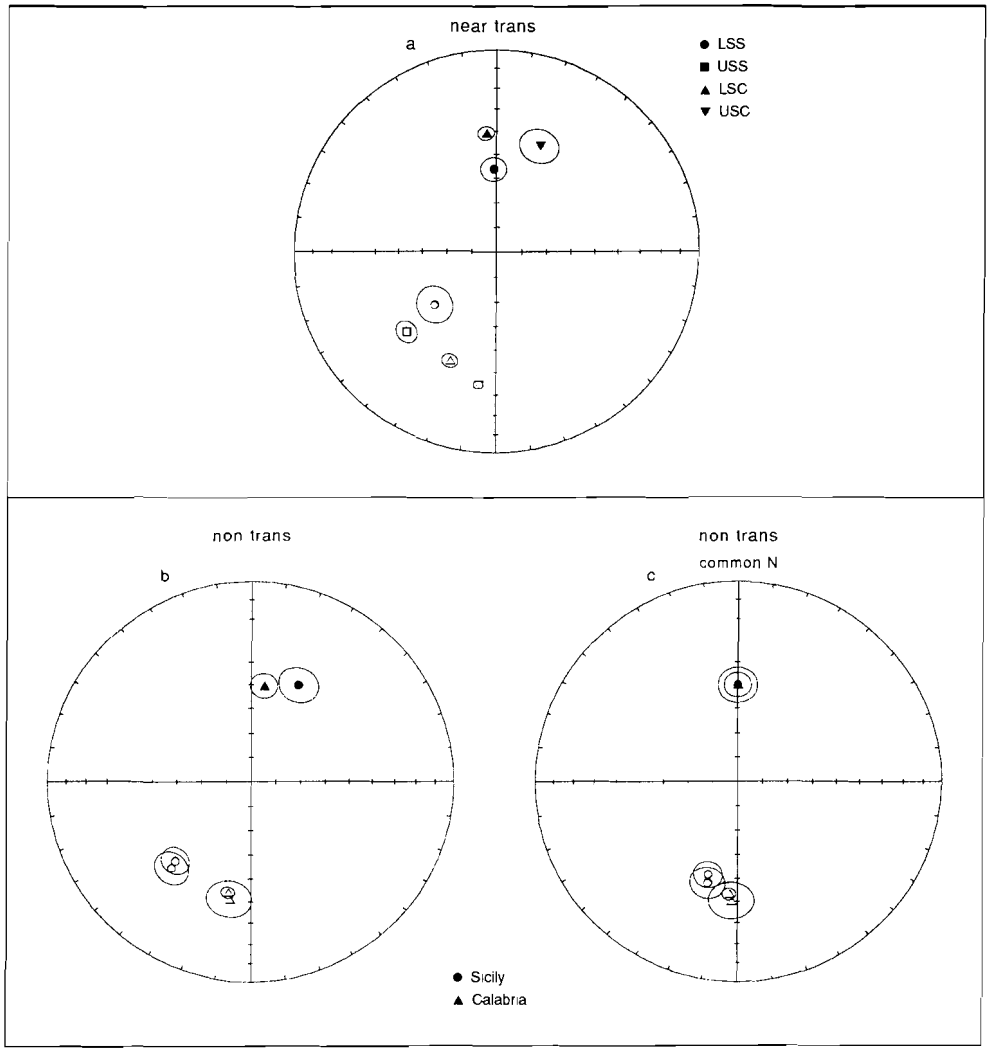

figure 5. Equal area projection of mean non- and near-transitional stable directions of the transition records with α_{95} circles. Closed (open) symbols denote downward (upward) projection. A) A normal near-transitional mean direction was not determined in the USS record, since stable normal directions were insufficiently available The mean and reversed near transitional directions have negative reversal tests (McFadden and McElhinney, 1990); B) Mean non-transitional directions resulting from magnetostratigraphic studies (Zijderveld et al., 1986; Langereis and Hilgen, 1991). Circles (triangles) denote the directions from the studies from Sicily (Calabria). The mean and reversed non transitional directions have negative reversal tests. The mean reversed non transitional directions of each section have a common true mean direction (CTMD) class B; C) same data as if B), after rotating the data in a way that the normal direction are exactly north with inclination 50° The two reversed mean non-transitional directions of the same chronozones have no Common True Mean Directions (CTMD).

The differences in the transitional directions of the two sections as well as the differences in offsets of the mean near-transitional directions (table 1, fig 5a) can be explained by variations in local redox circumstances in the layers in which the transition is recorded. The diagenetic magnetite model cannot account for the differences that are observed between the mean non-transitional directions in both sections, because the remanences of the primary and secondary magnetites are parallel in these parts of the sediment and thus a superposition will change the direction of the natural remanent magnetization. The mean non-transitional directions before and after each transition, determined by averaging the data from the magnetostratigraphic subzones from the studies by Zijderveld et al. (1986) and Langereis and Hilgen (1991) (fig 5b), all have negative reversal tests (McFadden and McElhinney, 1991) (table 1), but the mean non-transitional directions are closer to antipodality in the Calabrian section. If this is related to the observations that in the Calabrian section trajectories over which the transitions take place are shorter, the number of intermediate VGP points are lower and the fact that these intermediate points are nothing more than scatter (Zijderveld diagrams of figs 2,4), then this is a strong argument for the smoothing mechanism as proposed by Langereis et al (1992). The off-sets before and after both Sidufjall subchronozones are in each section significantly the same, since the reversed mean non-transitional directions have a common true mean direction (CTMD) The presence of CTMD was tested in each section was tested by doing the reversal test (see McFadden and McElhinney, 1991) with the reversed mean non-transitional direction before the subchron and the antipole of the reversed mean non-transitional direction after the subchron. The result was that the CTMDs are of class B in both sections. In addition, we rotated the three mean non-transitional directions of each section in a way that the normal directions had declinations that were exactly north and inclinations of 50° (5C), and determined whether the reversed mean non-transitional directions of each parallel subchronozone in the two sections has a CTMD. The result was that in both cases there was no CTMD. This suggests that the deflection from the antipole of the normal mean non-transitional directions have a local cause. This requires a mechanism dependent on local circumstances that deflects the directions of the remanence from the geomagnetic field over much longer trajectories than is expected by the model. In first order approximation the sediments reliable record the geomagnetic field, as is proven by magnetostratigraphic studies that show a convincing fit with the geomagnetic polarity time scale (based on sea-floor anomalies), on a smaller scale, however, the Trubi sediments do not reflect the higher order geomagnetic variations. In addition the reliability of other sediments to record polarity transitions still needs to be demonstrated, for instance by comparison of the sedimentary record with a record from a lava sequence from the same polarity transition.

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Chapter 6

The upper and lower Nunivak sedimentary geomagnetic transitional records from Southern Sicily

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Abstract 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 (LT) and a high temperature (HT) component. The two components do not represent the geomagnetic field because the changes of these components take place at lithological boundaries. In addition, the directional changes do not completely match directional changes of the very same transitions recorded in Calabria, some 250 km away. The character 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 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 a process of smoothing of the stable non-antipodal directions before and after the transitions. 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) as well as Channel et al,. (1982) found the two components residing in magnetite and hematite, 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 magnetite 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 with 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.

magnetostratigraphic results, Hilgen correlated 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

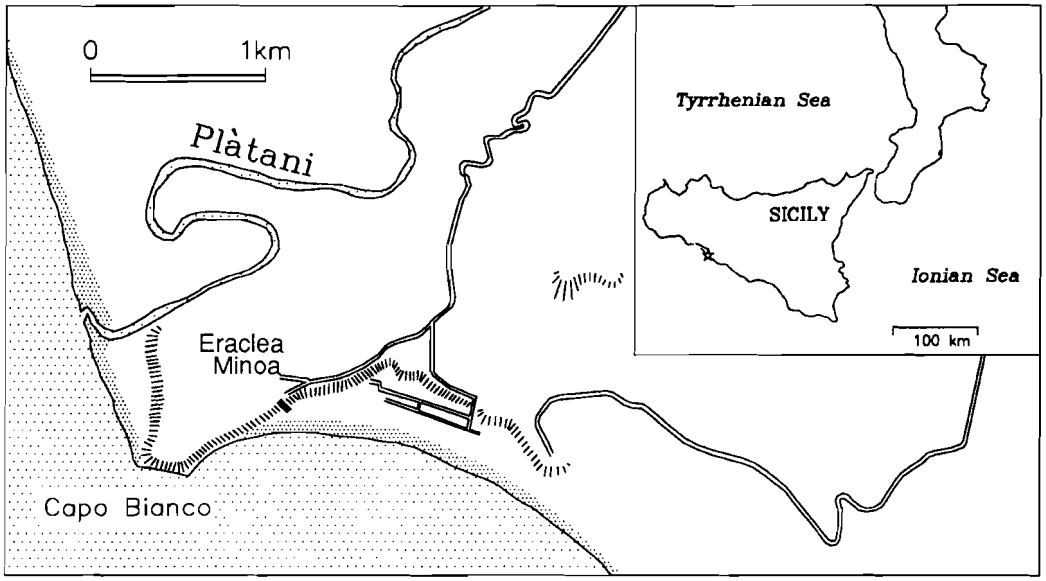

figure 1. Location of the Capo Bianco section in Sicily (Italy). Inset: Map of Sicily and Calabria. The transitions of this paper were sampled at the Capo Bianco section on Sicily and will be compared with the same transitions sampled by Linssen (1991) in Calabria, some 250 km away from Eraclea Minoa.

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 Caltanisetta basin of Southern Sicily (Fig. 1). These 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 cm/ka. The lithology consists of marine marls of the Pliocene Trubi formation (Fig. 2). This formation is composed of sedimentary cycles, consisting of carbonates (60 to 80% $CaCO_3$) 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 is 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

and a dip of 289.5° and 25.4° 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 metre) in order to expose the fresh (blue coloured) sediment. This method proved successful for almost the entire 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).

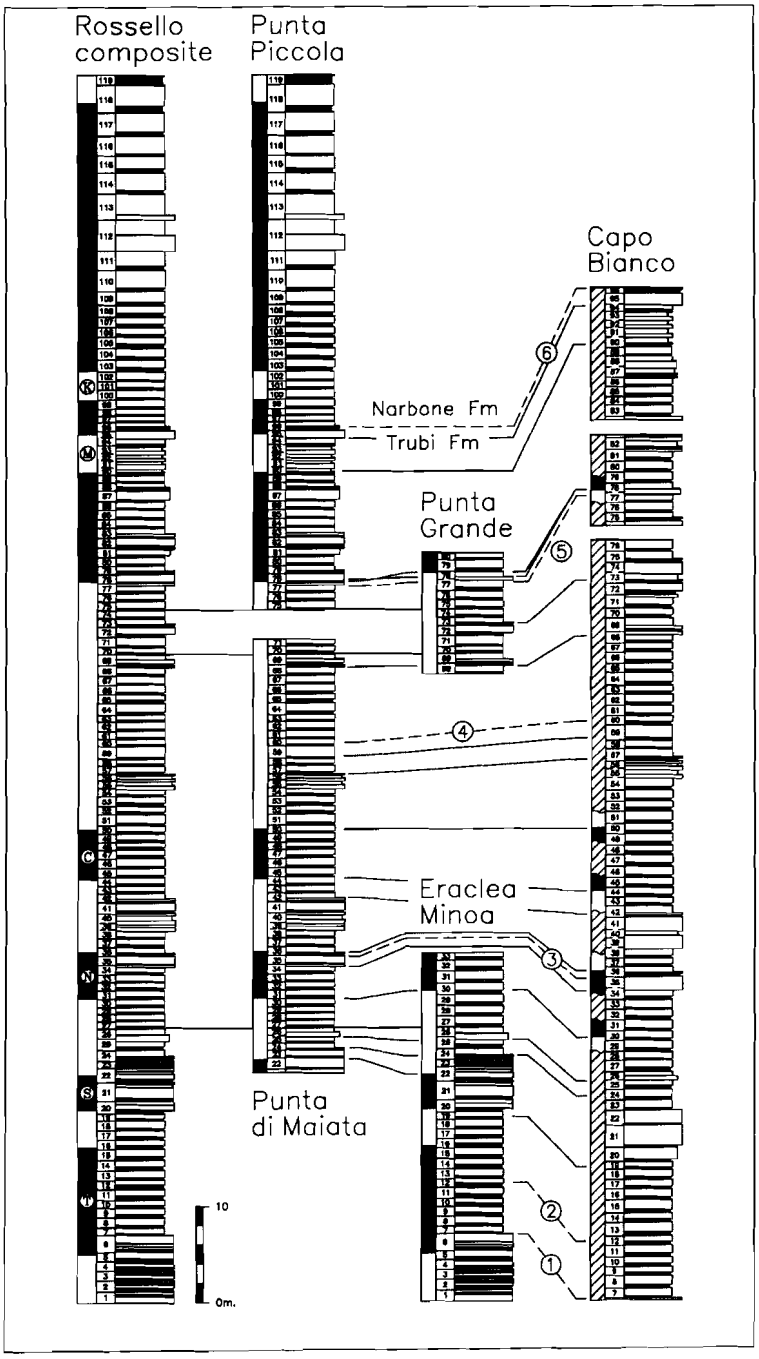

figure 2. Lithostratigraphy and magnetostratigraphy of the Capo Bianco and parallel sections that form the Rossello composite section (Langereis and Hilgen, 1991). Figures in circles refer to occurrences of microfauna. The subdivision of small scale cycles is after Hilgen (1987). The weathering colours (carbonate contents) of each sedimentary cycle are grey (70%), white (80%), beige (60%) and white (80%), whereas fresh colours show mere 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 36 and was sampled over 1 m. In the present paper the zero levels of the two transition records are arbitrarily chosen at pronounced layers parallel to sedimentary lines.

In the magnetostratigraphic study of the Rossello composite section the Characterisic Remanent Magnetization (ChRM) is based on the thermal demagnetization of a magnetite or high temperature (HT) component removed, mainly between 500 and 600 °C (cf. Hilgen and Langereis, 1988). A low temperature (LT) component, demagnetized between 330 and 500 °C, shows the same direction as the HT component well outside a transitional interval. But near a transition, con- 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, since 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 HT component. Rock magnetic

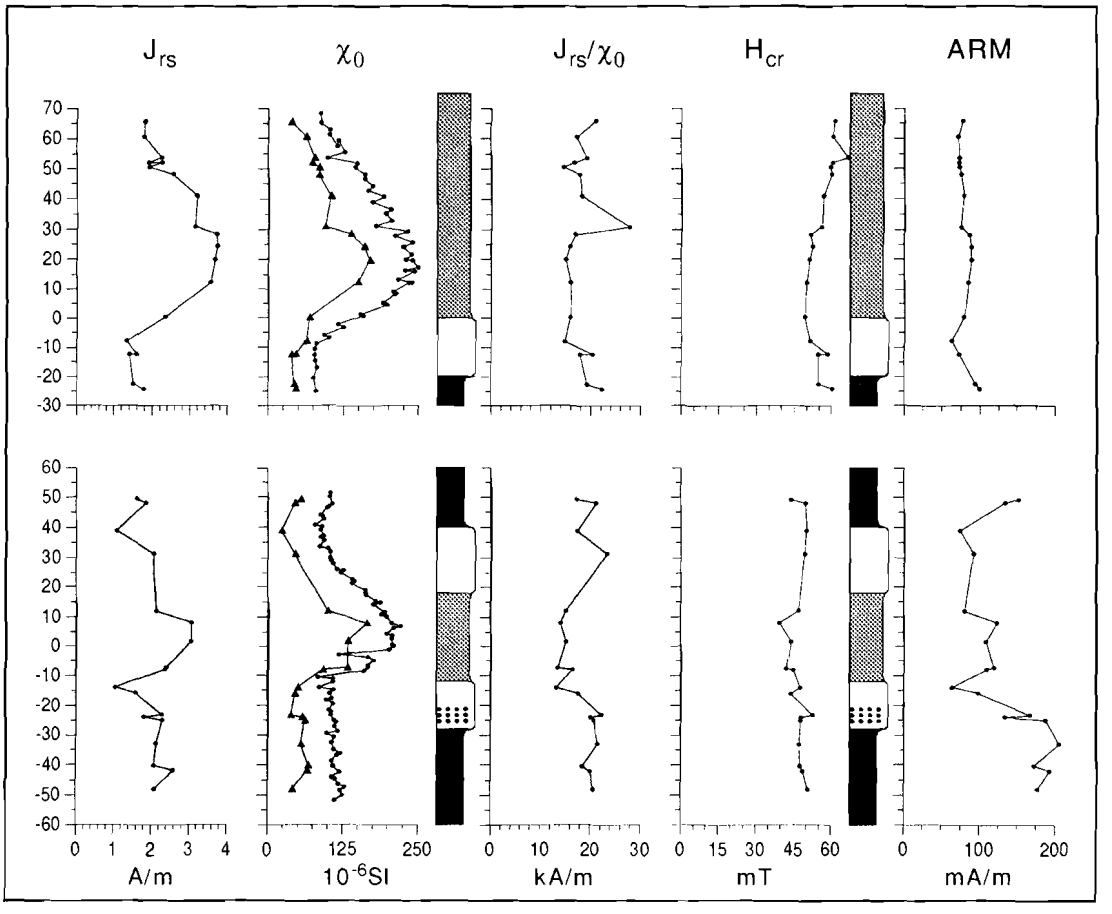

figure 3. Variation of magnetic parameters saturation IRM (J_{rs}), initial susceptibility (χ_0), the ratio J_{rs}/χ_0, remanent coercive force (H_{cr}) and Anhysteretic Remanent Magnetization (ARM) as functions of the stratigraphical levels and lithology. The upper part of the figure is the upper Nunivak 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 χ_0 diagrams are the values corrected for the high field susceptibility (fig. 4). J_{rs} and χ_0 have maxima in the beige parts of the lithology. The ratio of the parameters J_{rs} and the uncorrected χ_0 (J_{rs}/χ_0) lie 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 χ_0.

investigations are needed to determine in some detail the magnetomineralogy. We have determined the initial susceptibility (χ_0) and the (remanent) saturation magnetization (J_{rs}, J_s), the (remanent) coercive force (H_{cr}, H_c) and the anhysteretic remanent magnetization (ARM).

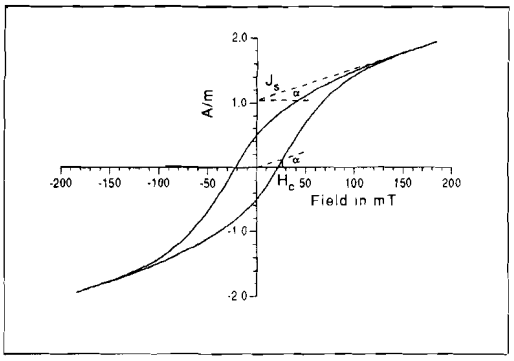

figure 4. Hysteresis loop. Horizontal (vertical) axis: inducing field (induced remanence). The linear trend at higher fields (dashed lines) is due to the paramagnetic susceptibility of the clay minerals. It can be derived from the curve as the tangent (of α). J_s: saturation remanence determined where 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.

The saturation isothermal remanent magnetization (J_{rs}) acquired in a maximum DC field of 2 Tesla (T) shows prominent maxima in the beige layers (Fig. 3a). Minima are usually found in the white layers just above the grey layers. The lithological dependence of the χ_0 (Fig. 3b) is in the UN section similar to that of J_{rs}, with a maximum of $250*10^{-6}$ (SI units) in the beige layer. In the LN section, the maximum of χ_0 in the beige layer is more pronounced than the maximum of J_{rs}. Since the clay fraction of the lithology may give a substantial paramagnetic contribution, it will increase the bulk susceptibility χ_0. We have used hysteresis loop experiments to quantify the amount of the paramagnetic contribution (Fig. 4). Correction for the paramagnetic susceptibility decreases the χ_0 over both entire intervals (triangles in Fig. 3b). The maxima in the beige layers however, are clearly retained and indicate not only an increase due to the clay fraction but also due to ferrimagnetic minerals. Both χ_0 and J_{rs} are dependent on the nature and concentration of the magnetic minerals. The ratio J_{rs}/χ_0 - largely independent of concentration provided that the dominant magnetic mineral is fine grained magnetite as is here the case (van Velzen and Zijderveld, 1990; van Hoof et al., 1992) - may give an indication of grain size changes and the typical values are larger than 20 kA/m up to 30 kA/m.. In this study the ratio shows typical values of about 20 kA/m, but the corrected for the clay susceptibility the ratio will increase some 50 % (Fig. 3c).

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. 3d.

In their rock magnetic study of the Trubi marls of Eraclea Minoa, Van Velzen and Zijderveld (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 of the present study fit the values from Van Velzen and Zijderveld (1990) extremely well (Fig. 5).

In those parts of the record where the LT and HT component have completely different directions, the two components can easily be separated, resulting in two separate decay curves (Fig. 6b). The decay of the NRM of the other samples were not separated in a LT and HT part but these curves also suggest the presence of a LT component by the two fold decay (Fig. 6b). To investigate the presence of a LT component in the entire section the thermal decay of the NRM was compared with the thermal decay of the Anhysteretic Remanent Magnetization (ARM; the remanence acquired in a 0.037 mT DC field superimposed on an AF field of maximum 250 mT). 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 χ during thermal demagnetization was monitored. The relative temperature dependence of χ as well as ARM was identical for the three levels. The ARM decreases with a plateau in the decay curve between 300 and 400 °C, similar to the NRM decay curve. The χ increases from 350 to a maximum at 450/500 °C, followed by a relative minimum at 550 °C and a strong increase at higher temperatures. The maximum at 450/500 °C might be due to an alteration 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 Zijderveld, 1992). Van Velzen and Zijderveld (1990) found in the samples from the Eraclea Minoa section no indication for any sulphur bearing magnetic minerals (i.e. greigite or pyrrhotite) in the Trubi sediments.

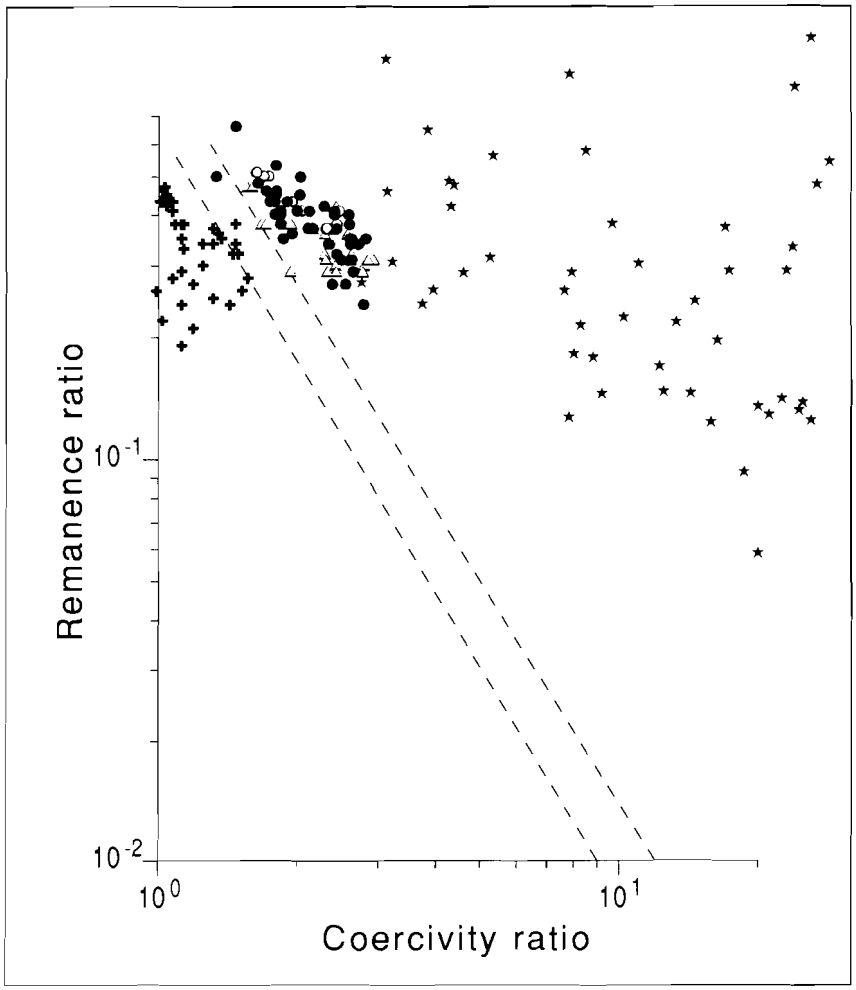

figure 5. Double logarithmic plot of coercivity ratio H_{cr}/H_c versus remanence ratio J_{rs}/J_s after van Velzen and Zijderveld (1990). Literature data for magnetite of known grain sizes fall on a single trend indicated by two dashed lines. Stars (crosses) denote data from goethite (pyrrhotite) (Dekkers,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

4. NRM components

The LT and HT components are most clearly seen in the diagrams where they are not parallel (Figs 7b and 7c, d, h, i) and the two-fold decay of the two NRM components is depicted in Figure 6b. During demagnetization of the LT component, the direction of the remanence in some samples start fluctuating at temperatures of 330 to 390 °C (Fig. 7h). At this trajectory the susceptibility starts increasing (see rock magnetic paragraph) so the fluctuations are most likely caused by oxidation of an iron-sulphur (pyrite) mineral and the subsequent formation of (SP) magnetite (van Velzen and Zijderveld, 1992). Where possible, directions of the LT and the HT components have both been determined.

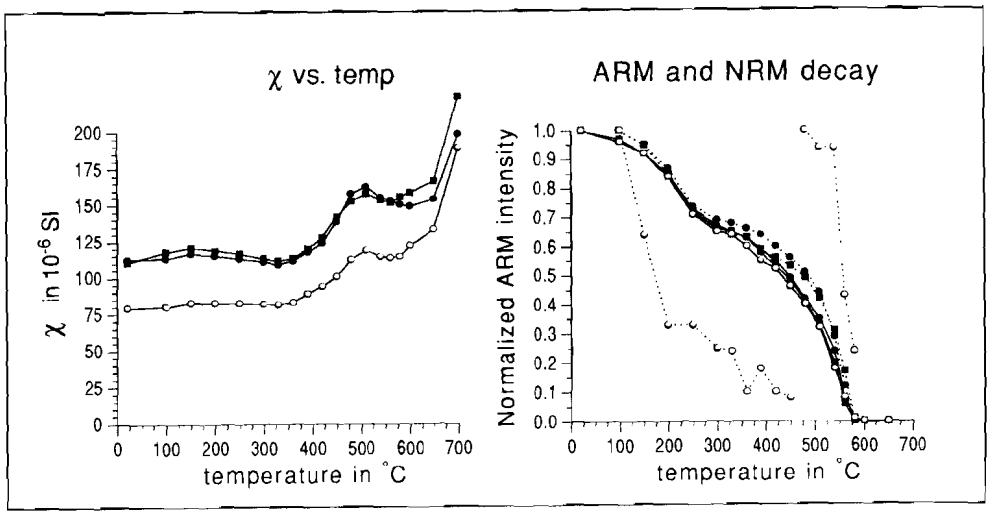

figure 6. Temperature dependency of χ, ARM and NRM. Black dots and squares are from levels -40 and -25 cm respectively from the lower Nunivak section where the HT and LT components were parallel. Open circles show the measurements from a sample from level -12 where LT and HT were anti-parallel. 6a: Changes in χ are identical in the tree samples. Increase at 450 °C indicates the presence op sulphur bearing minerals. (Dekkers, 1990). Increase at 650 °C indicates the generation of super paramagnetic magnetites (Van Velzen and Zijderveld 1992). 6b: Dotted line: NRM decay. Continuous line: ARM decay. The NRM of the samples from level -12 was easily separated in two anti parallel components. Decay of ARM and NRM are identical for all levels.

The demagnetisation diagrams show that the non-transitional reversed and normal polarity directions (Fig. 7a, b, e, f, g, i) show a tectonic clockwise rotation of the Caltanisetta basin as observed in the other Trubi sections on Sicily (Langereis and Hilgen, 1991) and is a strong indication of the primary nature of both the LT and HT components. Also 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 figure 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° to 360° back to 35° in the white layer, between levels -25 cm and 5 cm. At the -25 cm level also the intensities of both components collapse. A sudden jump to south/up directions is observed at level -5 cm. 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° in declination and 0° 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 level 40 cm followed by a decrease. The transition of the LT component is initiated at level 20 cm and it is finished at level 50 cm. The LT component does not reach a completely reversed direction, it shows more scatter and its intensities, collapsed at level -20 cm, remain low.

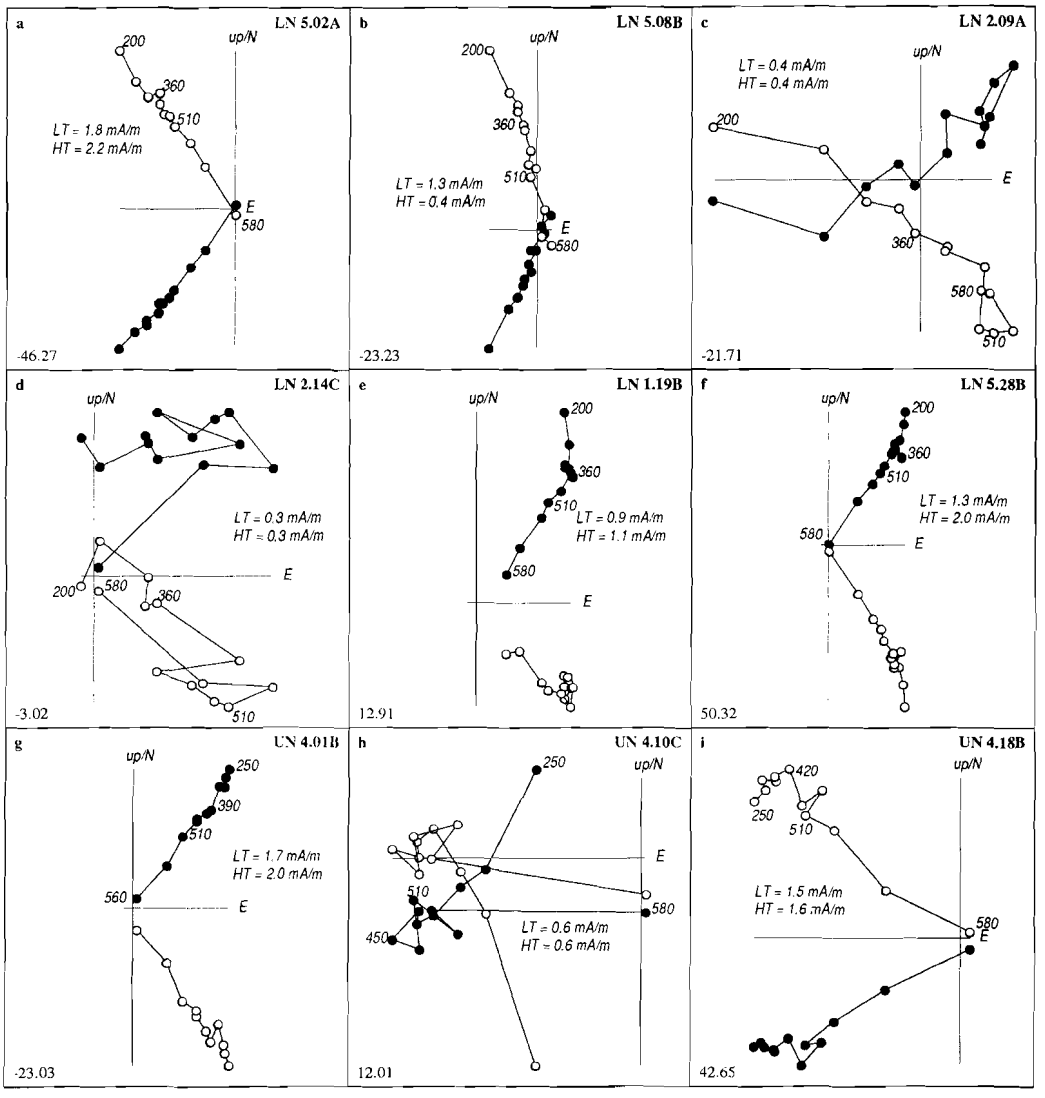

figure 7 Representative thermal demagnetisation diagrams of the lower (LN samples figs 7a-7f) and upper (UN samples figs 7g-7i) Nunivak reversals. Stratigraphical level (down left) refers to the stratigraphic columns of Fig 4. Solid (open) symbols are horizontal (vertical) projections. Temperature steps below 200° C are not shown to enhance the details 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 idem, 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 north 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.

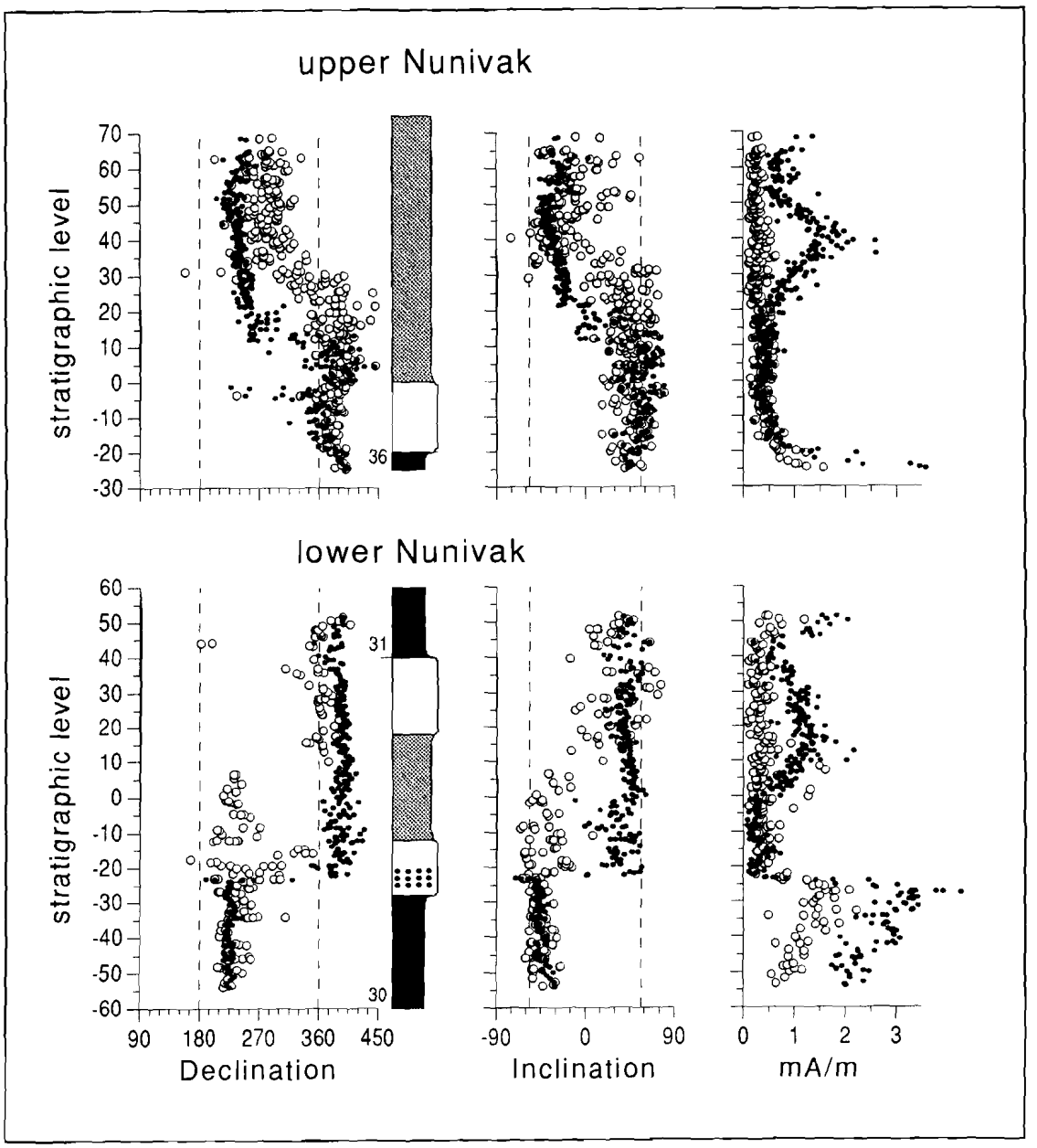

figure 8. Records of the declination, inclination and intensity obtained after thermal demagnetisation. For legend of sedimentary column: see fig 3. Figures along the stratigraphic column refer to the small scale sedimentary cycles (Hilgen, 1991). Circles denote the LT component; Black spots the HT component. Dashed lines indicate declination and inclination (57.5°) 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 abruptly 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 fast. 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.

6. Discussion

6.1 The Sicilian lower and upper Nunivak records.

The reliability of the NRM of sediments to represent the directions of geomagnetic field during its reversals is subject to discussion (van Hoof and Langereis, 1991; Langereis et al., 1992; van Hoof et al., 1992; Mary and Courtillot, 1992; Quidelleur et al., 1992). Because the (non-transitional) directions of both the LT and HT components 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 short time after sedimentation. However, their transitional characteristics of the LT and the HT components 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 this 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 also the intensities of either component 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, because 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, 1988; 1991); they were sampled at the Monte Singa section in southern Italy. The grey to white-beige-white sequence in the Sicilian Trubi is equivalent with 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. 9a). 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 is in the Calabrian upper Nunivak 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 summarised below:

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

The diagenetic conditions in the sediment varied with the amount of metabolisable organic carbon (Froelich et al., 1979). During or very shortly after deposition (primary) magnetite was formed. The diagenetic conditions were the most reducing in the grey layer where sulphate reduction took place. However, magnetite was probably preserved because HS^- react more easily with the less stable ironhydroxides (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 an $white_{(1)}$ layer.

After burial of the grey/$white_{(1)}$/beige/$white_{(2)}$ 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 Fe^{2+} migration from the grey layer.

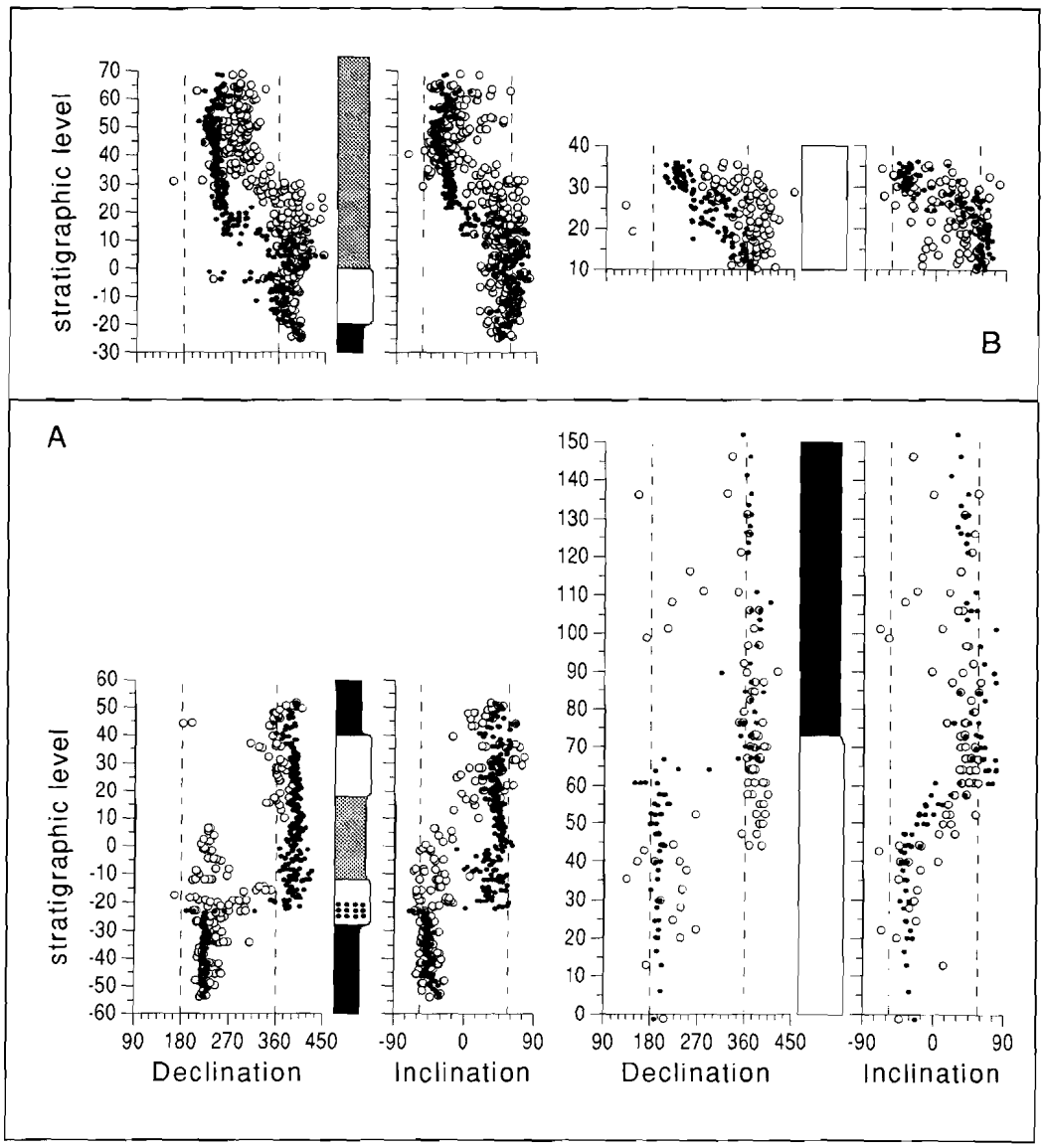

figure 9. Comparison of the two transitions from the present paper at the left hand side and the same ones from the Calabrian Trubi formation (Linssen, 1988; 1991) at the right hand side. A: Lower Nunivak: the vertical scale has been changed to calibrate the same lithological boundaries from white to 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

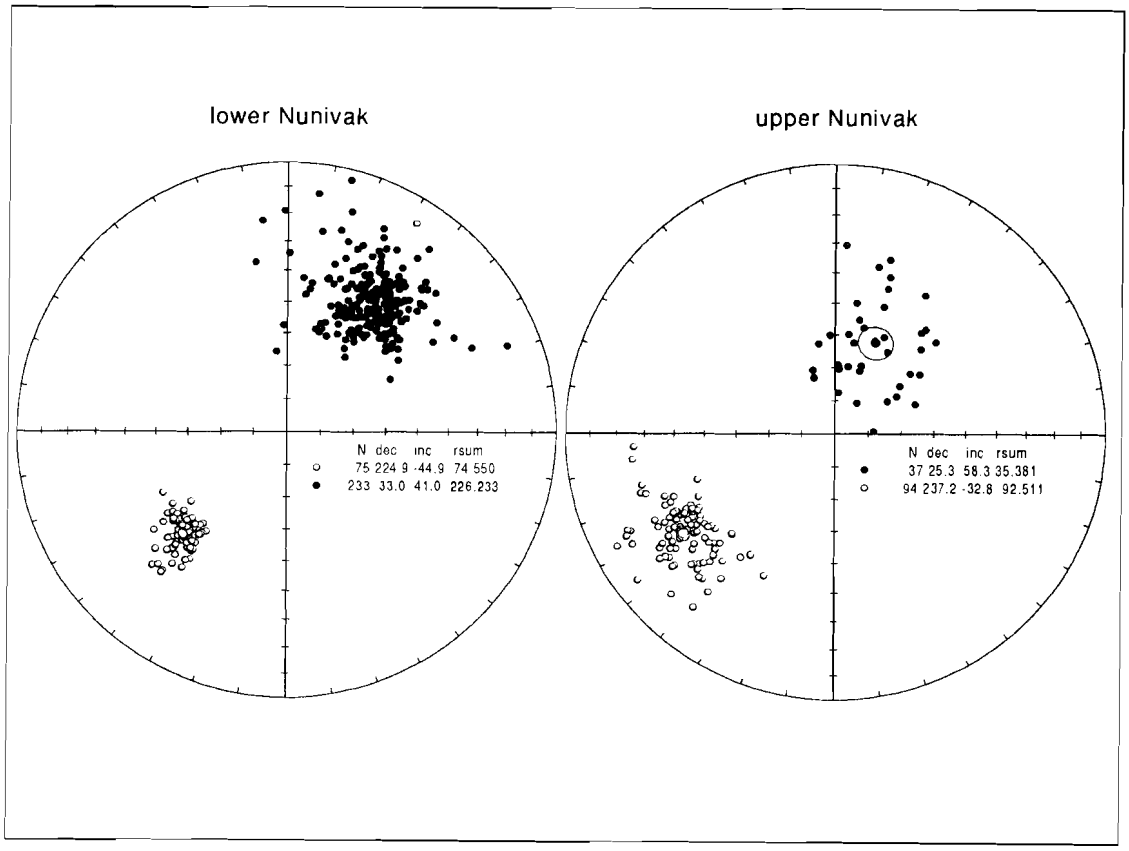

figure 10. Stable HT directions from the two transitions determined in the LN record below level -23 cm (before the transition) and above level -19 cm (after) and in the UN record between levels 0 and 8.5 cm (before) and above level 35 cm (after). The means of the stable directions before and after the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

This implies that in these sediments there is a time lag of at least one cycle white$_{(1)}$/beige/white$_{(2)}$ 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' fits well 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$_{(1)}$ 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$_{(1)}$ 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^{2+} 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 third 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 directions before and after a transition also result in a VGP path with a similar longitudinal confinement (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 vectorsum of pre- and post-transitional magnetisations (and even true intermediate directions) leads to smoothing of which the extent 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 since 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).

subchron, age in Ma	DEC	INC	N	R
4.85-4.65	223.2	-44.9	13	12.79
Nunivak	31.5	50.9	12	11.93
4.53-4.35	214.7	-41.0	17	16.78

Table 1: Mean stable directions and ages of the polarity zones before, during and after the Nunivak subchronozone after Langereis and Hilgen (1991). N: number of samples, R: unit vector sums.

Mean stable directions of the polarity zones before, during and after the Nunivak subchronozone (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,1991), thus indicating a significant non-antipodality.

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 America's 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

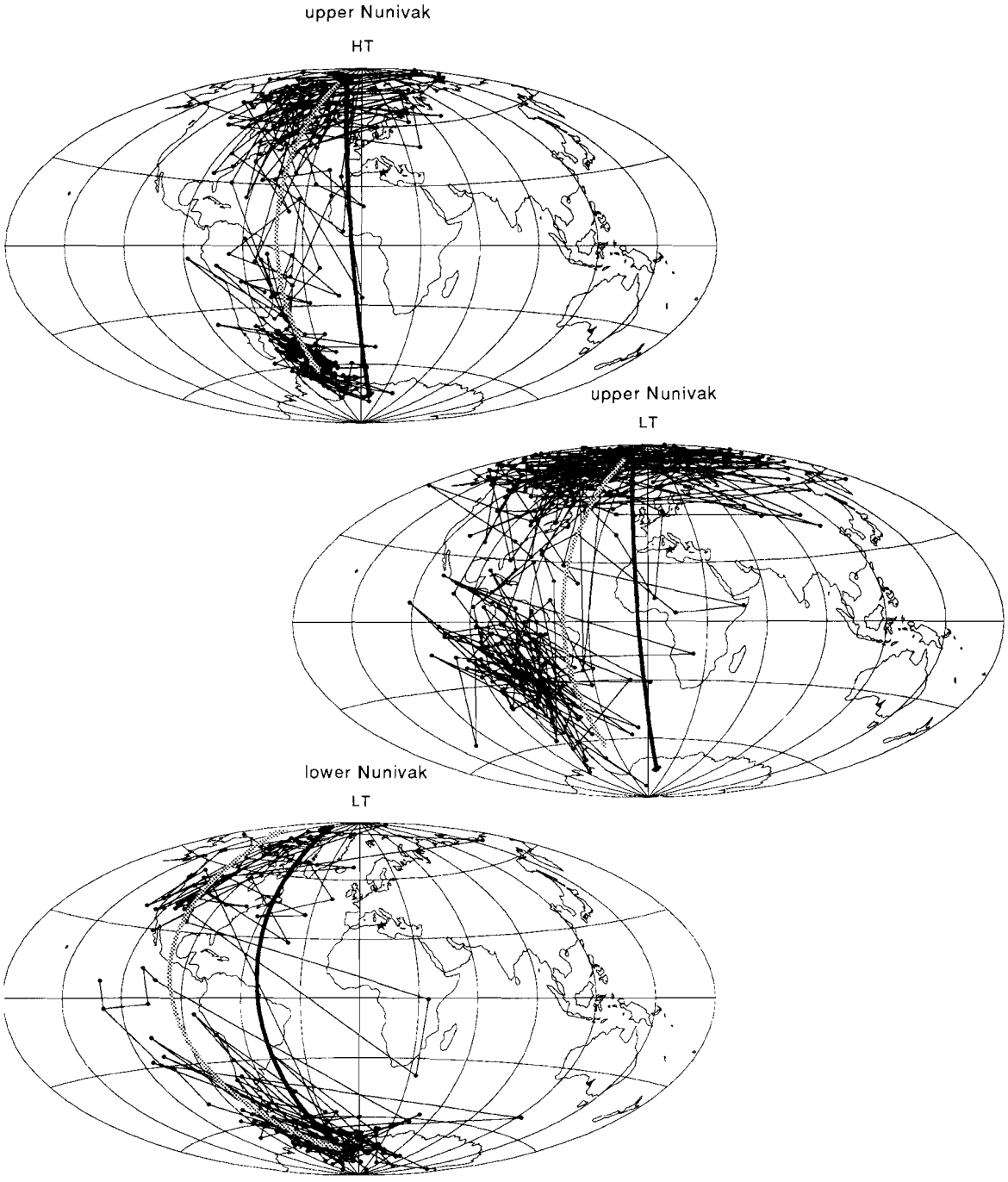

figure 11. VGP paths of the lower and upper Nunivak transitions. The solid black line shows the VGP path obtained by filtering the mean directions of under/overlying polarity zones resulting from magnetostratigraphy (Langereis and Hilgen, 1991), the shaded line represents filtering of near-transitional 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 are even better.

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 does therefore 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 the 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). Due to the smoothing, a long lived cluster of intermediate VGPs as found by Hoffman (1992) during the transition cannot be excluded.

Acknowledgements

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Chapter 7

A paleomagnetic record of the upper and lower Cochiti reversals from Southern Sicily (Italy)

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Abstract. A detailed paleomagnetic study of the upper and lower Cochiti reversals from marine marls on southern Sicily show that both are recorded by a low (LT) and a high (HT) temperature component. TRM experiments reveal that the small scale directional changes observed in the demagnetisation diagrams are no components of the NRM but that they are caused during demagnetisation. These components do not influence the directions at higher temperatures and show a logarithmic relation with temperature. The reversals are recorded as very complex transitions with directional changes some of which coincide with distinct lithological boundaries. The results from the upper Cochiti record are compared with the results of the same transition [1] sampled in the same lithology, but in a parallel section some 30 km east. An interval with normal directions from this study is recorded over a longer trajectory than in the parallel section. An earlier proposed diagenetic magnetite model [1] to account for lithology related directional changes in reversal records cannot explain all observations of the lower Cochiti record.

1. Introduction

In the study of the behaviour of the geodynamo during a change in polarity, sediments are generally assumed to show a more or less continuous record. However, the reliability of sediments to record the polarity transitions are increasingly subject to discussion. Apparently, due to sedimentary filtering artefacts in the remanence are introduced [2,3]. Langereis et al. [4] smoothed stable non transitional directions before and after reversal records from Mediterranean marine marls and the resulting synthetic virtual geomagnetic pole (VGP) paths show a striking coincidence with the observed VGP paths. They concluded that the intermediate directions could also be the result of smoothing, and are not necessarily representations of the Earth's magnetic field during the transitions. Not all information of transitional geomagnetic field behaviour will be obscured however, since a low frequency signal of the geomagnetic field will remain visible in the sedimentary record, provided the filter width is small with respect to the signal. For instance if the clusters of long-lived intermediate polarity states as found by Hoffman [5] in lava sequences are indeed features of the geomagnetic transitional fields, smoothing of a geomagnetic field like this will likely result in a VGP path that follows the longitude of this long lived polarity state.

Some reversal records have shown that a sedimentary NRM acquisition not only filters directional changes, it also may record a polarity more than once by different magnetic components that have varying lock-in depths [6,7,8]. This implies that the different components at the same stratigraphical level have acquired their remanence at a different times. While Dijksman [6] and Channel [7] found two such components residing in magnetite and hematite, van Hoof and Langereis [8] found the two components - a high-temperature (HT) and a low-temperature (LT) component - both residing in magnetites [1]. To complicate

matters even further, the varying lock-in depth may also cause one transition to be recorded at different sedimentary levels by the same component [1], and in that case the transition is preceded by an apparent 'excursion'. The varying lock-in depths, the filtering and the multiple registrations of the same transition can (in a first-order approximation) be explained by a diagenetic magnetite formation model [1]. This model is based on the geochemical properties of the sediment and describes the superposition of (early diagenetic) 'secondary' magnetites on 'primary' magnetites, even to the extend of complete substitution. The primary magnetite acquired its remanence very shortly after the deposition of the sediment, the secondary magnetites are formed under suboxic conditions after burial of an organic-rich layer.

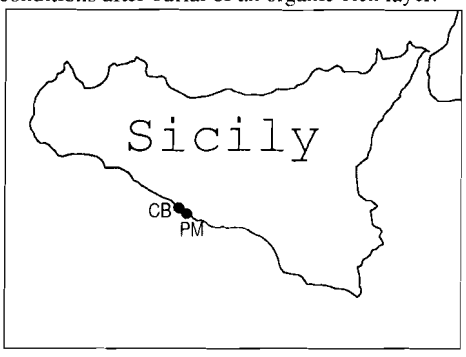

figure 1: Locations of the Capo Bianco (CB) section where the present upper and lower Cochiti records have been sampled, and of the Punta di Maiata (PM) section.

The fact that the HT and LT components have variable lock-in depths was observed in two reversed (R) to normal (N) polarity transitions, where the LT component clearly acquired its remanence before the HT component [8]. One of these two R-N polarity transitions was the lower Cochiti transition. In this paper we present the extended paleomagnetic, rock magnetic and geochemical results of this transition, together with the results of the upper Cochiti polarity transition, both sampled in the Capo Bianco section on Southern Sicily (fig. 1). In addition, we will compare the paleomagnetic and geochemical results from the upper Cochiti polarity transition with the results obtained from the upper part of the same polarity transition [1], sampled in the Trubi sediments of the Punta di Maiata section, some 30 km southeast of the Capo Bianco section.

2. Geological setting and sampling

Our current research concerns the reversals records - from the lower Thvera in the Gilbert Chron to the upper termination of the Olduvai in the Matuyama chron - that are located in Southern Italy. The Cochiti subchronozone is recorded in the Rossello composite section, which consists of the marine marls from the Trubi formation [9]. These sediments show a distinct cyclicity in both $CaCO_3$ content and weathering profile, a cyclicity which is related to the precessional cycle of the Earth's orbit [10]. The cycles range from number 1 at the Miocene/Pliocene boundary at 5.32 Ma, to number 119 at 2.3 Ma. Recently, Hilgen [11] has established an astronomical polarity time scale by correlating these sedimentary cycles to the astronomical solutions of the past 5 Myr [12]. The Capo Bianco section in which we sampled the two transitions is a small prominent cape in a series of cliffs along the southern coast of Sicily (fig. 1). This section forms the part of the Rossello composite section [9]; the bedding plane has a strike and dip of 281.0°/10.8°N (upper Cochiti) and 273°/19.6°W (lower Cochiti). The Trubi marls consist mainly of carbonates (60 to 80% $CaCO_3$) and a mixture of clay minerals; small scale sedimentary cycles are quadripartite and show a distinct grey-white$_{(1)}$-beige-white$_{(2)}$ colour layering (fig. 2). The average thickness of these cycles - in which the grey and beige marls represent the less indurated, $CaCO_3$-poor beds - is approximately 1 metre. The (midpoints of) individual grey layers have been correlated to individual minima of the precession index [12], providing a high-resolution time scale and a means to determine the average sedimentation rate per cycle; which is 4.5 to 5.5 cm/kyr [11].

Although the weathering profile shows sharp changes in colour and induration, the changes in the fresh, unweathered sediment are much more gradual. Often, the bottom part of the white$_{(1)}$ layers shows brown oxidation spots, even in the fresh marls. These spots appear to be important with respect to both paleomagnetic and geochemical properties, as will be discussed later.The lower Cochiti record comprises the complete interval between the centre of white$_{(2)}$ of cycle 43 and the centre of grey 46; the upper Cochiti record contains grey 49 to the centre of beige 50 (fig.2) [11]. In each record the zero-levels were defined at a distinct boundary between two layers.

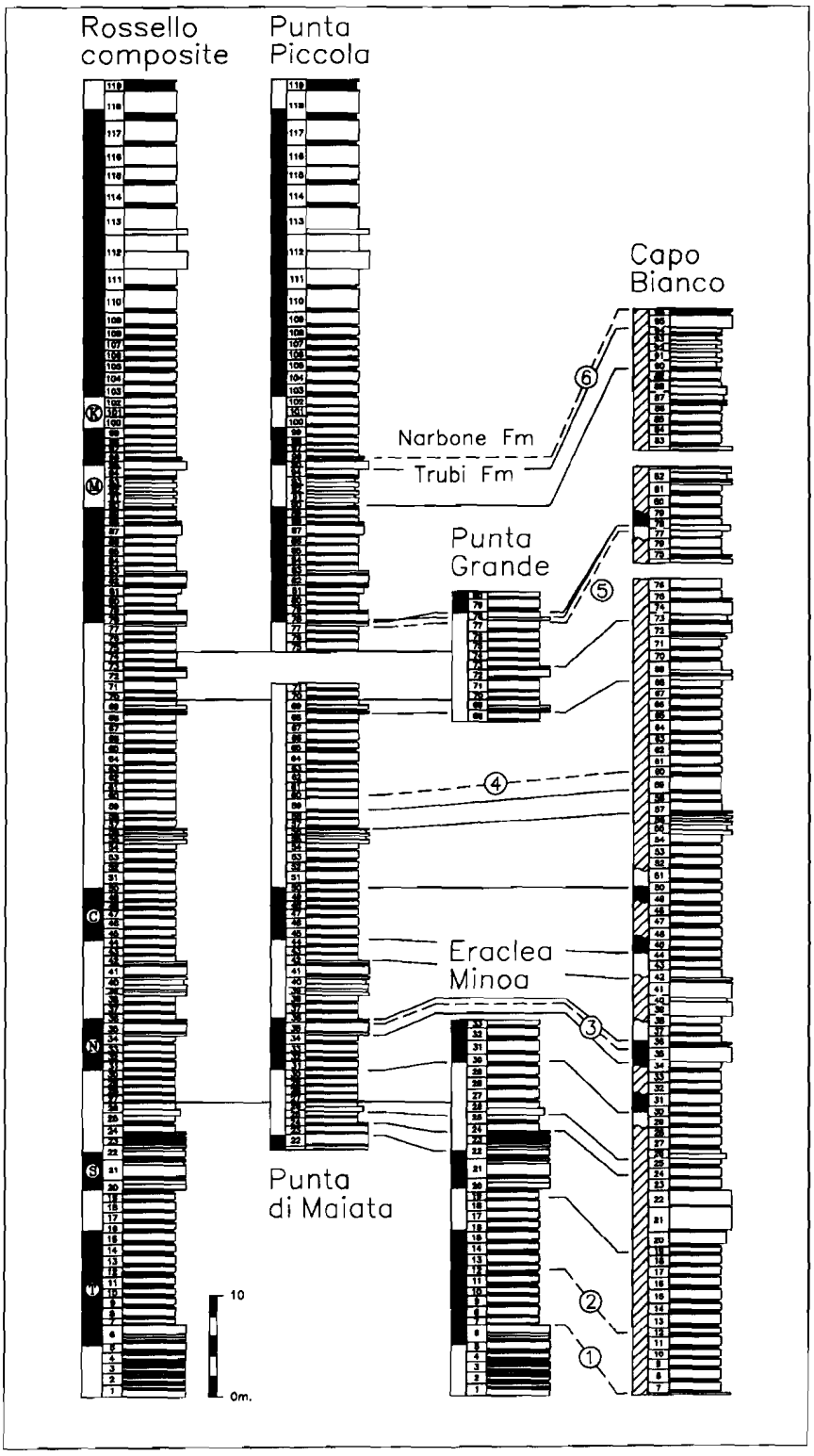

figure 2: Magnetostratigraphy and lithostratigraphy of the Capo Bianco section and the parallel section of Punta di Maiata. In the Capo Bianco section, the lower Cochiti transition record was sampled from white$_{(2)}$ 43 to grey 46, the upper Cochiti transition from grey 49 to beige 50. In the Punta di Maiata section, the upper sampled interval comprises the layers from grey$_{(2)}$ 49 to grey 53.

3. Rock magnetic properties

Earlier rock magnetic investigations of the Pliocene Trubi marls of Sicily [1,13,14] have revealed that the LT and HT components are carried by magnetites. The results of the rock magnetic and geochemical measurements in the upper Cochiti transition in the Punta di Maiata section have led van Hoof et al., [1] to propose an early diagenetic magnetite model that explains the observed magnetic record in terms of paleoredox conditions. In order to test the consistency of their observations throughout the lithology and consequently test their model, we have measured the rock magnetic properties of the lower Cochiti transition record.

Initial susceptibility (χ_0) shows a very obvious relation with the lithology; there are clear maxima in the beige layers (fig. 3a) and therefore this parameter follows the cyclicity of the lithology [1]. A part of χ_0 is due to the paramagnetic contribution of clay and super paramagnetic minerals.

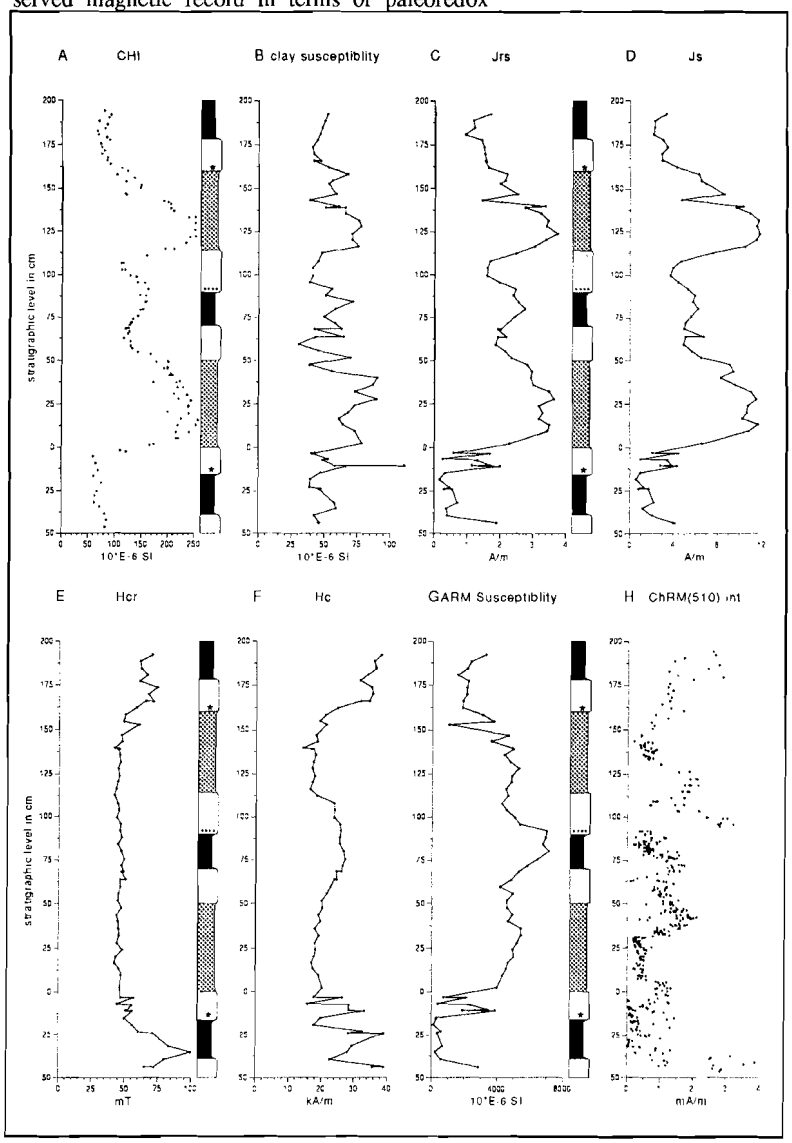

figure 3. Rock magnetic parameters of the lower Cochiti transition record from the Capo Bianco section. In the lithological column, white denote the white layers; black the grey layers and shaded the beige layers. Black dots in the white layers represent layers with brown spots; asterisks represent layers with a vague brown color. a) initial susceptibility b) high field susceptibility; c) remanent saturation magnetization J_{rs}; d) saturation magnetization J_s; e) remanent coercive force H_{cr}; f) coercive force H_c; g) susceptibility of ARM; h) intensity of the HT component.

The maxima in the beige layers of χ_0 are not compensated by maxima in the high field susceptibility (fig. 3b), therefore the shape of the χ_0 curve is due to magnetic minerals. This is confirmed by the variations in the remanent saturation magnetisation J_{rs} (fig. 3c) and in the saturation magnetisation J_s (fig. 3d). The trend of the remanent coercive force H_{cr} (fig. 3e), and the coercive force H_c (fig. 3f) is quit different from that of χ_0, J_{rs} and J_s.

The maximum in H_c in the centre of the record is not present in H_{cr}. The susceptibility of anhysteretic remanent magnetisation (χ_{ARM}), proportional to the ARM intensity, is not proportional to J_{rs} and J_s. It shows a clear maximum in the centre part of the record (fig. 3g). The HT intensity (fig. 3h) has no relation to any of the rock magnetic parameters. During thermal demagnetisation the direction of the remanence often starts to fluctuate. To investigate whether these fluctuations are real components of the NRM, we have thermally demagnetised pairs of specimens from the same core. The NRM of one specimen was completely demagnetised in an alternating field (AF) before the thermal demagnetisation. The diagrams of the thermal demagnetisation of the NRM (fig. 4a) were smoothed by fitting three straight lines through the points of respectively HT, LT and viscous component (fig. 4b). Subsequently, thermal demagnetisations of the AF demagnetized specimens (fig. 4c) are superposed on the smoothed diagrams. This results in a diagram (fig. 4d) with fluctuations that are very similar to the initial, unsmoothed diagram (fig. 4a). We therefore conclude that the fluctuations of the remanence are not related to components of the NRM but they are introduced during thermal demagnetisation.

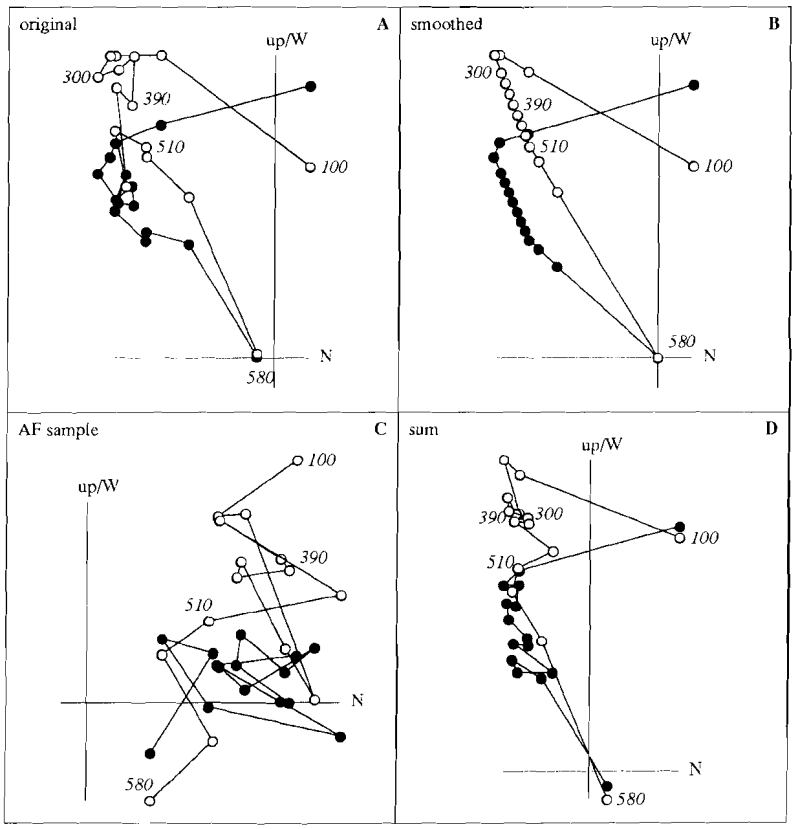

figure 4. A) original thermal demagnetization diagram, with reversed LT and HT directions. B) the same diagram after fitting three straight lines through respectively HT, LT and secondary component thermal trajectories. C) thermal demagnetization of a specimen that was demagnetised in an alternating field. D) same diagram as in figure B, with demagnetisation results of figure C superposed on it.

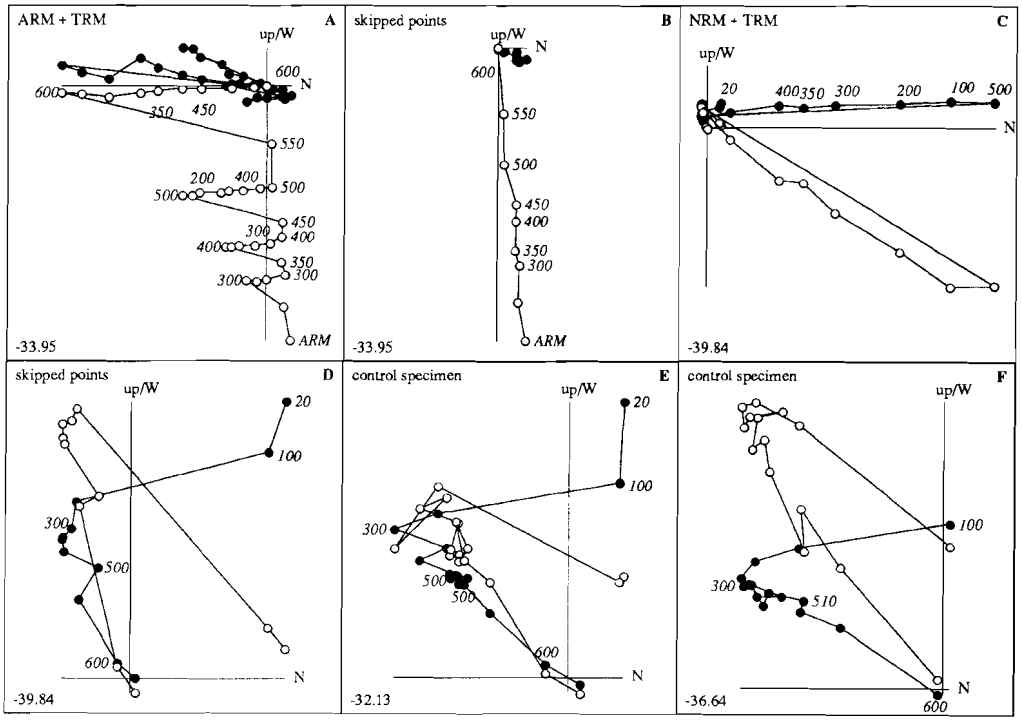

figure 5. A) Thermal demagnetisation of ARM. ARM direction is steep/down. After the first heating steps of 300°, 400°, 500° and 600°C, the cooling took place in a controlled (the Earth's) magnetic field. Direction of the field is South/shallow. After cooling in the controlled field, the previous demagnetisation steps were repeated. B) Same diagram as figure A, the demagnetization steps that were repeated have been omitted from the diagram. The direction of the ARM is not changed by the induced TRM direction. C) Thermal demagnetization of a reversed NRM. Same procedure as in figure A was followed, but only the usual demagnetisation steps, the cooling in the controlled field after 500°C and the subsequent demagnetisation of the induced TRM are shown. Controlled field direction was North/down. D) Same specimen as figure C, but the TRM at 500° and its demagnetisations have not been plotted. E) Thermal demagnetization of control specimen with reversed NRM. The same demagnetization steps as in figure C have been used, but no TRM was introduced. The character of this diagram does not differ significantly from the diagram from D. F) Control specimen with reversed NRM during the usual thermal demagnetization. No significant difference is observed with diagram D. Diagrams E and F indicate that the TRM does not influence the NRM at higher temperatures.

Heider and Dunlop [15] demagnetised ARM remanences and gave the specimen a thermoremanent magnetisation (TRM) by cooling the specimen after one of the demagnetisation steps in a controlled magnetic field perpendicular to the ARM direction. A part of the specimens acquired a TRM, with a direction intermediate between the ARM and the controlled magnetic field direction. Hence, if a component that is introduced during thermal demagnetisation, influences the remaining ARM components, it follows that the fluctuations we found during demagnetisations most probably influence the NRM component. To study this in more detail, we have demagnetised a set of specimens in which we first introduced an ARM (0.038 mT DC magnetic field superposed on an alternating field). The demagnetisation steps were from 200° C up to 600° C with increases of 50° C. At the steps 300, 400, 500 and 600° C we cooled the samples in a controlled magnetic field, instead of in a zero field and we repeated the preceding demagnetisation steps. Control specimens from the same cores or from the same stratigraphic level and containing an ARM or NRM, underwent the same demagnetisation treatment, but the cooling steps of these specimens all took place in a zero field. Figure 5a shows the demagnetisation of ARM with the cooling steps in the controlled magnetic field, figure 5b shows the demagnetisation of the same specimen, but the demagnetisation steps of the induced TRMs have been omitted. It is clear that the TRM introduced by cooling

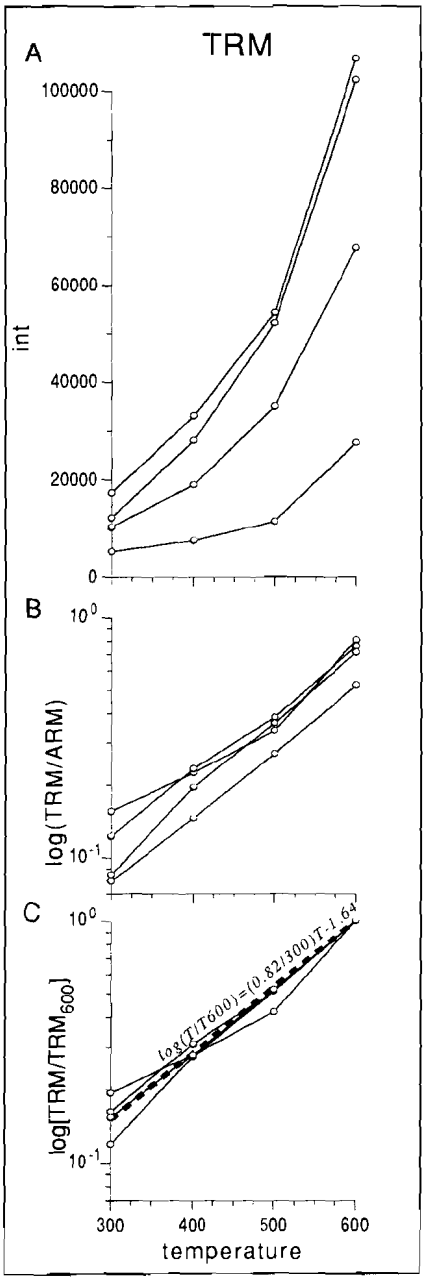

figure 6. A) temperature dependency of TRM intensity. B) Idem for $^{10}\log(^{TRM}/_{ARM})$. TRM intensity at 600°C is some 0.7-0.8 times the ARM intensity at 20°C. C) Idem for $^{10}\log(^{TRM}/_{TRM(600°C)})$, where a linear increase with the temperature is observed.

after the steps 300, 400, 500 and 600° C in the Earth's magnetic field does not influence the ARM direction because the points after demagnetisation of the TRM have the same directions as the initial ARM direction. The procedure was repeated with specimens having an NRM instead of an ARM. The TRMs induced after the cooling steps in the Earth's magnetic field are relatively strong (fig 5c), but plotting the demagnetisation diagrams without the (demagnetisation of the) TRM gives similar diagrams as the demagnetisation of the control specimens that had no TRM. It may therefore safely be concluded that the scatter observed during thermal demagnetisations does not influence the NRM direction.

In addition, the absolute intensities of the TRM, introduced after each cooling step in the ambient magnetic field were measured. The TRM intensities increase with increasing temperature (fig. 6a). The TRM intensity at 600°C is 0.7 to 0.8 times the initial ARM (fig. 6b). Normalizing the TRM with $TRM_{600°}$ shows that the logarithm of the normalized curve is linear with the temperature and that the linear relation is approximated by $^{10}\log(TRM/TRM_{600°})=(0.82/300)T-1.64$ (fig. 6c).

The susceptibility increases during thermal treatment at temperatures higher than 300° C (fig. 7) probably due to oxidation of iron-sulphides into magnetites at these temperatures [16].

4. NRM components

4.1 Demagnetisation

Thermal demagnetisation of the natural remanent magnetisation generally shows the presence of three components, as usually found in these Trubi marls [9]. Apart from a small laboratory-induced component removed at 90-100 °C, there is often a secondary component that has a present-day field direction; it is removed at 200-250 °C (fig. 8).

The characteristic remanence of the Trubi sediments consists of two components [8]. A low-temperature (LT) component is removed between 250°C and 450 - 510 °C and a high-temperature (HT) phase usually is removed at 580 °C.

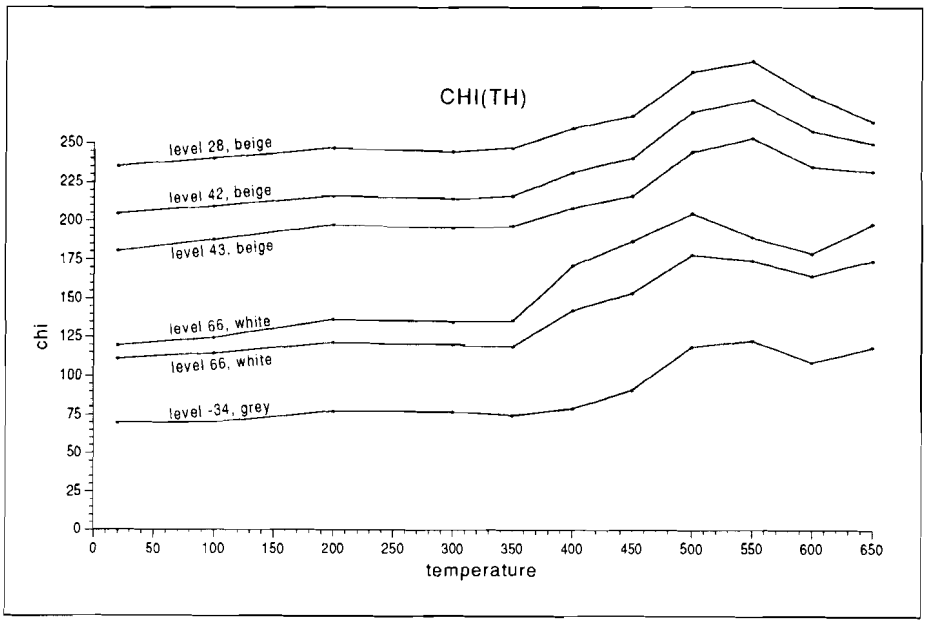

figure 7. Susceptibility as a function thermal demagnetization temperature steps. The increase between temperatures 350° and 550° C is probably due to newly formed magnetites from oxidation of iron-sulphur bearing minerals.

The HT and LT components have acquired their remanences shortly after deposition because the demagnetisation diagrams show that both stable reversed and normal polarity directions have the tectonic clockwise rotation of the Caltanisetta basin (35°) as observed all the other Trubi sections on Sicily, and moreover, the magnetostratigraphic results of the Rossello composite section show a significantly positive correlation with the geomagnetic polarity time scale [9].

4.2 The reversal records

Both transition records show a very complicated pattern of directional changes (fig. 9). The directional changes of the LT and HT components are similar even though there is an often considerable scatter in the LT component.

The lower Cochiti transition, recorded as a R-N-R-N transition is preceded by an R-N-R 'excursion' with declinations up to 330° at level -10 cm, and followed by a N-R-N 'excursion' to west/up directions at level 140 cm. The first R-N transition takes place between levels 25 and 35 cm, via west/shallow directions, the LT component reverses some 15 cm higher in the sediment than the HT component. This observation is not unusual, and has been described extensively by van Hoof and Langereis [8]. The first transition is followed by a gradual steepening to 55° until 60 cm where the directions suddenly swing back to R. The second R-N reversal via west/shallow directions is between levels 75 and 90 cm, again followed by a steepening to 50-60°. After the final excursion, the declinations do not reach the 35° tectonic rotation as would be expected.

Previous studies of the Trubi sediments [17] have indicated that reliable parameters to normalize the intensities are present. We will therefore discuss only the unnormalized intensities. The intensity record of the HT (LT) starts at a maximum of 3-4 (1-1.5) mA/m and decreases to a minimum at level -30 cm. After the excursion at level -10 cm there is a maximum of 1.5 (2) mA/m at level 15 cm followed by a sudden (gradual) decrease in the HT (LT) intensities. Here, the LT intensities are higher than the HT intensities. At the end of the changes in declination during the first R-N transition, the HT intensities increase while the LT in-

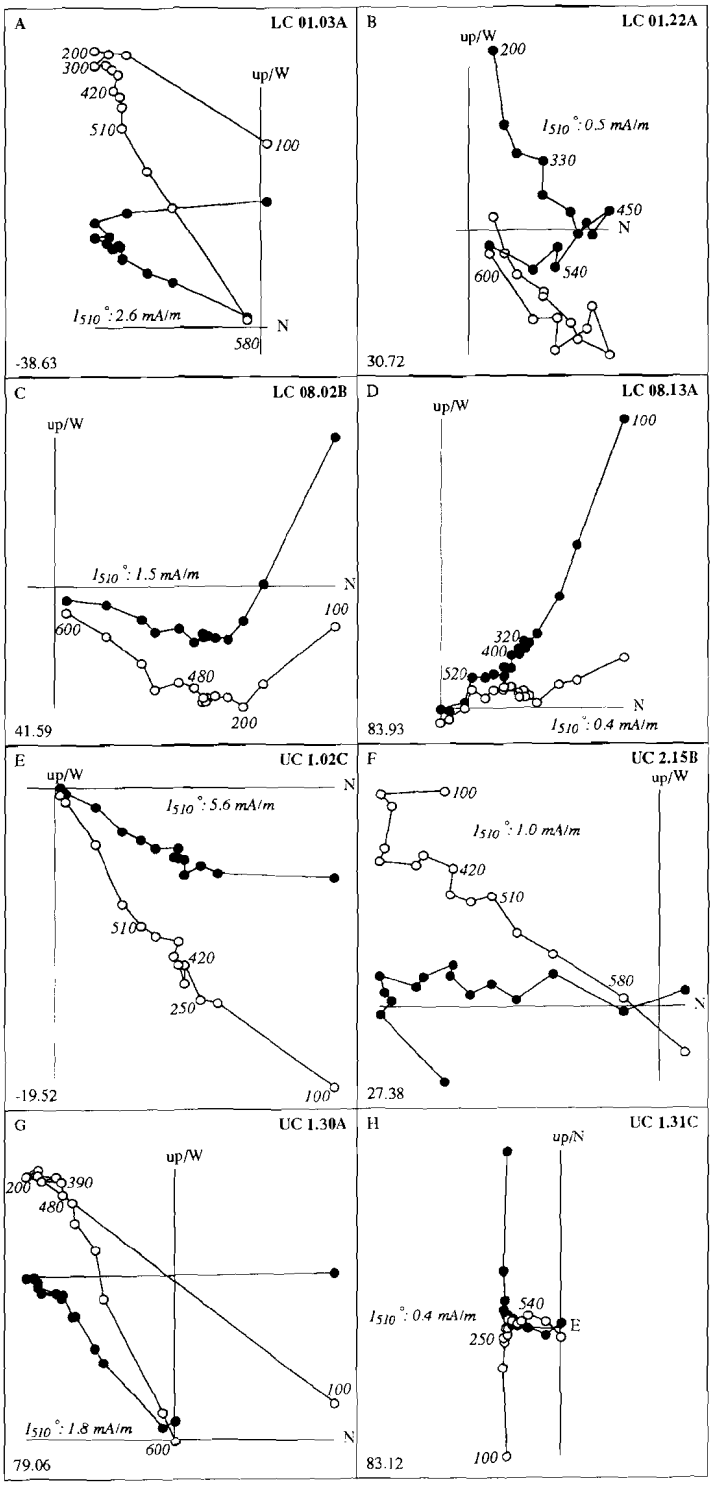

figure 8. Thermal demagnetisation diagrams of the lower Cochiti record (A-D) and upper Cochiti record (E-H). A) Reversed LT and HT components. B) Reversed LT and normal HT component. C) Normal LT and HT components. D) Intermediate components with north/west declinations and negative shallow inclinations. E) normal LT and HT components, F) and G) reversed LT and HT components. H) intermediate directions in up/north projection to enhance the components in directions.

tensities go to a minimum, and the HT intensities are again higher than the LT intensities. After the sudden N-R transition a local minimum at level 65 cm is observed. A local maximum after the second R-N transition is seen at level 95-100 cm, precisely where brown oxidation spots are observed. After this transition the HT (LT) jumps to a maximum of 3 (2) m A/m followed by a local minimum at 110 cm, and an even stronger minimum at the second excursion near 135 cm. At the top of the record the intensities are somewhat lower than at the bottom.

The upper Cochiti transition is recorded as a N-R-N-R-N-R-N-R transition although in the final R-N-R excursion only a few points show clear N directions, especially in declination. At level 0 cm, the declination suddenly jumps to R while the inclination reverses gradually between 0 and 25 cm. At level 33 there is a again fast R-N transition followed by a N-R transition between 50 and 60 cm where the inclination shows a few oscillations between N and R directions. The next R-N transition is between 75 and 90 cm. At level 120 cm, another immediate N-R transition is observed, followed by the poorly documented final R-N-R excursion between 125 and 140 cm. During the transitions observed in the upper Cochiti record the intensities of LT and HT show corresponding minima. During the stable polarity intervals, the HT intensity is higher than the LT intensity.

5. Discussion

The directional behaviour of the transitions is very complex and many of the directional changes closely coincide with lithological boundaries. In the lower Cochiti record, the first excursion takes place in the oxidized level of white$_{(1)}$44, the N-R directional change at level 65 cm is close to the white$_{(2)}$44/grey 45 boundary; In the upper Cochiti record, the fast N-R change of the declination takes place at the grey 49/white$_{(1)}$49 boundary where also brown spots were observed. The R-N changes at 85 cm are at the top of white$_{(2)}$49, and the third N-R change is initiated at the brown spotted grey 50/white$_{(1)}$50 boundary. Therefore, it is almost certain that the directional changes are not caused by a complex geomagnetic transitional behaviour, but they are caused by diagenetic changes that show a distinct different character, depending on lithology. As a consequence, the lock-in depth relative to the sediment water interface may change and the same reversal may be recorded several times [8].

Langereis et al. [4] calculated the mean stable non-transitional directions by averaging the data of the geomagnetic polarity subzones obtained in the magnetostratigraphic studies [9] and the mean stable near transitional directions by averaging the stable polarity data just before and after each transition. Each transition from one mean stable direction to the next was smoothed by a moving window. The resulting synthetic VGP paths showed a striking coincidence with the observed VGP paths. The lower Cochiti and upper Cochiti transition records were considered too complicated to denote intervals from which the mean stable near-transitional directions are determined. In order to approximate these directions, all datapoints in the intensity maxima (i.e. the trajectories in which intensities exceeded 2 mA/m) before and after the transition records were averaged (table 1). The observed and synthetic VGP paths are shown in figure 10. The observed VGP paths of the lower Cochiti transition, after correction for the 35° rotation are strongly confined to North and South America and VGPs are equally distributed along the path. The synthetic VGP paths of the mean stable non transitional directions pass the Atlantic; the synthetic VGP path of the stable near-transitional coincide remarkably well with the observed VGP paths. The VGP paths of the upper Cochiti transitions are confined to an area enclosed by great circles over the east coast of North America and over India. The synthetic non-transitional VGP path is over the Atlantic while the synthetic near-transitional VGP path forms the western limitation of the area with observed VGP paths.

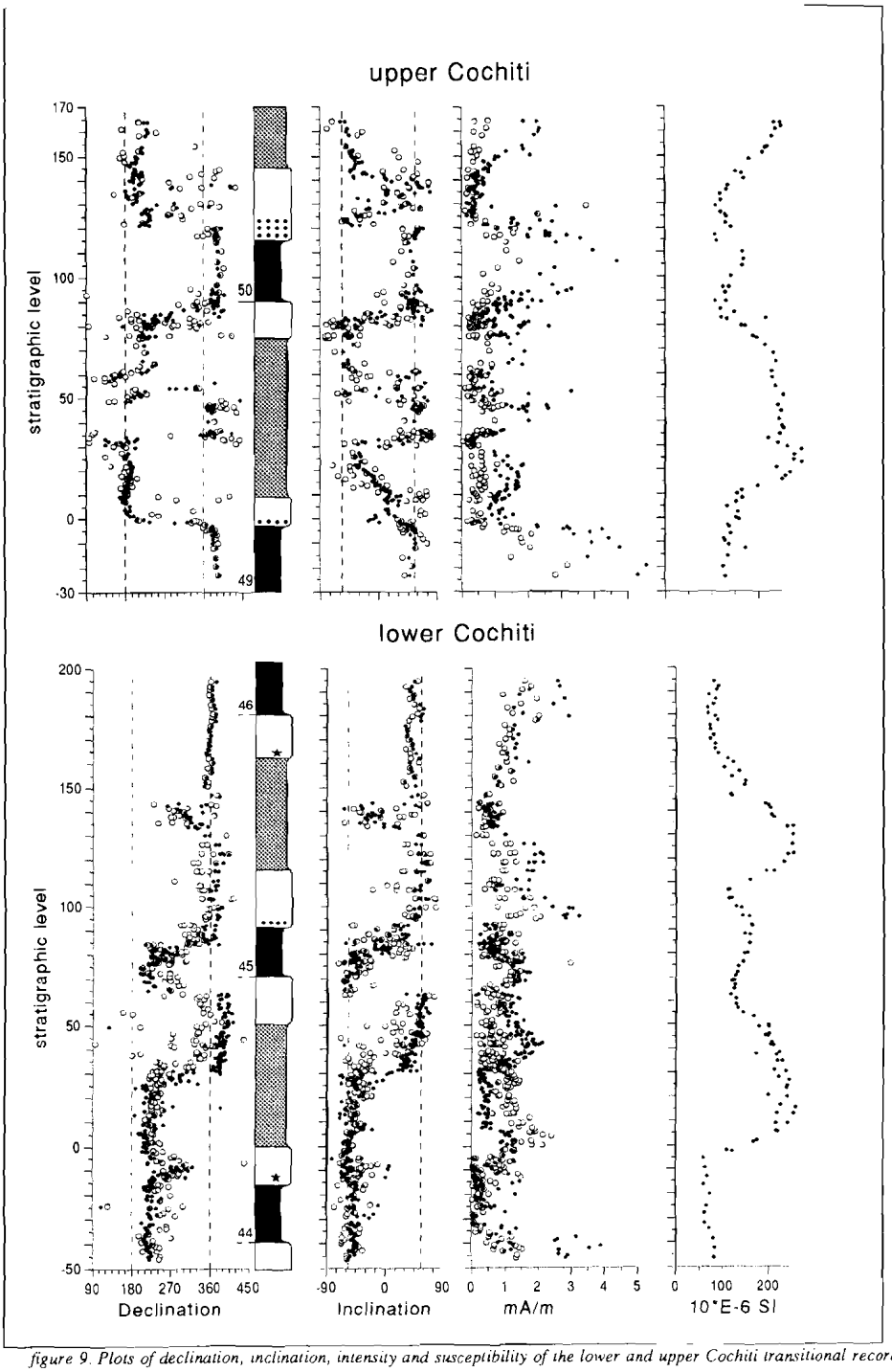

figure 9. Plots of declination, inclination, intensity and susceptibility of the lower and upper Cochiti transitional records of the Capo Bianco section. For the lithological column, see figure 3. The figures next to the column are the numbers of the cycles [9]. The 0-levels were defined at a distinct litholgical horizon in each record. In the plots of declination, inclination and intensity, the black dots (circles) represent the HT (LT) component.

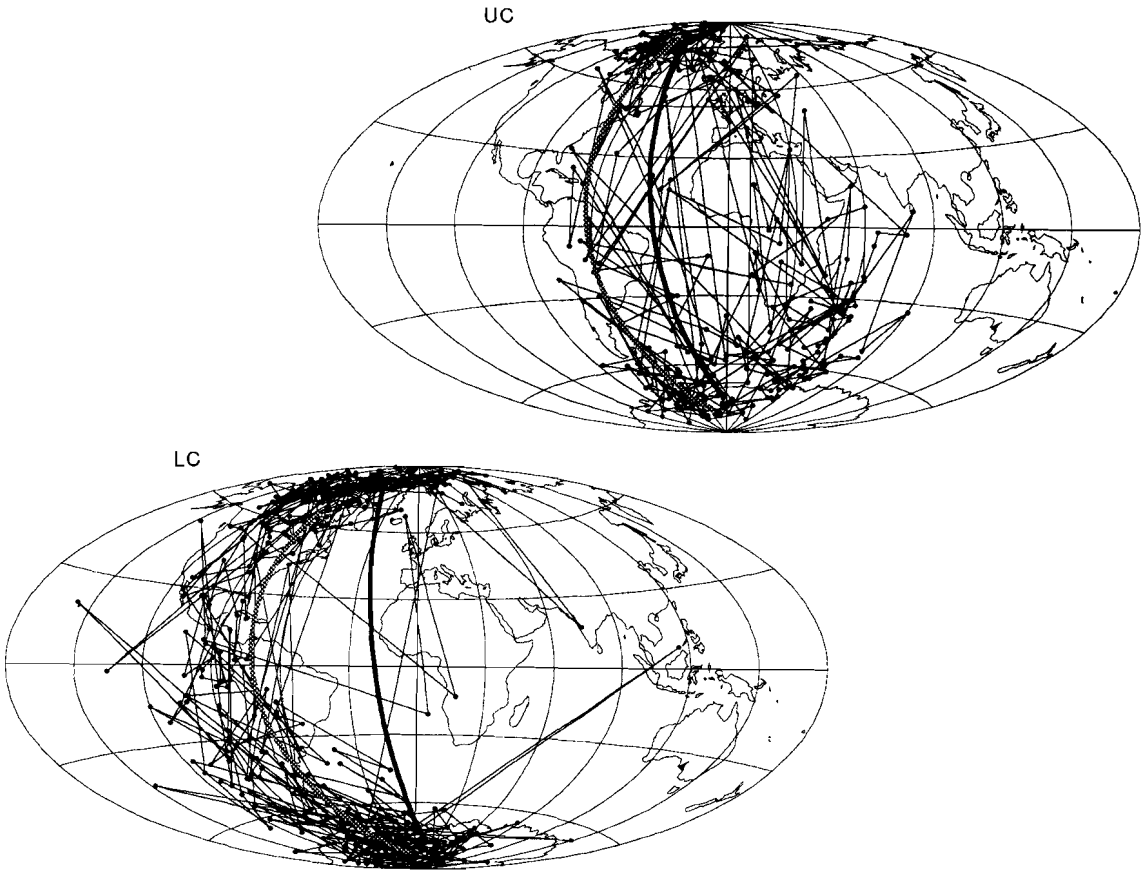

figure 10. VGP paths of the HT components of the lower and upper Cochiti transitions. The path of the lower Cochiti record is confined to North and South America; the path of the upper Cochiti record is confined within the great-circles over the east coast of North America and India. Black (Grey) lines: VGP paths of smoothed mean non-(near-)transitional directions.

	dec	inc	N	Rsum	gamma	gammac
after UC	186.5	-47.9	9	8.912	12.1	6
before UC	348.7	52	14	13.931		
after LC	332.6	48.4	9	8.922	19.6	5.5
before LC	183.2	49.2	15	14.93		
3.40-3.88	184.2	-39.4	49	48.28	6.3	5.2
Cochiti	-5.5	49.6	8	7.95		
3.97-4.10	179.7	-40	17	16.78	10.3	5.9

Table 1: Near transitional directions before and after the two transitions. N: number of data; Rsum: unit vector sums; gamma: angle between the two directions before and after the transition; gammac: crititical angle.

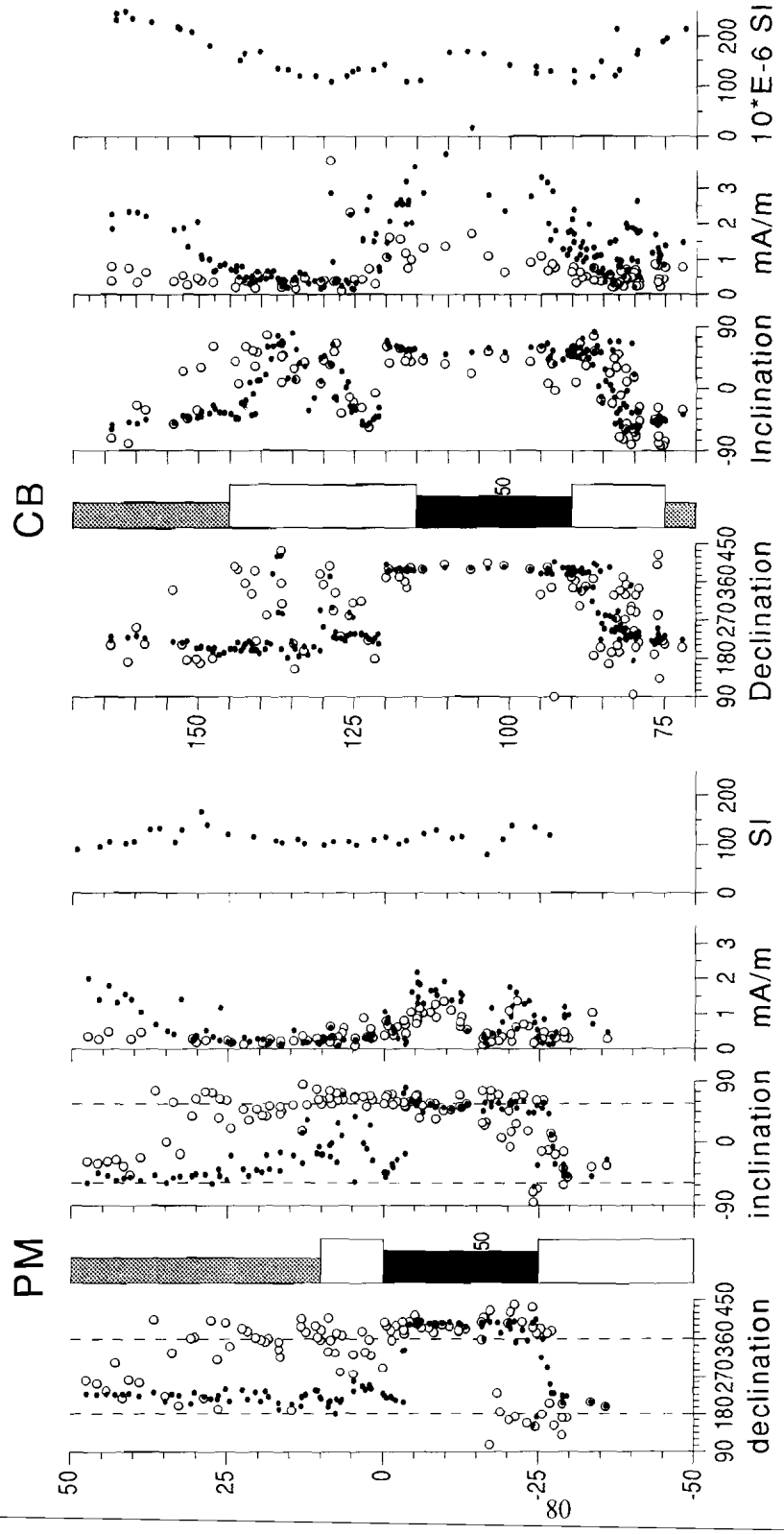

figure 11 The corresponding trajectories straddeling gray 50 of the upper Cochiti transition in the Punta di Maiata and the capo bianco sections. The normal directions in the upper Cochiti record of Capo Bianco (CB, this study) are recorded over a longer sedimentary interval than in the Punta di Maiata record (PM, [1]).

The sedimentary filtering and changing lock-in depths can be explained by an early diagenetic magnetite model, based on the paleomagnetic and geochemical properties of three subsequent cycles sampled in the Trubi sediments of the Punta di Maiata section [1]. Primary magnetites, which have acquired their remanences - probably a chemical remanence magnetisation (CRM) - shortly after deposition may be superposed or substituted by secondary magnetites - also carrying a CRM - which are formed after burial. If the formation of the secondary magnetites is shortly after a reversal of polarity and the primary magnetites are formed before the transition, the directions of the remanences of the two types of magnetite are more or less anti-parallel. The amount of newly formed secondary magnetites is dependent on the redox conditions in the sediment. Under anoxic circumstances, no secondary magnetites will be formed. The geochemical data indicate [1] that the grey layers only probably underwent anoxic conditions and should therefore only contain primary magnetite. As a consequence, the remanences in the grey layers should reflect the geomagnetic field shortly after deposition. In the other layers, the superposition of the primary and secondary magnetite remanences will result in post-transitional, intermediate or pre-transitional directions, depending on the intensities of the post-transitional and pre-transitional CRMs, and the time (duration) during which they were formed. Therefore, perhaps with the exception of the grey layers, below the chronostratigraphic level of the polarity transition, the observed remanences may carry all possible directions, pre-transitional, intermediate or post-transitional.

This model [1] only partially fits the current observations. According to the model, the chronostratigraphic level of the upper Cochiti transition is above the grey 50 layer, since grey records the geomagnetic field during deposition, and since it has pre-transitional directions. However, in the lower Cochiti record the data are not in agreement with the model: the grey layer of 45 shows intermediate and post-transitional directions. Above it, in beige 45, an excursion is observed. If the remanence in grey 45 represents the geomagnetic field during deposition, then the observed transition recorded in grey 45 represents the actual lower Cochiti geomagnetic reversal. Then the 'excursion' at the centre of the subsequent beige layer could not be explained by a superposition of pre-transitional magnetites carrying normal directions and post-transitional magnetites carrying reversed directions, since at this sedimentary level the primary as well as the secondary magnetites should have had post-transitional normal directions. We find it difficult to believe, however, that the excursion in beige 45 has a geomagnetic origin, the more since the directions above it still do not show the familiar 35° rotation. They rather show directions that are very similar to those of the G/G boundary, where there is evidence for (partial) CRM resetting [18]. New geochemical analyses are now being performed to refine the model and to explain the observations in the lower Cochiti record better.

Acknowledgements

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CHAPTER 8

THE GILBERT/GAUSS SEDIMENTARY GEOMAGNETIC REVERSAL RECORD FROM SOUTHERN SICILY

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Abstract. A detailed paleomagnetic record of the Gilbert/Gauss reversal boundary from southern Sicily shows that the remanence characterising the reversal is carried by a high-temperature magnetite component. A low-temperature component is intermediate between the high temperature component and a present-day overprint, suggesting that it is obscured by weathering. The actual reversal record is clearly affected by filtering caused by early diagenetic acquisition of a chemical remanent magnetisation (CRM). Intermediate VGPs show a general confinement to a longitude over the Americas. More specifically, they are restricted to two clusters, a less distinct one off SE South America, and a very prominent one near NE North America. The remarkable coincidence with two 'Hoffmann clusters' [Hoffmann, 1992] suggest that even a CRM acquisition filter cannot entirely obsure long-lived transitional features.

INTRODUCTION

Since Brunhes [1906] discovered the existence of polarity transitions recorded in rocks, the natural remanent magnetisation (NRM) from lavas and sediments has been used to study the geomagnetic polarity changes in the geological past. The recordings of transitions of the geomagnetic field in a lava sequence consist of spot readings, because each lava flow represents one discrete point in time, although considerable detail may occasionally be observed [Mankinen et al., 1985] or typical field features such as recurring clusters may be seen [Hoffman, 1991; 1992]. In a sedimentary record, the remanence is often assumed to represent a continuous registration of the paleomagnetic field and, provided the sedimentation rate is high enough, a detailed sampling of a polarity transition would yield an almost continuous record of the geomagnetic behaviour during transition. The NRM in sediments appears to be much more complicated, however, than a straightforward registration of the geomagnetic field [Hoffmann and Slade, 1986; Karlin, 1990; Rochette, 1990; van Hoof and Langereis, 1991; 1992; Langereis et al., 1992].

Evidently, this 'artefact behaviour' has consequences for 'true field behaviour' theories, such as the recently advocated relation between longitudinal confinement of the virtual geomagnetic pole (VGP) transitional paths over North and South America [Clement, 1991; Tric et al., 1991] and physical processes in the Earth's interior [Laj et al., 1991]. This true field behaviour theory is both supported - by recurring VGP clusters from volcanic data in the same longitudinal band [Hoffman, 1991;1992] - and not supported - by VGPs from lava's scattered all over the globe [see e.g. Valet et al., 1992]; it is critically examined in a statistical sense [Valet et al., 1992] and again defended [Laj et al., 1992]; it encounters considerable caution because of possible sedimentary artefacts [Langereis et al., 1992], while it is also argued that these cannot represent the general case [Weeks et al., 1992]. In other words, the debate is still alive and lively, although the observations and interpretations published sofar are not necessarily mutually exclusive. For instance, the observed longitudinal bias in the stable (non-transitional) field during the past 5 Myr [Constable, 1992] and the advocated filtering of the geomagnetic signal [Langereis et al., 1992] are not difficult to reconcile [Jackson, 1992]. Similarly, it is easily seen that recurring VGP clusters [Hoffmann, 1991] that may correlate to specific inclined dipolar field configurations and near-radial flux concentrations [Hoffmann, 1992] may lead to longitudinal VGP confinement if the NRM signal is filtered due to acquisition in a sedimentary environment. But we will come back to this filtering process in the discussion.

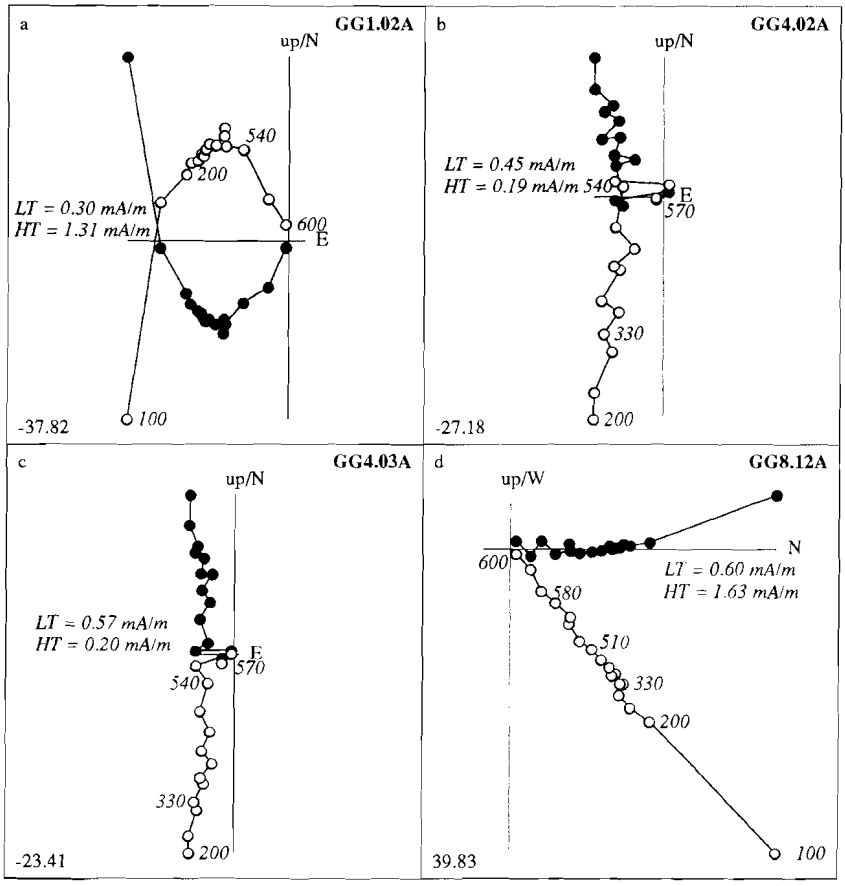

Figure 1. Thermal demagnetisation diagrams of the Gilbert/Gauss reversal. Stratigraphical level (down left) refers to the stratigraphic columns of figure 2; solid (open) symbols are horizontal (vertical) projections. Numbers denote temperature steps, between 580 and 600 °C small steps (5 °C) were used. Temperature steps below 100 °C (1a,d) or below 200 °C (1b,c) are not shown to enhance details at higher temperatures. Intensities of the low-temperature (LT) component and the high-temperature (HT) component are given. a) Clearly reversed HT component, including a 35° tectonic rotation; the LT component has a direction intermediate between the HT component and the present day field direction; b,c) Intermediate directions: westerly/shallow up (b) and westerly/shallow down (c); d) Normal HT component, showing no rotation.

To study the records of paleomagnetic polarity reversals in sediments, we have sampled 14 polarity transitions of the Gilbert and Gauss Chronozones in the Pliocene Trubi and (lower part of the) Narbone marls of Sicily and Calabria, several of which have already been published [Linssen, 1992; van Hoof et al., 1992; and references therein]. The studies of these rocks have revealed that in general the reversals seem to be recorded by two magnetite components, probably with only slightly different grain size distributions and/or composition, but with different blocking temperature spectra. One is the low-temperature (LT) component removed between 200/250° and 480°/510°C, the other is the high-temperature (HT) component removed between 480°/510°C and 600°C. Recognition of characteristic stable (non-transitional) directions is facilitated by a typical clockwise (35°) rotation of the Sicilian marls [Langereis and Hilgen, 1992]. This enables us to distinguish true normal HT components from a present-day direction overprint.

From the sequence of 14 transitions, the record of the Gilbert/Gauss transition is presented in this paper. This transition - identified by an earlier magnetostratigraphic study [Langereis and Hilgen, 1991] - was sampled at Capo Bianco (37° N, 13°30' E), near Eraclea Minoa in southern Sicily; the bedding of the section has a strike and dip of 227°/10°NW. The lithology - marine marls of the Pliocene Trubi formation - form a repetition of grey, white, beige and white coloured beds, so called quadruplets that range from number 1 at the Mio-Pliocene boundary up to number 119 at the

Gauss/Matuyama boundary. The lithology has been extensively discussed earlier [Hilgen, 1991; and references therein]; the sedimentation rate is approximately 5 cm/kyr. In some quadruplets, the grey layer cannot clearly be distinguished from the surrounding white layers; this was also the case with quadruplet 78 in which the Gilbert/Gauss transition is located. It concerns an interval with a maximum in carbonate content [cf. Hilgen, 1991] and although before sampling the weathered part of the marls is always removed to obtain the fresh, blue coloured marl, in such high-carbonate layers it is not possible to reach the usual freshness. Sampling was done by taking oriented cores at stratigraphic intervals of 1 cm. Using the drilling orientation, width of the saw cut and tilt of the bedding plane, the stratigraphical level of each specimen was calculated, yielding a resolution much better than 1 cm.

THE REVERSAL RECORD

Typical thermal demagnetisation diagrams are shown in figure 1. In the Trubi marls, the present-day field component is usually removed at 200/250 °C, but in the Gilbert/Gauss record a substantial part seems to persist at higher temperatures, probably as a result of the increased weathering due to a higher carbonate content. The direction of the NRM removed between 200/250 °C and 480/510° C is therefore intermediate between the secondary and high-temperature (HT) component (fig. 1a), very similar to the results of the high-carbonate upper and lower Thvera reversal records [van Hoof and Langereis, 1992]. Even though this low-temperature (LT) component is not obvious in the demagnetisation diagrams of the marls with a high carbonate content, we still have determined the directions of the NRM demagnetised over this temperature trajectory for comparison with existing

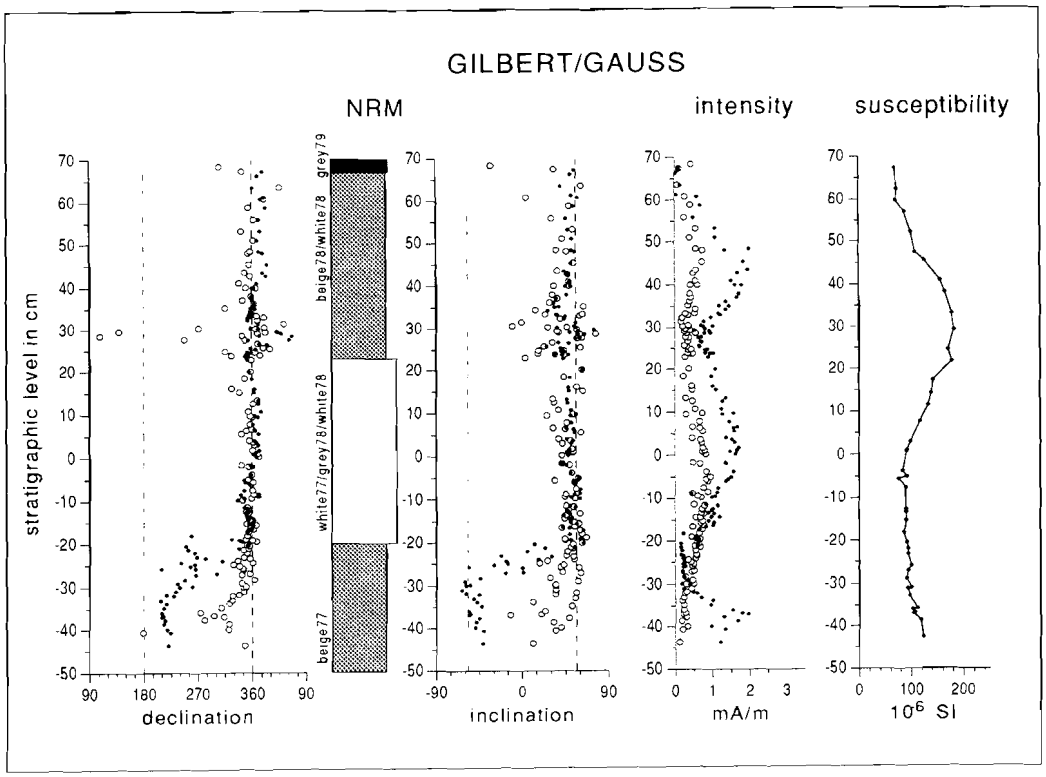

Figure 2. Records of the declination, inclination, intensity (after thermal demagnetisation), and initial susceptibility. The lithological expression of quadruplets 77 and 78 is not clearly visible; white77/grey78/white78 is exposed as one single high-carbonate layer, while beige78/white78 is exposed as a beige layer. Circles (dots) denote the LT (HT) component; dashed lines are declination/inclination (57.5°) of the geocentric axial dipole. A transition of the HT component from reversed to normal occurs between level -20 and -30 cm, but the actual transition may occur higher in the sediment due to CRM resetting (see discussion in text).

records, but we realise that we may be looking - at least in some intervals - at an overlap in unblocking temperature spectra only. The intensity of this LT component is relatively small and is usually seen as a cluster in the demagnetisation diagrams. The HT component is removed at temperatures higher than 480 °C, but the most rapid decay is observed only at temperatures higher than 510 °C. Often, it is only at these highest temperatures that this component shows a linear decrease towards the origin.

The registration of the Gilbert/Gauss boundary in the HT component (fig. 2) shows a decrease in intensities between levels -35 cm and -15 cm, i.e. typically at the beige to white boundary [see van Hoof et al., 1992]. At this level, declinations become westerly and the inclinations appear to steepen. Between levels -30 and -20 cm the intensities are at a minimum and despite the scatter it is clear that the directions change from reversed to normal via westerly declinations and shallow inclinations. Declinations do not regain, however, their typical 35° rotation values, but remain - after tilt correction - north directed. Subsequently, intensities recover their pre-transitional values at

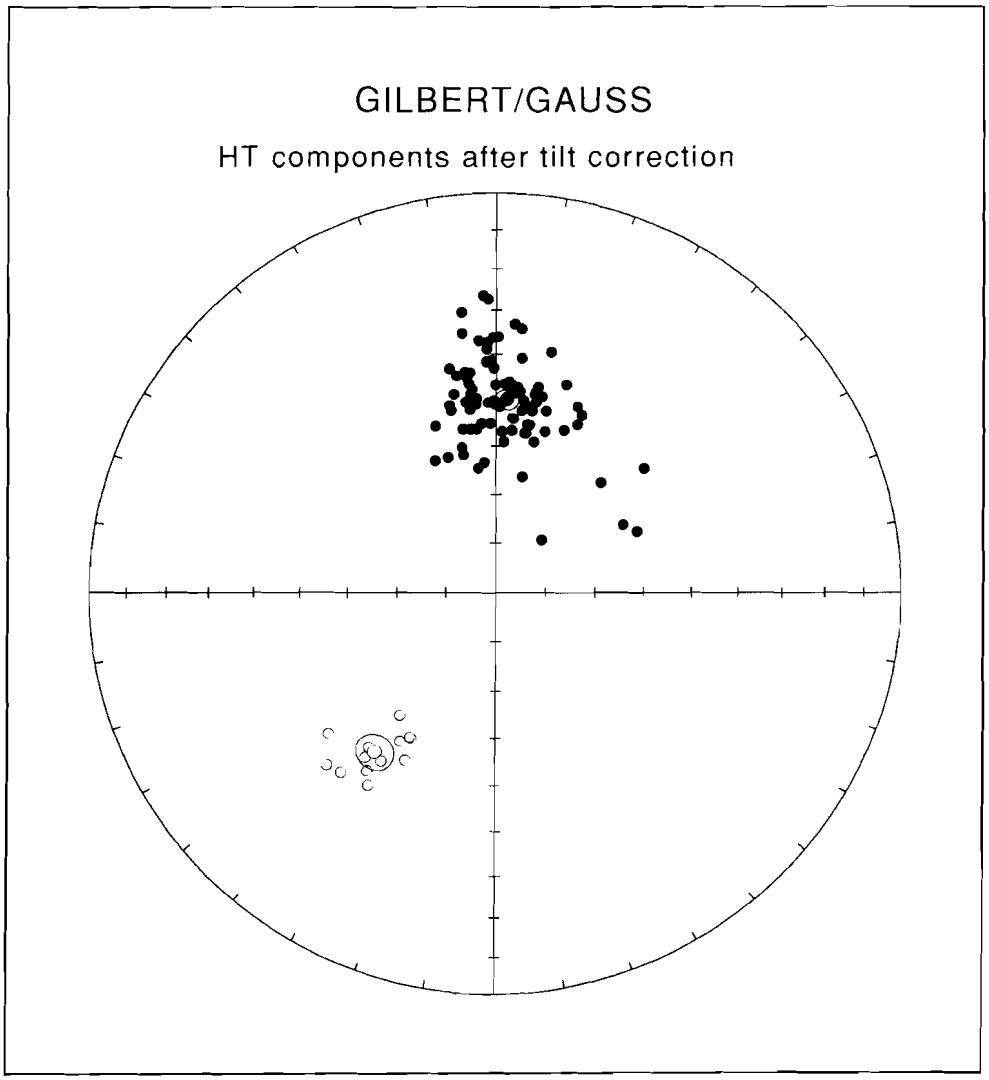

Figure 3. HT directions (after tilt correction) just before and after transition between -20 and -30 cm (fig. 2). The reversed mean directions shows a clear tectonic offset (Table 1), while the mean normal direction does not show this rotation.

level 0 cm, but show again a significant minimum between 25 and 30 cm. In general, declinations still do not reach their expected value (35°), but at this intensity minimum there is a deviating HT cluster having easterly declinations of 40-75° and steep inclinations. Only when the intensities regain their initial values, above 30-35 cm, declinations are approximately 15-30°, indicating (close to) stable normal directions. In the topmost part of the section, the very low intensities are the main cause of the scattered (LT and HT) components. These are regarded as less reliable since they can only be determined with great uncertainty; the better diagrams seem to show the same clockwise rotation.

Results from earlier transitions [van Hoof and Langereis, 1992] have revealed that the susceptibility record (fig. 2) depends strongly on lithology, having maxima in the beige layers. Hence, the susceptibility is not a a reliable tool to normalize the intensities of the NRM, although at the intensity minimum at 25-30 cm the susceptibility is highest and would tend to decrease 'relative intensities' even further. We rather interpret the higher susceptibilities, however, to be caused by a higher paramagnetic contribution of the clay minerals, and thus the maximum probably denotes the 'middle' (= highest clay content) of the beige-white layer of cycle 78 (fig. 2) where otherwise there is no clear visible distinction between beige and white.

clustering, may be due to true geomagnetic field behaviour, but it may also be caused, or at least be 'contaminated', by two other phenomena. One concerns a secondary (present-day) overprint, and the other concerns the filtering mechanism due to early diagenetic CRM acquisition.

A present-day overprint, especially in the high-carbonate quadruplet of the Gilbert/Gauss transition, may have considerable influence because the overlap in the blocking temperature spectra of the secondary component and the HT component may be persistent up to the highest temperatures [van Velzen and Zijderveld, 1992]. In this particular case, it would introduce an offset in HT (and LT) directions, since the reversed (normal) HT component is expected to have a southwest (northeast) declination due to the 35° rotation of the basin, and a relatively shallow inclination due to an inclination error. Hence, a persisting secondary component will offset the reversed (normal) declination to west (north) and decrease (increase) the inclination, causing an apparent non-antipodality, as is indeed seen in figure 3 and Table 1 with respect to the declination (but not with respect to the inclination).

In addition, a decrease in intensity of the geomagnetic field during the transition may cause a low-intensity reversed direction (southwest declination, negative inclination) to be increasingly overprinted by the normal (north-down) secondary component, so that the resulting HT directions are

	Dec	Inc	N	Rsum	ctmd
Non before	219.2	-39.4	49	48.28	} no
Non after	32.8	40.1	19	18.92	
Near before	216.5	-48.7	13	12.90	} no
Near after	3.2	50.1	92	89.86	

Table 1. Mean directions - after tilt correction - of stable non-transitional (based on mean directions of entire polarity zones) and near-transitional (based on mean directions from intervals below -35 cm and above 17 cm) HT components, respectively. Common true mean directions (ctmd) determined according to McFadden and McElhinney [1990]. Near-transitional means before tilt correction are 228.1/-49.5 and 13.7/56.4

Discussion

The VGP path was calculated after a correction for the 35° rotation of the location. The path is largely confined to a well-known longitude, that over the Americas (fig. 4). Characteristically, VGPs seem to be concentrated in two clusters, a less distinct one at the southern tip of South America, and a very prominent one at the northeast coast of North America. These clusters coincide remarkably well with two of the 'Hoffmann clusters' [Hoffmann, 1992; 1992]. This confinement, or

intermediate, having westerly declinations and shallow inclinations. The persistent overprint will then largely obscure any real geomagnetic features during the transition. We note that before tilt correction, however, the north directions between -20 cm and 40 cm are significantly different from the present-day field direction (fig. 3), and hence they cannot represent a recent overprint (only).

The other explanation of the observed VGP path is a filtering mechanism due to early diagenetic CRM acquisition. Here, we emphasize that this filtering mechanism is unlike a simple post-depositional remanent magnetisation (PDRM)

filter which results in a moving window that must be very large with respect to reversal duration to obscure long-term geomagnetic field features [Rochette, 1990; Weeks et al., 1992]. But in this case we have sediments that were deposited during cyclically changing paleoredox conditions, resulting in an alternation of dysoxic/anoxic layers (grey, and middle of beige) containing mainly primary magnetites, and suboxic/oxic layers (white-beige, and beige-white) in which both primary and secondary magnetites are present [van Hoof et al., 1992; van Hoof, 1993]. Hence, the direction of the recorded NRM is either the - possibly PDRM smoothed - primary magnetite direction (grey, middle of beige), or the vector sum of the directions of the two types of magnetite; the two types may occasionally be distinguished by slightly different blocking temperature spectra (e.g. LT & HT components: van Hoof and Langereis [1991]). Or, the primary magnetite may even be entirely reset by the secondary magnetite. The latter occurs especially in white and in the transition from white-to-beige and beige-to-white. This resetting occurs during (sub)oxic conditions by downward migration of the redox front from the sediment-water interface. The resulting formation of secondary magnetite (but only/mainly in white-beige and beige-white) may occur to a considerable depth and can result in apparent excursions before the actual transition [van Hoof and Langereis, 1991; 1992].

Using this geochemically constrained model, we may envisage the following, tentative scenario for the Gilber/Gauss reversal record (cf. figs. 2,4). True reversed directions are recorded in beige 77 (-45 to -33 cm), and have relatively high intensities. Intermediate directions with very low intensities are found in the top of beige 77 (-33 to -20 cm) and they give mainly VGPs near SE South America. Although this may be due to a transitional (low-intensity ?) inclined dipolar state [Hoffmann, 1992], it is equally well explained by the vector sum of a reversed component residing in primary magnetite and a normal one residing in secondary magnetite. Remarkably, and contrary to earlier observations, the LT component reverses lower in the sediment than the HT component. But here it concerns also the only interval where LT intensities are higher than HT intensities, and we speculate that the influence of the secondary overprint on the LT component, as discussed above, is the main cause for this 'anomalous behaviour'. The typical north declinations of the white lithology (white 77 / 'grey' 78 / white 78; -20 cm to 25 cm) are probably due to resetting by secondary magnetite. VGPs, when corrected for the 35° rotation, are strongly and consistently concentrated at NE North America.

Just below 30 cm, we find the small but clearly deviating interval with easterly/steep HT components and reversed to intermediate LT components, and low intensities. Since this interval

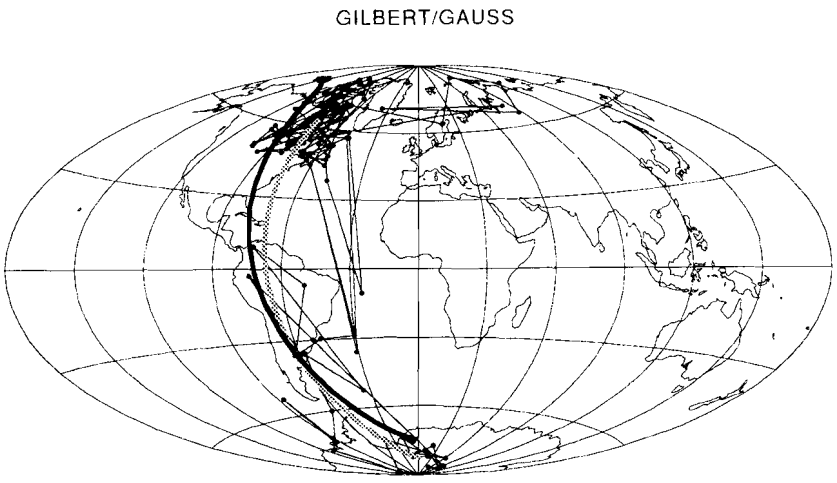

Figure 4. VGP path of the Gilbert/Gauss transition. The solid black line shows the VGP path obtained by filtering the mean directions of under/overlying polarity zones; the shaded line represents the filtering of near-transitional directions determined from the reversal record (fig. 3, Table 1). The clusters - a less distinct one off South America, and a prominent one in North America - coincide remarkably well with those observed by Hoffmann [1992].

occurs at a susceptibility maximum (indicative of a higher clay content, i.e. the 'middle' of a regular beige layer), these directions may represent primary magnetite, and hence may give either 'true field' (intermediate) directions, or they represent the vector sum of primary and (some) secondary magnetite. Directions between 30 cm and 40 cm, still in the clayey (high susceptibility) part of the beige layer, again give consistent VGPs located at NE North America. Above 40 cm, directions are close to the usual stable directions, but since this interval represents the beige-white (cycle 78) transition, it is a good candidate for CRM resetting, i.e. it may record stable, post-transitional directions from, for instance, cycle 79.

This tentative scenario is at least consistent with the observations of other reversals from the Trubi marls. Also, we realise that the correspondence of our clusters (in this, but also in some other records: see e.g. Langereis et al. [1992]) with those of Hoffmann is striking. But some questions still cannot be answered satisfactorily, such as: what are the time constraints of the secondary magnetite resetting, does it occur gradually over a long period of time, depending on redox conditions, or does the (partial, or complete) resetting occur over a short time interval, but to a considerable depth ?

If we assume, for instance, that the actual geomagnetic transition has been completed at approximately level 40 cm, then the 'true' directions between 30 cm and 40 cm (giving the NE North America VGP cluster) may have caused the interval between -20 and 27 cm to be reset by CRM acquisition. This would require a very rapid downward migration of the redox front ('oxygen burn-down'; cf. van Hoof et al. [1992]), a process that is stopped at the middle of beige at -30 cm, while between -30 cm and -20 cm - the transition from beige to white - a partial CRM resetting causes the observed intermediate directions and low intensities. This would imply that the VGP cluster has a duration of at least 2 kyr (~10 cm of sediment), but possibly longer, if the beige-white part of 78 (40 cm to 60 cm) has been reset, in turn, by post-transitional stable directions.

Without firm temporal constraints on this paleoredox and hence CRM resetting mechanism the above assumption must, for the time being, remain speculative. It is clear, however, that the detailed (paleomagnetic/rock magnetic/geochemical) study of these cyclic sediments, and the scrupulous examination of all 14 reversal records in relation to the lithology in which they are recorded, may give important information. Not only on the details of the NRM acquisition process in these marls, that may be important for other (perhaps not *visibly* cyclic) sediments as well. But also to provide time constraints, once the redox mechanism is better understood, on important geomagnetic field features such as inclined dipolar states.

Whatever the disadvantage of these cyclic sediments for true transitional field recording may be, it may then turn into an advantage: the fact that these marls form the basis of a high-resolution astronomical polarity time scale [Hilgen, 1991] provides a potential for accurate estimates of time and duration of the processes involved, an estimate that is difficult to obtain from homogeneous sediments.

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Chapter 9

The Upper Kaena Sedimentary Geomagnetic Reversal Record From Southern Sicily

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A detailed paleomagnetic and rock magnetic record of the upper Kaena polarity reversal was determined from late Pliocene marine marls in southern Sicily. Two separate characteristic magnetization components were distinguished: a low-temperature (LT) component between 200° and 500°C and a high-temperature (HT) component between 500° and 580°-610°C. The two components record the reversal in a slightly different way, but both were acquired in an early stage following deposition, most probably by authigenic (biogenic) formation of magnetic minerals. The actual (R-N) reversal, via E and horizontal inclinations, seems to take 3.3 kyr and is followed by a stable normal interval (3.3 kyr) and subsequently by a significant excursion (50°-60°) in declination (3.6 kyr). Relative intensities were lower during a time-span of 5.6 kyr and show a minimum of 10% of the nontransitional values, but during the excursion, no intensity changes are apparent. The VGPs of the HT component are confined to a longitude between India and Australia; those of the LT component oscillate between this same longitude and one passing over North and South America. We conclude that most if not all observed features are largely determined by the remanence acquisition process which is complicated by authigenic formation of magnetic minerals and cyclically fluctuating paleoredox conditions. This has not only resulted in considerable smoothing but also in delayed NRM acquisition in specific sedimentary intervals. We believe that the observed "excursion" is a sedimentary artifact and not related to geomagnetic field behavior during transitions. We emphasize that many reported "geomagnetic" features may in fact be due to the sedimentary remanence acquisition process.

INTRODUCTION

Detailed records of geomagnetic reversals may be obtained from different sources, most important of which are those recorded in lava sequences and those in sedimentary sequences. Both recording media have their advantages and drawbacks. The main advantages of sedimentary sequences concern appropriate time control and continuous registration of the geomagnetic signal. In addition, long sedimentary sequences from the same area may yield useful information concerning the characteristics of a number of successive polarity reversals [*Valet et al.*, 1988]. However, a major drawback of sedimentary records is the insufficient understanding of the acquisition of the remanence which may involve complex processes of authigenic formation of different magnetic minerals at different depths in the sediment during accumulation [e.g., *Lund and Karlin*, 1990; *Karlin*, 1990*a,b*; *Stolz et al.*, 1990; *van Hoof and Langereis*, 1991].

Long-term magnetostratigraphic research in the Central Mediterranean by our laboratory has so far resulted in the identification and location of 25 successive reversals from 7.5 to 1.5 Ma in the upper Miocene of Crete [*Langereis*, 1984] and in the Pliocene of Sicily and Calabria [*Zijderveld et al.*, 1986; *Zachariasse et al.*, 1989, 1990; *Langereis and Hilgen*, 1991]. A succession of three older reversals (11-12 Ma) from the same area is known from Zakinthos [*Laj et al.*, 1988]. Not all reversals are suitable for detailed paleomagnetic research, for exemple, due to low intensities and/or unfavorable rock magnetic properties. From the nine upper Miocene reversals, five were extensively investigated earlier [see *Valet et al.*, 1988, and references therein]. From the lower Pliocene in Calabria, five successive reversal records (upper Thvera, upper and lower Sidufjall and upper and lower Nunivak in the Gilbert Chron) have been reported by *Linssen* [1988, 1991]. From the upper Pliocene in Sicily, three more successive reversal records (upper and lower Mammoth and lower Kaena in the Gauss Chron) were also studied in detail by *Linssen* [1991]. This paper presents the details of the fourth successive reversal, the upper Kaena reversal, from the same sedimentary sequence.

GEOLOGICAL SETTING AND SAMPLING

The Mammoth and Kaena reversals in the Gauss Chron were unambiguously identified by a detailed magnetostratigraphic study of the Punta Piccola section in the Caltanisetta basin of southern Sicily (Figure 1) [*Zachariasse et al.*, 1989, 1990]. The Punta Piccola section has a bedding plane with strike and dip of 315° and 11° NE and consists of marine marls of the Pliocene Trubi formation in which the Mammoth subchronozone is located and of the very similar Pliocene Monte Narbone formation in which the Kaena subchronozone is located (Figure 2). The marls of the Trubi and Monte Narbone formation consist mainly of carbonates (50-80% $CaCO_3$) and a mixture of clay minerals [*Hilgen*, 1987; *de Visser et al.*, 1989; *Linssen*, 1991]. The average sedimentation rate per polarity zone in the Punta Piccola section can accurately be determined and is remarkedly constant: 4-5 cm/kyr; the sedimentation rate in the Kaena polarity subzone is 4.3 cm/kyr [*Zachariasse et al.*, 1989].

Hilgen [1987] recognized a long succession of small-scale sedimentary cycles, so-called quadruplets. These quadruplets form a repetition of grey, white, beige, and white coloured beds (Figure 2) and were deposited during cyclic sedimentation periods which are related to the *Milankovitch* [1930] cycle of precession of the Earth's orbit [*Hilgen and Langereis*, 1989] with

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an average duration of 21.7 kyr [*Berger and Loutre*, 1988]. The low-carbonate grey marls probably represent a period of increased continental runoff and a mainly suboxic environment [*de Visser et al.*, 1989]. The high-carbonate white marls represent a more oxic environment, but only those from the southwestern part [sicily, southern Calabria] of the Trubi formation contain the intercalated low-carbonate beige interval [*Hilgen*, 1987]. This beige interval is probably due to increased wind-blown African input [*de Visser et al.*, 1989]. In some of the grey marls from the Monte Narbone formation a dark brown and sometimes finely laminated sapropelitic layer is intercalated [*Hilgen*, 1987], as is the case in the sampled upper Kaena record [Figure 2). Although the weathering profile shows quite sharp changes in color and induration, the changes in fresh, unweathered sediment are much more gradual: light blue (white layer] to dark blue-grey (grey layer). The color boundaries between grey marls and sapropel are mottled and gradational, from dark grey to dark green to dark brown.

Sampling was done in two successive field trips. During the first trip, we sampled a stratigraphic interval of 80 cm in which the reversal had been located earlier [*Zachariasse et al.*, 1989] and which consisted of a lower white marl, a grey marl including a sapropelitic layer, and an upper white marl. During the second trip, an interval of 25 cm where we had pinpointed the actual reversal was resampled to obtain more detail. Only a short interval of less than 5 cm within the sapropel could not be sampled due to fragility (lamination) of the sediment. We have sampled by taking oriented cores of 25 mm diameter, more or less parallel to the bedding plane at very close intervals (<1 cm) and from a freshly cut vertical plane. In order to get a higher stratigraphic resolution a number of specimens were split into upper and lower halves. The stratigraphic position of each specimen is accurately determined by taking into account drilling orientation, bedding plane and width of the saw cut. The stratigraphic levels in the present paper are in centimeters with respect to the bottom of the sapropel (Figure 2). The sampled upper Kaena interval therefore ranges from 25.07 to 25.87 m in Figure 2.

ROCK MAGNETISM

The upper Kaena reversal occurs in a layered lithology. In particular, this reversal is pinpointed close to a sapropelitic layer which has a lithology distinctly different from that of the surrounding marls. It is therefore important to ensure that the carriers of the characteristic remanent magnetization are of primary origin even though they may not be exactly the same in the various lithologies. Rock magnetic parameters give an indication of the content and nature of the magnetic minerals, and for this reason we have determined the initial susceptibility (X_0) and several (remanent) hysteresis parameters: the (remanent) saturation magnetization (J_{rs}, J_s) and (remanent) coercive force (H_{cr}, H_c).

Susceptibility and Hysteresis Parameters

Acquisition of isothermal remanent magnetization (IRM) of a number of samples from different lithologies up to a maximum DC field of 2 T (Figures 3*a* and 3*b*) reveals that samples outside the sapropel in general show a saturation IRM (J_{rs}) in fields of 125-200 mT. Samples from the sapropel itself are saturated at higher fields (0.5 T) and, in addition, have higher J_{rs} values (Figures 3 and 4). The sapropel samples appear to have both a low- and a high-coercivity IRM acquisition curve. The low-coercivity component has a saturation field similar to that of the nonsapropel samples. The relative intensity of this component may be estimated at approximately 20-30% of the intensity at 0.5 T if it is assumed that the J_{rs} acquisition of the high-coercivity component is linear in low fields [*Dekkers*, 1988] (Figure 3). This results in an absolute J_{rs} intensity of the low-coercivity component which is of the same order as the J_{rs} of the samples with only one component. Thermal demagnetization of the J_{rs} shows that approximately 95% of the total saturation remanence

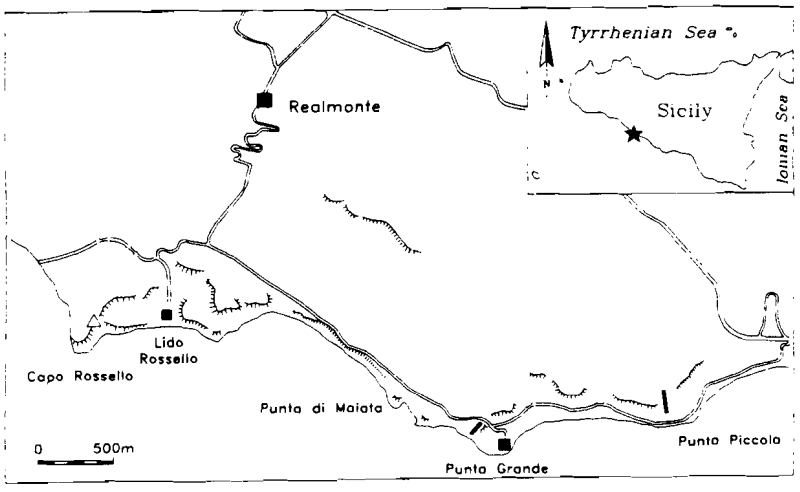


Fig. 1. Location of the Punta Piccola section in Sicily (Italy) in which the upper Kaena interval was sampled.

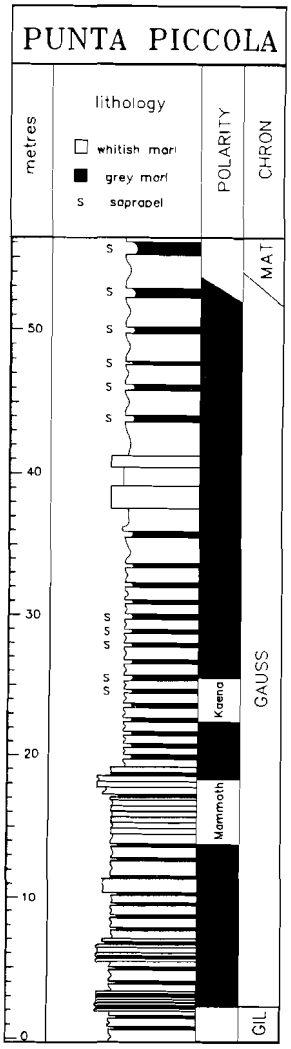


Fig. 2. Polarity sequence and lithology of the Trubi and Monte Narbone formations in the Punta Piccola section [after *Hilgen*, 1987; *Zachariasse et al.*, 1990] The 19-m level corresponds to the Trubi-Narbone transition. The upper Kaena reversal (indicated by an arrow) was sampled in detail over an interval of 80 cm, from 25.07 to 25.87 m. In the lithological column, the white layers are internally composed of a white, beige, and a white marl, the black intervals corresponds to a grey marl. The white marls have a higher carbonate content, while the grey and beige marls are relatively poor in carbonate [*Hilgen and Langereis*, 1989], probably due to increased continental runoff (grey) or increased African wind-blown input (beige) [*de Visser et al.*, 1989].

unblocks up to 580°C (Figure 3*c*). Together with the relatively low saturation fields (125-200 mT), this points to a magnetite dominated magnetic mineralogy. The remaining 5% shows maximum blocking temperatures of 580°-600°C, slightly higher than the maximum blocking temperature for pure magnetite. This suggests the presence of some cation-deficient magnetite [*Heider and Dunlop*, 1987; *Linssen*, 1988, 1991].

J_{rs} shows a clear lithological dependence: in the white layers, both above and below the grey layer, values are constant (3.5 A/m), while in the grey layer there is an increase (to 8 A/m) culminating in a maximum (of 15 A/m) in the middle of the sapropel. The lithological dependence of the initial susceptibility X_0 (Figure 4) is similar to that of J_{rs}: values are constant (250 x 10^{-6} S.I.) in the white layers and show a maximum (500-600 x 10^{-6} S.I.) in the sapropel. Neither X_0 nor J_{rs} is a useful magnetomineralogical indicator on its own because each is dependent on the nature and concentration of the magnetic minerals. The ratio J_{rs}/X_0 is more useful: it is largely independent of concentration, and provided that the dominant magnetic mineral is magnetite, it may give an indication of the grain size. The ratio shows typical values of 10-15 kA/m (Figure 4), lower than expected for fine-grained magnetite (>20 kA/m) [Dunlop, 1986]. In the grey layer and especially in the sapropel, values are significantly higher (15-25 kA/m).

However, the clay fraction has a relatively large influence on the bulk susceptibility X_0 because of its paramagnetic contribution. Hysteresis loop experiments show that this contribution is approximately 110 x 10^{-6} SI in the white layers, and as much as 250 x 10^{-6} SI in the sapropel (see Figure 5). Correction of X_0 for the susceptibility of the clay fraction does not significantly change the observed trend in X_0, but it causes the values for J_{rs}/X_0 to increase to 20-25 kA/m in the white layers and to as much as 50 kA/m in the sapropel. This clearly points to fine-grained magnetite as the dominant magnetic mineral in the white layers, whereas the high values in the sapropel point either to even finer-grained magnetite or to an additional magnetic mineral.

The remanent coercive force H_{cr} is largely independent of the concentration of magnetic material, and it is not influenced by paramagnetic clay minerals. Typical values for fine-grained magnetite are 40-60 mT [*Day et al.*, 1977; *Hartstra*, 1982; *Dunlop*, 1986], which are the values observed in the white layers (40-50 mT). In the grey layer/sapropel, H_{cr} values increase to 75 or even 100 mT which again indicates either even finer-grained (SD) magnetite or an addition of a high coercivity mineral (e.g., hematite, goethite). Although pyrrhotite also has a higher H_{cr} than magnetite for the expected grain size range, the observed maximum values of 90-100 mT would imply that the magnetomineralogy in the sapropel would be entirely dominated by this mineral [*Dekkers*, 1988], which is not the case.

The ratio $H_{cr}/(J_{rs}/X_0)$ [*Thompson and Oldfield*, 1986] is diagnostic for many magnetic minerals and may give useful information on the carriers of the remanence. The observed ratio $H_{cr}/(J_{rs}/X_0)$ is somewhat high (2-3) for very fine grained magnetites (Figure 4). Correction for the paramagnetic susceptibility of the clay fraction yields values of 1.2 in the white layers, compatible with fine-grained magnetite. Corrected values for the sapropel are only slightly higher, approximately 1.5, and in combination with the observed higher H_{cr} values (up to 75-100 mT) possibly indicate a trace amount of goethite [*Dekkers*, 1988]. Addition of goethite causes a relatively strong increase of $H_{cr}/(J_{rs}/X_0)$, whereas addition of pyrrhotite would decrease this

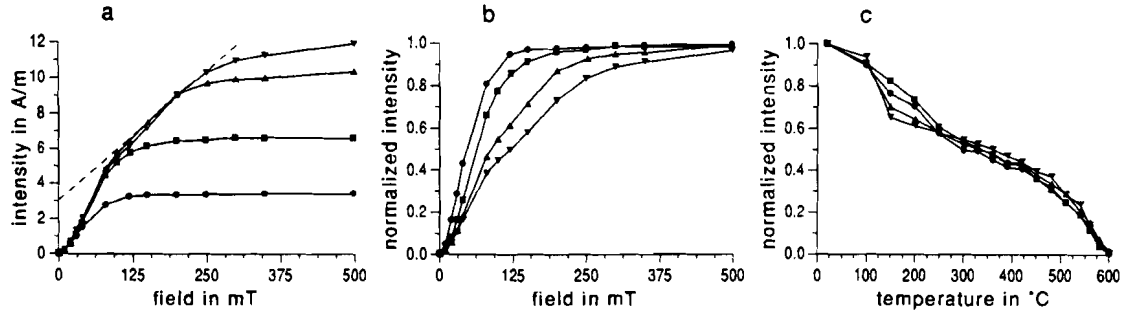


Fig. 3. (*a*) Absolute and (*b*) normalized IRM acquisition curves of samples from different lithologies, up to a maximum field of 1600 kA/m (2 T). Samples from the white marl (circles, level -22.8 cm) and from the grey marl (squares, level -5.1 cm) are saturated at 100-150 kA/m. Samples from the sapropel (triangles and inverted triangles, levels 1 and 2 cm, respectively) are only saturated at 400 kA/m. The curves show a low- and high coercive component. The low-coercive component is saturated ~100 kA/m; linear extrapolation of the high-coercive component back to the ordinate (dashed line) shows that the low-coercive contribution of the toatl IRM is 20-25%. (*c*) Subsequent thermal demagnetization of IRM shows similar decay curves except between 100 and 250 °C. Maximum unblocking temperatures are close to 580 °C, but a remaining 5% unblocks at slightly higher temperatures, possibly indicating some cation-deficient magnetite.

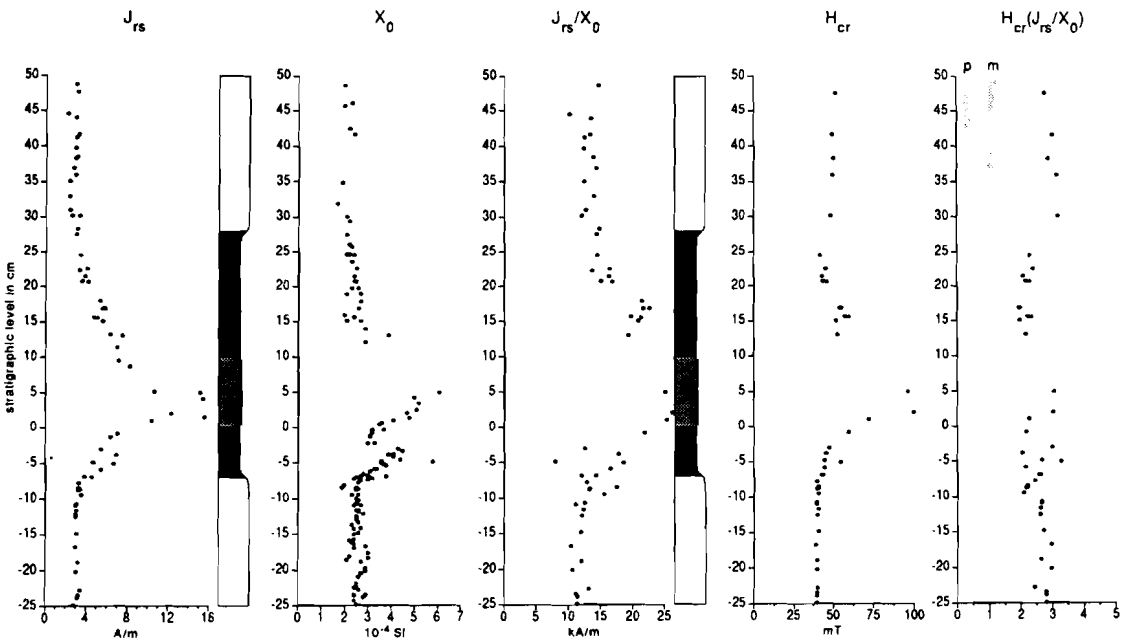


Fig. 4. Saturation IRM (J_{rs}), initial susceptibility (X_0), remanent coercive force (H_{cr}), and some interparametric ratios versus stratigraphic level and lithology: white marls (white), and a grey marl (black) including a sapropelitic layer (shaded). J_{rs} and X_0 show constant values in the white marls but reach a maximum in the grey/sapropelitic layer. Their ratio J_{rs} / X_0 is not constant either, indicating an addition of finer-grained magnetite and/or a (high-coercive) mineral. A similar trend is found for H_{cr}; the high maximum (100 mT) probably indicates addition of high-coercive material. This is supported by the $H_{cr}/(J_{rs}/X_0)$ ratio which shows a rather constant value (ranging 2-3) throughout the entire lithology. The ratio is dimensionless, since H_{cr} has been converted from mT to kA/m. The shaded lines denote typical values for fine-grained magnetites (m) and pyrrhotites (p).

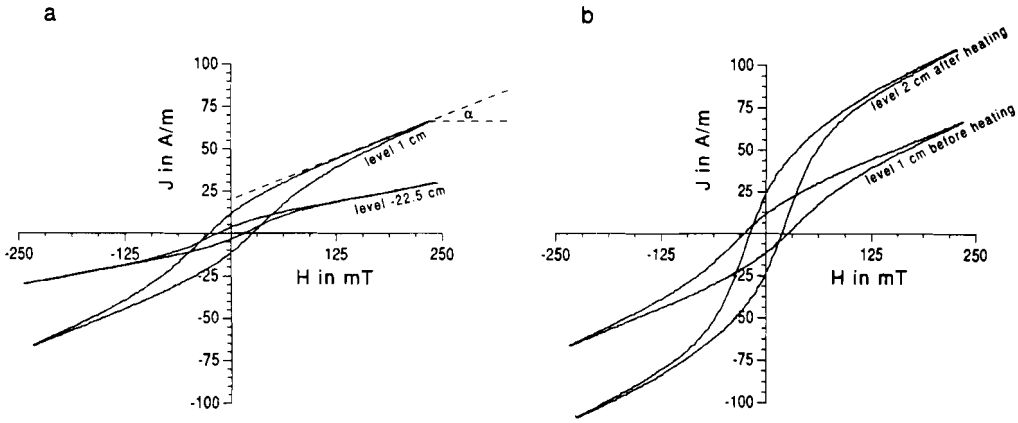


Fig. 5. (*a*) Hysteresis loops of samples from the white marl (-22.5 cm) and sapropelitic layer (1.0 cm). Saturation occurs at approximately 100 kA/m (-22.5 cm) and at 200 kA/m or higher (1 cm). The linear trend at higher fields (dashed lines) is due to the paramagnetic susceptibility of the clay minerals. It can be derived from these curves as the tangent (of α) and is approximately 110 (-23.9 cm) and 250 x 10^{-6} S.I. (1.0 cm). (*b*) Hysteresis loops of sapropel samples before heating (1.0 cm, same sample as in Figure 5*a*) and after heating to 360°C (2.0 cm). After heating, the sapropel samples have a higher saturation remanence.

ratio. This makes it rather unlikely that pyrrhotite is the mineral responsible for the observed changes in X_0, J_{rs}, and H_{cr}, despite the relatively high sulphur content in a sapropel [*Calvert*, 1983]. The presence of goethite, on the other hand, need not be surprising: it may be due to increased weathering in the dark brown and more permeable (laminated) sapropel. A similar line of reasoning with respect to the $H_{cr}/(J_{rs}/X_0)$ ratio excludes a possible small addition of hematite [*Dekkers and Linssen*, 1989].

Hysteresis loops (Figure 5*a*) show that a nonsapropel sample is saturated in a field of approximately 125 mT, whereas a sapropel sample is hardly saturated in the maximum field of 250 mT. This supports the idea that the sapropel samples may contain some high-coercivity mineral, i.e.,goethite. since these samples are not entirely saturated, only minimum values of J_s and H_c can be determined; complete saturation would slightly increase these estimates, corresponding with a decrease of the ratios H_{cr}/H_c and J_{rs}/J_s of the sapropel samples. The ratios H_{cr}/H_c and J_{rs}/J_s (table 1) of the nonsapropel samples are similar to the ratio's found by *Van Velzen and Zijderveld* [1990] who explain the discrepancy between their data and those of *Dunlop* [1986] by the presence of goethite and superparamagnetic (SP) magnetite grains. From the sapropel samples only, the ratios of the level 1.0 cm sample fit the data from *Van Velzen and Zijderveld* [1990], whereas the ratios from level 4.9 cm do not.

So far the conclusion that can be drawn from the observed rock magnetic parameters and some interparametric ratios is the likely presence of fine-grained magnetite in the white layers and even finer-grained magnetite in the grey/sapropelitic layer together with some (high-coercivity) magnetic mineral, possibly goethite. Without some additional data, for example, from X ray diffraction or microprobe analysis of magnetic concentrates, these rock magnetic parameters are not sufficient to make a more exact discrimination of the magnetic mineralogy. It is clear, however, that the grey layer and especially the intercalated sapropel show properties that markedly deviate from the underlying and overlying part of the record.

Susceptibility and Hysteresis Parameters During Thermal Demagnetization

During thermal demagnetization of the natural remanent magnetization (NRM), a number of specimens showed some scatter at temperatures higher than approximately 360°C. A small number of samples were heated to this temperature, and subsequently the rock magnetic measurements described above were repeated (Table 1). Most conspicuous is that after heating to 360°C also the sapropel specimens are also saturated at 125 mT, a value which is equal to the saturation fields of the nonsapropel samples before heating. This may be due to the removal of existing goethite.

J_{rs} itself does hardly change outside the sapropel, whereas in the sapropel, one sample showed a minor decrease and another sample showed a very strong increase (Table 1). The susceptibility after heating (X_{360}) increases the entire section with respect to X_0. The relative decrease of about 25-30% of J_{rs}/X_{360} with respect to J_{rs}/X_0 is the same for sapropel and nonsapropel samples. Growth of magnetite grains may have taken place in all specimens.

H_{cr} values from the sapropel, which showed maxima before heating, now show almost the same values as the nonsapropel samples. At the same time, all post-heating H_{cr} values are lower than those before heating. This is probably due to the removal of goethite and/or growth of magnetite grains: H_{cr} values decrease only slightly in the white layer but appreciably in the grey layer/sapropel.

In general, most rock magnetic properties that were characteristic for sapropel samples before heating, disappear

TABLE 1. Rock Magnetic Parameters From the Upper Kaena Interval, Before and After Heating to 360°C

Level, cm	Saturation, kA/m		J_{rs}, A/m		X_0, 10^{-6} S.I.		J_{rs}/X_0, kA/m		H_{cr}, kA/m		$H_{cr}/(X_0/J_{rs})$		H_{cr}/H_c		J_{rs}/J_s	
	Before	After	Before	After	Before	After	Before	After	Before	After	Before	After	Before	After	Before	After
41.6	200	100	3.2	2.7	240	250	13.3	10.8	40	26	3.03	2.44				
24.5	100	100	3.4	3.5	240	300	14.1	11.6	33	29	2.33	2.50				
20.7	100		3.6		280		13.0		35		2.70					
13.0	100	100	7.4	7.3	390	420	18.9	17.2	42	29	2.22	1.69				
4.9	400	100	16.9	14.1	610	670	27.8	20.8	78		2.78		4.19	2.00	0.48	0.56
2.0	400	100	12.3	20.4	470	1000	26.3	20.4	80	30	3.03	1.47		2.00		0.41
1.0	400		10.4		410		25.6		58		2.27		2.90		0.42	
-5.1	150	100	6.7	7.0	360	650	18.5	10.9	44	30	2.38	2.78		2.54		0.26
-13.0	100	100	3.0		250		12.2		32		2.63					
-23.0	100		3.4	3.4	260	370	13.2	9.3	32	30	2.44	3.23		2.61		0.38
-24.0	100		3.2		280		11.4		32		2.78		2.56		0.40	

Area from 4.9 to 1.0 cm is the sapropelitic layer. The ratio $H_{cr}/(X_0/J_{rs})$ is dimensionless, since H_{cr} has been converted from mT to kA/m. See text for further discussion.

upon heating to 360°C. The resulting changes in magnetic parameters are mostly due to growth of (preexisting) magnetite grains during heating. In the sapropel some goethite may be present. The presence of pyrrhotite cannot be ruled out, but considering, for example, the $H_{cr}/(J_{rs}/X_0)$ ratio, it is thought to play a relatively minor role. Although the rock magnetic parameters from the grey/sapropelitic layer are somewhat atypical with respect to the parameters of the white layers, the dominant carrier of the remanence is magnetite. This conclusion agrees with the results of thermal demagnetization (see next section). *Van Velzen and Zijderveld* [1990], in a rock magnetic study of the very similar Trubi marls from the Eraclea Minoa section in southern Sicily, also concluded that the most important magnetic mineral in these marls is fine-grained (SD) magnetite, with a small amount of goethite.

NRM COMPONENTS

Measurements of the natural remanent magnetization (NRM) were done with a three-axis vertical 2G Enterprises cryogenic magnetometer. Progressive stepwise thermal demagnetization was carried out using a μ-metal shielded tunnel furnace. In general, small temperature increments of 30° or 50°C were used. The demagnetization properties of the Pliocene marls from the Trubi and Monte Narbone formations are generally very constant, and we first give a general description before discussing the details of the upper Kaena interval.

NRM Components in the Trubi and Monte Narbone Marls

Typical thermal demagnetization diagrams of the Pliocene marine marls from southern Sicily generally show several distinctly different components [see *Langereis and Hilgen*, 1991]. Often, a small component is removed at low temperatures (90-120°C); since its direction is mostly related to storage, a viscous (laboratory induced) origin is likely. A component removed at 200°-250°C has a present-day field direction: it is easily distinguishable from the characteristic remanence removed at higher temperatures since it has a steeper inclination and a northerly declination, whereas the characteristic remanence shows a consistent 35° (average) rotation during the early Pliocene which reduces to 25° in the upper part of the Gauss Chron. A characteristic remanent magnetization (ChRM) is removed between 200°-250° and 580°-610°C. This ChRM actually consists of two components [*Langereis and Hilgen*, 1991], a low-temperature (LT) component removed between 200°-250° and 480°-510°C and a high-temperature (HT) component removed between 480°-510°C and the highest temperatures used. The LT component decays most strongly between 200° and 330°-360°C. Often, there is almost no decay between 360° and 480°C, but some scatter is frequently observed which is probably due to magnetic minerals being formed during heating. The origin of the LT component is uncertain, but the strong decay between 200° and 330°-360°C may indicate that either pyrrhotite or greigite is a carrier of the remanence in these marls [see *Leslie et al.*, 1990; *Linssen*, 1988, 1991]. The HT component decays most strongly between 540° and 580°C, resulting in a distinctive and narrow peak in the blocking temperature spectrum, which may be characteristic of ultra-fine-grained SD magnetite in a narrow grain size range and possibly of biogenic magnetite [cf. *Chang and Kirschvink*, 1985; *Lovley et al.*, 1987; *van Hoof and Langereis*, 1991].

Typically, outside polarity transition intervals, the LT and HT component have the same direction, including the characteristic tectonic rotation of the Caltanisetta basin. Inclinations of the ChRM are consistently shallower than the inclination of the present-day field at the locality. This inclination error is most likely due to sediment compaction [see *Laj et al.*, 1982; *Anson and Kodama*, 1987; *Arason and Levi*, 1990] since the marls with a higher carbonate content [cf. *Hilgen and Langereis*, 1989] and thus less compaction consistently show a smaller inclination error. This inclination error may be taken as an indication of an early origin of the magnetization [*Laj et al.*, 1982]. Together with the fact that the observed polarity pattern in these Pliocene marine marls shows an excellent correlation with the polarity time scale [*Zijderveld et al.*, 1986; *Langereis and Hilgen*, 1991],

these observations provide strong evidence that the ChRM has been acquired in a very early stage after deposition.

NRM Components in the Upper Kaena Interval

Throughout the upper Kaena interval, the remanence removed between 200° and 330°-360°C is small, some 15-20% of the total ChRM intensity (Figure 6*a* and 6*f*), whereas usually values of 30-50% are found. since the maximum blocking temperatures of pyrrhotite are within this temperature range [*Schwarz*, 1968, 1975; *Dekkers*, 1988], this supports our conclusion that in the upper Kaena interval (almost) no pyrrhotite is present. In some samples, the remanence direction starts to fluctuate at temperatures higher than approximately 360°C up to 480°-510°C (cf. Figure 6*a*). These fluctuations are probably caused by growth of existing magnetite grains and/or by production of SP magnetite grains from clay minerals during laboratory heating. This is illustrated by the often large increase of the magnetic susceptibility during stepwise thermal demagnetization at temperatures of 390°-420°C (Figure 7). At temperatures higher than 450°-480°C susceptibilities decrease again but remain always higher than values at room temperature. Remarkedly, susceptibilities of samples from the sapropel do not show the clear maximum between 390° and 480°C, but gradually increase to values some 30% higher than those at room temperature (Figure 7).

If these fluctuations in NRM direction take place, the direction of the LT component is determined by least squares fitting [*Kirschvink*, 1980] of demagnetization steps between 200°-250°C and 360°-390°C, if not, from steps between 200°-250°C and 450°-480°C. The direction of the HT component is ,where possible, determined from demagnetization steps higher than 480°-510°C; the origin was always included in the least squares fitting, even if the HT component does not clearly converge towards the origin (as in Figure 6*d*). The strong remanence decay of samples outside the reversal between 510°-540°C and the maximum temperature (Figures 6*a*, 6*b* and 6*f*) indicates fine grained magnetite as a carrier of the HT component. Maximum blocking temperatures (590°-610°C) are higher than expected for magnetite. This is probably not due to the presence of hematite, as discussed in the section on rock magnetism (and see, e.g., Figure 3) but may rather be due to the presence of cation-deficient magnetite [*Heider and Dunlop*, 1987; *Linssen*, 1991]. The majority of the vector diagrams with intermediate directions

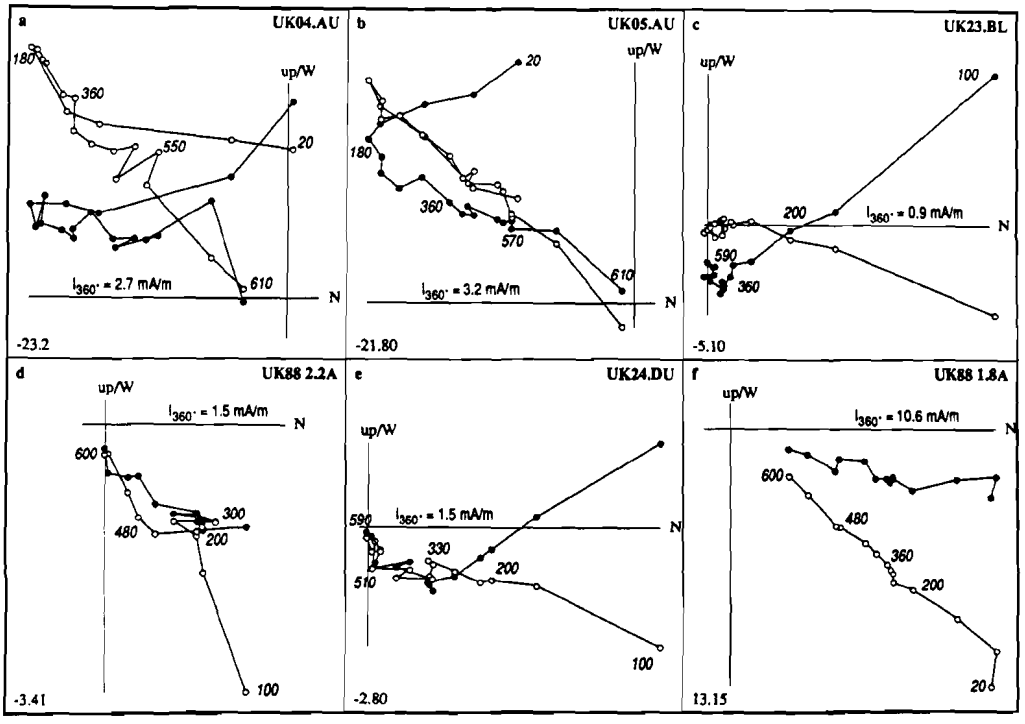


Fig. 6. Thermal demagnetization diagrams from the upper Kaena interval. The intensity at 360°C ($I_{360°C}$) and the stratigraphic level (in centimetres; down left) is given in each diagram; open (solid) symbols denote projection on the vertical (horizontal) plane; numbers are temperatures in degrees Celcius. (*a*) and (*b*) Stable reversed directions show some scatter at temperatures higher than 390°C due to growth of newly formed magnetite (cf. Figure 7) but (*f*) less noticeable in stable normal directions. (*c*), (*d*), and (*e*) Intermediate directions show a difference in low-temperature (LT, 200°-360°C) and high-temperature (HT, 480°-580/610°C) components.

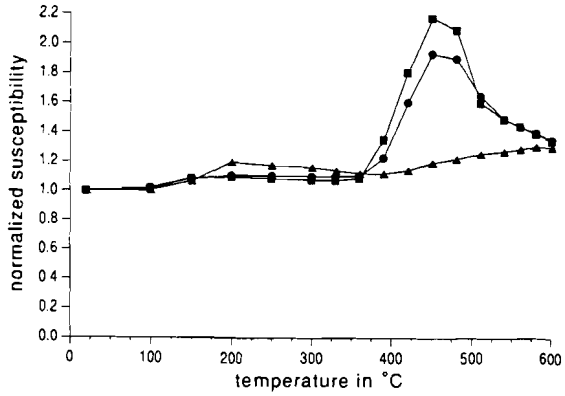


Fig. 7. Susceptibilities, normalized to their value at room temperature, during stepwise thermal demagnetization of nonsapropel samples (circles, squares) show a considerable increase between 360° and 450°C, and decrease again at higher temperatures. This probably indicates that growth of magnetite grains occurs, and/or that magnetite is formed from clay minerals during heating. A sapropel sample (triangles) does not show a maximum but more or less steadily increases to a 30% higher value.

(Figures 6*c*, 6*d* and 6*e*) are not as well behaved as the nontransitional diagrams, probably because the ChRM intensity is lowered during the reversal: a same amount of scatter due to growth of magnetite will then have a much stronger disturbing influence.

The nontransitional reversed (Figures 6*a* and 6*b*) and normal (Figure 6*f*) ChRM directions from the upper Kaena interval both show a distinct clockwise rotation. The mean stable, nontransitional directions from the upper Kaena interval (Figure 8) agree reasonably but not quite with the mean directions of the entire underlying and overlying polarity zones from the Punta Piccola section (Table 2). The actual angle between the mean reversed and normal stable directions of the upper Kaena interval is 11.5°, whereas this difference is only 4.6° in the Punta Piccola section.

The demagnetization diagrams within the transitional interval (Figures 6*c*-6*e*) show that LT and HT components seem not to have the same (intermediate) direction. Therefore, we have determined, where possible, both the LT and HT components separately throughout the entire sampled interval (Figure 9). In nontransitional samples (Figures 6*a*,6*b*,6*f*) the secondary component is clearly different from the LT + HT component. It may be argued that in transitional samples (Figures 6*c*-6*e*) the LT component is more strongly influenced by the secondary component due to the relatively low LT intensities during a reversal, while the intensity of the secondary component remains approximately the same. However, *van Hoof and Langereis* [1991] showed that during two older (N-R) reversals (lower Cochiti and lower Nunivak) from the same marine marls the LT component is reversed, and thus not a recent overprint, while the HT component is normal, and concluded that there is a magnetization depth lag between the two components. The very similar magnetic properties between the just mentioned records and the present record as well as the clearly different directions of secondary and LT components in nontransitional samples give us no a priori reason to assume that the LT component represents a (partial) overprint by secondary (subrecent) remanences in transitional samples. Admittedly, the coincidence of a sapropel being present in the upper Kaena record may give rise to increased weathering, but it can be seen that just in this sapropel LT and HT components have the same characteristic direction (Figure 9).

Although very rapid changes have been reported both in sediments and in lava flows [*Laj et al.*, 1988; *Coe and Prévot*, 1989], we feel that a single ChRM direction strongly deviating with respect to adjacent samples or an unreliable direction, due to low intensity and/or large scatter, must be interpreted with caution. Such directions are denoted by diamonds in Figure 9 and were rejected in the calculation of the VGP paths.

THE UPPER KAENA REVERSAL RECORD

Nontransitional Stable Directions

Outside the transitional interval, the directions of the HT component follow the same trend as the LT component. Mean directions of the HT component from the upper Kaena interval immediately before and after the reversal are not antipodal (by 11.5°), especially with respect to the declination (Table 2). They

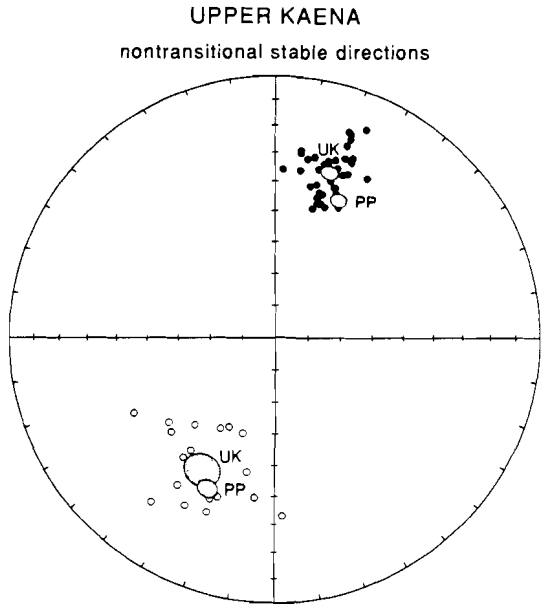


Fig. 8. Nontransitional stable reversed (open circles) and normal (solid circles) directions from the upper Kaena (UK) interval. The mean directions including their α_{95} are given by the shaded area. For comparison, the mean directions of the entire underlying and overlying polarity zones from the Punta Piccola (PP) section are also given. See also Table 2.

TABLE 2. Mean Normal (N) and Reversed (R) Stable (Nontransitional) Directions From the Upper Kaena (UK) Interval and From the Entire Underlying and Overlying Polarity Zones From the Punta Piccola (PP)Section

	n	D, deg	I, deg	k	α_{95}	r	γ_0	γ_c	ctmd
UK (N)	45	18.1	34.2	80.8	2.4	44.46			
							11.5	5.9	-
UK (R)	23	208.6	-42.3	33.1	5.3	22.33			
PP (N)	32	24.5	42.2	122.3	2.3	31.75			
							4.6	4.1	-
PP (R)	12	203.9	-37.6	219.2	2.9	11.95			
UK-PP (N)							9.4	3.1	-
UK-PP (R)							5.9	6.1	B

In both cases, the reversal test, denoted as ctmd (common true mean direction [see *McFadden and Lowes*, 1981] is negative. Also, the mean normal directions from the upper Kaena interval and the Punta Piccola section do not share a ctmd, while the mean reversed directions do, with classification B. Reversal (or ctmd) test and classification have been determined using simulation after *McFadden and McElhinny* [1990], except in the case of PP (N) versus PP (R). The actual (γ_0) and critical angle (γ_c) between every two tested populations are given

show a negative reversal test [*McFadden and McElhinny*, 1990] (Table 2). Average inclinations of the HT component (-42.3° and 34.2°) are shallower than the geocentric axial dipole field inclination for the present latitude (58°). The directions of the LT component show considerably more scatter, which may be mainly due to the lower intensity of this component (Figure 9) and the consequent difficulty in determining a reliable direction. We have tested whether the mean reversed (normal) stable direction from the upper Kaena interval and the mean reversed (normal) direction from the entire polarity zones in the Punta Piccola section have a common true mean direction [*McFadden and Lowes*, 1981] (Figure 8 and Table 2]. We have done this by inverting one of the two reversed (normal) directions and subsequently using the reversal test of *McFadden and McElhinny* [1990]. This test uses the same statistics as the earlier test of *McFadden and Lowes* [1981], except that it may be performed by simulation if the hypothesis of a common precision parameter must be rejected. The reversed directions have a common true mean direction, with classification B [*McFadden and McElhinny*, 1990], while the normal directions have not (Table 2).

Transitional Directions

The actual reversal is characterized by an eastward movement of the HT declination starting at level -15 cm; it changes via east (90°) at level -5 cm to normal polarities at 0 cm, where they show the familiar clockwise rotation. The HT inclination shows a gradual change from reversed to normal values; no steep inclinations are observed during the reversal. At level 0 cm the inclination is close to normal values. Inclinations of the LT component follow a similar gradual trend as HT inclinations, but

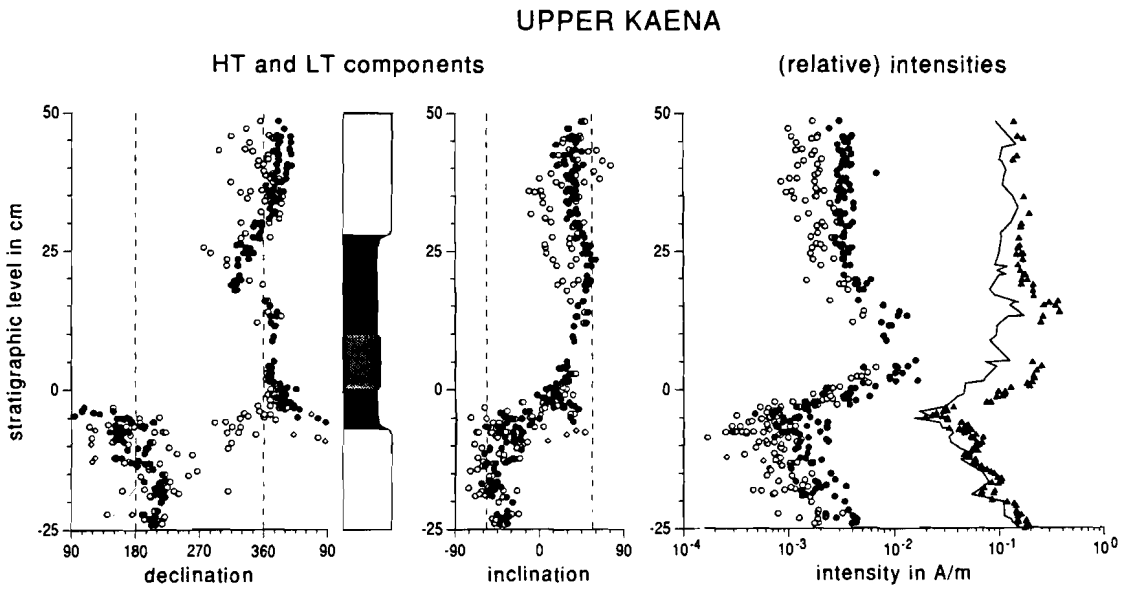


Fig. 9. Declinations, inclinations and intensities of the HT component (solid circles) and the LT component (open circles) during the upper Kaena transition interval. Diamonds denote unreliable directions (see text) and are not used for calculation of VGPs (Figure 10). Transitional directions, between -15 and 0 cm, proceed anticlockwise via east (HT component) or oscillate between east and west (LT component). After completion of the transition, an excursion in declination is seen between 17 and 30 cm; inclinations are slightly steeper. Relative intensities are denoted by triangles ($I_{360°C}$ / X_0) and by a solid line ($I_{360°C}$ / J_{rs}) and reach a minimum at the level of greatest directional changes, at -5 cm. During the excursion, no apparent changes in intensities are found.

LT declinations differ from those of the HT component. Whereas the HT declination has a clear easterly component, the LT declinations show both easterly and westerly deviations during the transition. We note, however, that in this interval LT intensities are extremely low (Figure 9) and the reliability of the LT directions may be strongly reduced. The ChRM intensity starts to decrease at the -20 cm level, slightly before directional changes start to take place (at -15 cm). The intensity minimum is some 10% of prereversal and postreversal values and occurs at -5 cm (Figure 9), at exactly the same level of the largest directional changes. The intensity seems to regenerate to prereversal values at approximately 5 cm, slightly after directions have returned to normal values (at 0 cm).

The directions remain normal between levels 0 and 14 cm, and the average declination agrees well with the generally observed rotation. In this interval, the ChRM intensities show their maximum values.. Between levels 14 and 30 cm, the declination shows a considerable deviation of some 50°-60° to the west, and the inclination shows values which are generally some 15° steeper than the average inclination. It would seem that a directional excursion is recorded after the actual upper Kaena reversal. The ChRM intensities, however, do not show a noticeable decrease. Above 30 cm, the directions show a stable normal polarity direction which has a mean declination still less than expected for these marls in the Punta Piccola section (Table 2).

Relative Intensities

The observed changes in magnetic mineralogy over the reversal record do not allow the intensities of the LT and/or HT component to be used as a measure of the relative paleo-intensity of the geomagnetic field during the reversal. Usually, ARM is used as a normalizing factor to determine relative paleontensities (NRM/ARM ratio) [see *Levi and Banerjee*, 1976; *King et al.*, 1983]. On the base of earlier results of older reversal records in the same marine marls from Calabria and Sicily, the presence of pyrrhotite was suspected, in which case the method developed by *King et al.* [1983] cannot be used [*Linssen*, 1988, 1991]. Therefore, we routinely determined X_0 and J_{rs} (see Figure 4) rather than ARM for a large number of samples.

We have calculated the ratio of ChRM intensity (at 360°C) and X_0 as a first order approximation of the paleointensity (triangles in Figure 9). The magnetomineralogical effect of the sapropel seems not to be entirely eliminated since there is still a maximum corresponding to the maximum of the ChRM intensity (Figure 9) and other rock magnetic parameters (Figure 4). This is probably due to the fact that X_0 is largely determined by the paramagnetic clay fraction. Therefore, this ChRM/X_0 ratio is not representative for geomagnetic field intensity changes. since we have determined the paramagnetic contribution only for a few samples, we cannot use the (corrected) X_0 to normalize the ChRM intensity. Instead, we have used the ratio ChRM intensity (at 360°C) and J_{rs} intensity (at 2 T) as a measure of the relative paleointensity (line in Figure 9). The J_{rs} value (Figure 4) of the same or the closest possible stratigraphic level was taken to calculate this ratio for each specimen. The increase of ChRM intensity in the sapropel is now largely compensated by the increase of J_{rs} in the sapropel and it confirms that this increase of the ChRM intensity is a purely mineralogical effect.

Virtual Geomagnetic Poles

The virtual geomagnetic poles (VGPs) of both LT and HT components were calculated after applying an average correction of 25° for the clockwise rotation of the location, i.e., that of the Punta Piccola section. Correction for the inclination error was not applied. The HT component VGP path (Figure 10) initially follows roughly a path some 60° east of the site, loitering southwest of Australia, then quickly passes the equator etween India and Australia and goes via Japan to high northern latitudes approximately 180° east of the site. Furthermore, the VGP path shows an easterly migration of more than 120° during the reversal. The excursion following the reversal shows VGPs more or less antipodal to the VGP path of the actual reversal itself. They are confined to North America and remain at latitudes higher than 30°N. The LT component VGP path (Figure 10) shows a more complicated picture. Rather than following a particular great circle, it oscillates between two meridians: one passing over Australia and one passing over South and North America. Surprisingly, no VGPs intermediate between these two great circles are found. Initially, most VGPs are on the southern hemisphere of the Australian path while relatively few are found on the (South) America path. During the final stage, VGPs are only found on the (North) America path, while there are no VGPs on the northern hemisphere of the Australia path.

DISCUSSION

The upper Kaena record is the last reversal from the Punta Piccola section in a series of four successive reversals which were sampled by our laboratory. Recently, *Linssen* [1991] reported the records of the lower Mammoth, upper Mammoth and lower Kaena reversals. Such a sequence of successive reversals may yield useful information on the time dependent character of transitional field behavior during longer time spans, provided that the records reflect true geomagnetic field behavior.

The average sedimentation rate during the Kaena subchronozone is 4.3 cm/kyr on the basis of the magnetostratigraphic results [*Zachariasse et al.*, 1990] and their correlation to the polarity time scale *Berggren et al.*, [1985]. A more precise estimate is based on the exact location of both lower and upper Kaena reversal [*Linssen*, 1991; this paper] and the correlation of the individual sedimentary cycles to the astronomical (precessional) solutions [*Hilgen*, 1991], yielding an average sedimentation rate of 4.5 cm/kyr. The sedimentation rate during the deposition of the grey marls, including the intermittent occurrence of a sapropelitic layer, is probably slightly higher than the average 4.5 cm/kyr due to increased continental runoff [*de Visser et al.*, 1989]. Although slight changes in sedimentation rate will not essentially alter any duration of features in the upper Kaena reversal, they may cause noticeable changes in paleoredox conditions, as will be discussed later.

A conventional interpretation of the record would suggest that the major directional changes take place in a sedimentary interval of approximately 15 cm or 3.3 kyr, followed by an interval of stable normal directions (15 cm or 3.3 kyr) and the excursion (16 cm or 3.6 kyr). The period of lowered intensities associated with the reversal takes some 25 cm (5.6 kyr), while

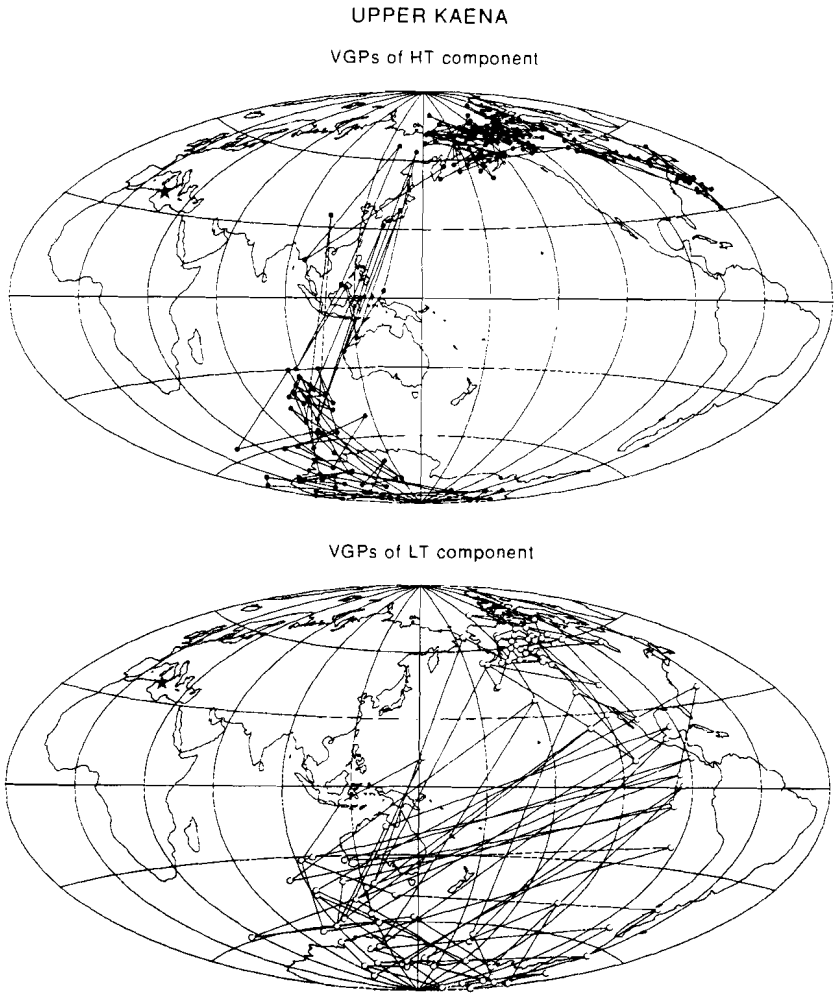


Fig. 10. Aitoff projection of virtual geomagnetic poles (VGP) of the HT and LT component from the upper Kaena reversal. Star denotes site location. The VGP path of the HT component shows a significant eastward migration during the reversal. The HT component VGPs of the "excursion" are restricted to eastern North America. The VGPs of the LT component oscillate between a path over India/Australia and a path over the Americas. VGPs of the excursion are similar to those of the HT component VGPs (not shown for readability).

the total interval from consistently stable reversed to stable normal directions takes 50 cm or 11 kyr. These durations seem to agree well with existing estimates of 4-10 kyr [e.g., *Clement and Kent*, 1984). However, before interpreting the reversal record in terms of geomagnetic field behavior is important to first establish the extent to which the present sedimentary record actually represents geomagnetic field changes.

Smoothing of the NRM

A major drawback of sedimentary reversal records is the lack of detailed knowledge of the NRM acquisition mechanism, although very recently a number of studies on important aspects related to the acquisition and recording process have been published [e.g., *deMenocal et al.*, 1990; *Lund and Karlin*, 1990; *Karlin*, 1990*a,b*; *Leslie et al.*, 1990; *Rochette*, 1990; *van Hoof and Langereis*, 1991].

A common mechanism by which many types of sediment, typically deep-sea sediments and lake sediments, acquire their remanence is postdepositional remanent magnetization (PDRM) [*Irving*, 1957], which acquired below the bioturbated zone [0-10 cm) during early dewatering and consolidation of the sediment. Estimates for the PDRM lock-in zone range from 10 to 100 cm below the sediment surface. Recent data from the Brunhes-Matuyama reversal indicate a 42-cm magnetization depth lag (half fixing depth) with a lock-in zone of 10-20 cm in high deposition rate sediments (180-370 cm/kyr) [*Okada and*

Niitsuma, 1989], and a PDRM acquisition depth of ~16 cm for eight moderately slow deposition rate deep-sea cores (1-8 cm/kyr) [*deMenocal et al.*, 1990]. Although PDRM acquisition inevitably involves smoothing of the geomagnetic record, it is often assumed that PDRM acquisition preserves the true geomagnetic record (albeit with a depth/time lag), provided sedimentation rates are high enough, at least more than 1 cm/kyr.

The smoothing process of reversal records has been discussed by various authors [e.g., *Hoffman and Slade*, 1986; *Rochette*, 1990]. The most evident type of smoothing is due to specimen thickness. Secular variation is generally assumed to be averaged out in moderately slowly deposited sediments; here, a single specimen represents some 500 years. However, the British secular variation master curve during the past 3000 years [*Clark et al.*, 1988] shows differences in declination of some 70° within this time span and occasionally differences of more than 40° within 500 years. If smoothing would occur due to sample thickness only, one would expect the secular variation to be expressed both before and after the reversal. In order to estimate the maximum amplitudes of secular variations, we have applied a 500-year filter to the British master curve, using time steps of 100 years, i.e., approximately the sampling resolution of the present record. It can be seen that amplitudes of more than 30° may then still be expected, although high frequency details are lost (Figure 11). Comparison with the nontransitional upper Kaena directions shows that prereversal stable directions show on average similar amplitudes, whereas postreversal stable directions show a maximum amplitude of 15° (Figure 11). However, it must be realized that the apparently similar amplitude of the prereversal directions is for a large part due to a few outlying data points. Although the large amplitudes present in the British master curve during the last 3000 years need not be representative for older time spans, we suspect that some additional smoothing, other than from specimen size only, has taken place. This additional smoothing is most probably related to the mechanism of remanence acquisition in these marine marls.

Smoothing will (strongly) affect short-scale features like secular variation, but the general pattern of the reversal may be preserved if the smoothing window is smaller than the reversal duration. Furthermore, it is often assumed that the smoothing window must be smaller than the observed duration of rapid directional changes. If this were true, we would arrive at a smoothing window of less than ~5 cm or 1 kyr, while the onset of the excursion even shows a 40° change in declination within 2 cm or 400 years (see Figure 9). However, *Hoffman and Slade* [1986] have shown that due to smoothing directional changes may appear more rapidly than in the original reversal record. Also *Rochette* [1990] has argued that in the case of nonantipodal stable directions and a smoothing window large compared to reversal duration, this assumption is not true. In such a case, the intensity change is observed over the width of the smoothing window, while the directional change, although smooth, is much more rapid: transitional directions follow a great circle defined by the nonantipodal stable directions. Nonantipodality mainly in inclination leads to a zonally confined path, while nonantipodal declinations produce a VGP path some 90° away from the site.

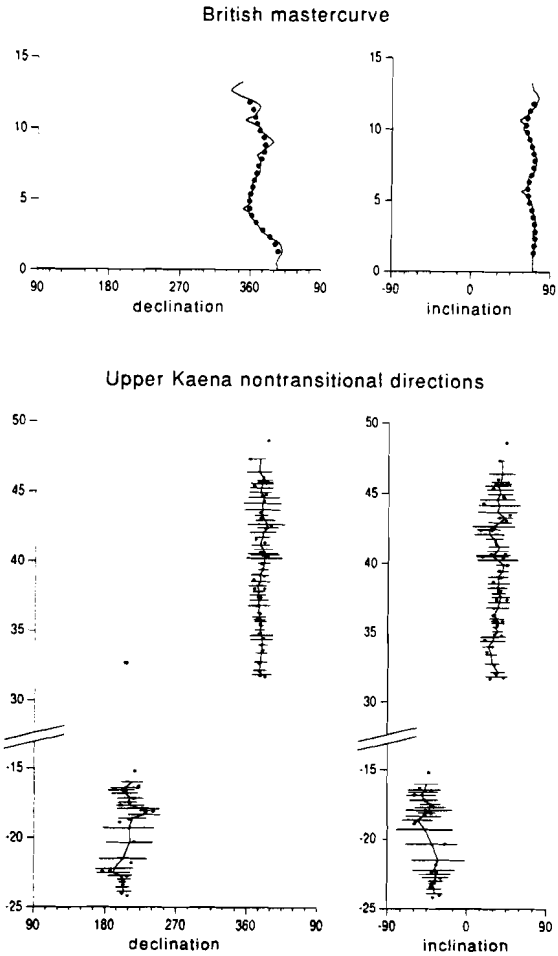


Fig. 11. British secular variation master curve (thin line) [after *Clark et al.*, 1988], sampled with a 500-year filter (dots) to represent specimen thickness, at time intervals of 100 years (sampling resolution). The time scale (1000 B.C. to 1950 A.D.) has been converted to stratigaphic level using a sedimentation rate of 4.5 cm/kyr. For comparison, also the stable pretransitional and posttransitional upper Kaena directions are plotted (dots). Line denotes a three-point moving average; bars indicate the three-point α_{95}. It can be seen that the directional variations of the upper Kaena interval are significantly less than those derived from the British master curve, indicating smoothing in addition to that resulting from specimen thickness only.

In our record, we find nonantipodal directions in the upper Kaena interval, mainly due to a ~10° difference in declination [Table 2). Indeed, intensity changes take longer than directional changes [5.6 and 3.3 kyr, respectively) and VGP paths follow a longitude well away from the site longitude (Figure 10). However, the HT component VGPs are located east rather than west of the site as should be expected, while LT component

VGPs are found at both 90° east and 90° west. On the other hand, true nonzonal behavior is quite indifferent to filtering, and a commonly found feature of reversal records like a typically nonzonal and longitudinally confined path may indicate a relatively stable feature of the transitional field since both smoothing and true record give similar results [*Rochette*, 1990].

Many reversal records show the occurrence of loitering (VGP "hangups") and rapid rebounds/excursions which are not easily explained by smoothing, although slight variations in the shape of the averaging filter may account for these features [*Rochette*, 1990]. It is therefore important to discuss the acquisition process and corresponding lock-in mechanism in some more detail, since the rythmically bedded Pliocene marls may induce a complex sedimentary filter indeed.

NRM Acquisition Processes in a Cyclic Environment

Geochemical and rock magnetic investigations on whole sediment and magnetic concentrates from both low and high carbonate marls from the Calabrian Trubi formation by *Linssen* [1991] have shown that the major magnetic minerals (pyrrhotite, magnetite) in these marls must have been formed by authigenic processes in the early stages after sediment deposition. *van Hoof and Langereis* [1991] have suggested that these minerals are possibly of bacterial origin, because indications for biogenic magnetites were also found by *Chang and Kirschvink* [1985] in very similar upper Miocene marine marls from Crete [*Langereis*, 1984]. *Linssen* [1991] concluded that pyrrhotite, in coexistence with pyrite, is relatively more abundant in the low carbonate (grey) marls, while magnetite is relatively more abundant and less fine-grained (PSD rather than SD) in the high carbonate (white) marls. This may indicate that in the grey marls the progressive series of redox reactions involved in the decomposition of organic matter [*Froelich et al.*, 1979] has progressed into sulfate reduction, while in the white marls they are mainly restricted to Mn and Fe reduction. The cyclic variation in color, carbonate content, and rock magnetic properties strongly suggests that these lower Pliocene marine marls have been subject to fluctuations in [paleo)redox conditions. In the cyclic Monte Narbone marls the same processes have most probably played a role, considering the very similar sedimentary properties [*Hilgen*, 1987; *de Visser et al.*, 1989] and rock magnetic properties [*Linssen*, 1991; *Langereis and Hilgen*, 1991].

Both magnetotactic and dissimilatory iron reducing bacteria have been proposed as the main producers of ultra fine authigenic magnetite in anoxic sediments [*Lovley et al.*, 1987; *Bazylinski et al.*, 1988] as well as in suboxic sediments, at the top of the iron reduction zone [*Karlin et al.*, 1987; *Karlin*, 1990*a*]. In suboxic sediments deposited at moderate rates (1-5 cm/kyr) the reaction sequence does not seem to progress into sulfate reduction [*Karlin et al.*, 1987], although *Linssen* [1991] has observed both pyrite and pyrrhotite in magnetic concentrates from the Trubi marls. The finely laminated structure of the (top part of the) sapropel is due to the absence of burrowing and indicates oxygen depleted conditions. Suboxic conditions are furthermore in agreement with a possible higher organic input due to increased continental runoff [*de Visser et al.*, 1989]. Indeed, the sedimentological and rock magnetic properties of the grey and white layers show a striking similarity with those of suboxic sediments from the Bettis area in the NE Pacific which were studied in detail by *Karlin* [1990*a*]. The fresh colors of the grey marls and sapropel are the same (from dark/brown to green and grey) and, if representative of their original, primary colors during deposition, they indicate a preserved paleoshift in redox conditions from surface oxidation to Mn and Fe reduction [*Lyle*, 1983; *Karlin*, 1990*a*]. Apparently, paleoredox conditions must have been (cyclically) fluctuating to preserve these colors, because in steady state redox conditions the Fe redox boundary and the magnetite enrichment zone would have migrated upward at a constant rate with sedimentation. In addition, magnetic properties of the grey/sapropelitic layer such as higher J_{rs}, higher H_{cr} (Figure 4) and higher saturating fields (Figure 3) are similar to those found for the transition zone between nitrate and iron reduction in the Bettis sediments [*Karlin*, 1990*a*]. Thus, following the conclusions of *Karlin* [1990*a*], authigenic formation of pure and ultra fine-grained magnetite is the most likely mechanism for the observed rock magnetic changes. A change to a lower sedimentation rate during deposition of the overlying white marl, with less organic input, will result in a shift in paleoredox conditions, similar to trapping Mn oxides and organic carbon [*Finney et al.*, 1988], these nonsteady state conditions can promote preservation of the authigenic magnetite [*Karlin*, 1990*a*]. Clearly, additional geochemical data would be valuable to substantiate the presence of buried paleoredox indicators.

It may be expected that authigenic magnetite formation and the resulting acquisition of a chemical remanent magnetization (CRM) or a PDRM together with cyclically fluctuating redox conditions will result in a complex mixture of various natural remanent magnetizations, acquired at various times and at various depths. Indeed, *van Hoof and Langereis* [1991] have shown that in two older reversal records, the lower Nunivak and lower Cochiti from the same marine marls, there is a relative magnetization depth lag between LT and HT component, the LT component becoming locked at shallower depth than the HT component. This relative depth lag is 10-35 cm and is restricted to the beige and white marls. In the grey marls, no relative depth lag was observed, which is similar to the results from the upper Kaena interval (Figure 9). This relative depth lag could be due to a larger grain size of the LT magnetic mineral (pyrrhotite?) and hence a shallower PDRM lock-in depth. Alternatively, this relative depth lag may be due to a time lag in CRM acquisition, by which the LT magnetic minerals grow through their critical blocking volume well before the HT magnetic mineral [magnetite). In addition to a relative depth lag, both LT and HT component show an absolute magnetization depth lag, again in the beige and white marls only - which amounts to at least 60 cm (LT) and 70-80 cm (HT) in the lower Cochiti record. This is well below the PDRM acquisition depth of ~16 cm due to early sediment consolidation, as found in similar deposition rate sediments [*deMenocal et al.*, 1990], and it favors a delayed ChRM acquisition as a more likely mechanism. We note, however, that still little is known about other factors affecting the natural variability of the lock-in process.

A considerable absolute magnetization depth lag is also found in relation to the upper Kaena reversal. Detailed paleomagnetic sampling of the entire (reversed) Kaena subchronozone in the Punta Piccola section included a sapropelitic layer in the grey marl ~100 cm below the sapropel of the upper Kaena interval. The preliminary paleomagnetic results show that in this layer intermediate and normal polarity LT and HT components are recorded. These components represent a NRM which acquired the post upper Kaena normal polarity geomagnetic field in a small and selective interval, the sapropelitic layer, at a depth of more than 100 cm. This may be due to continued authigenic CRM acquisition in this specific depth interval, or to PDRM acquisition of already existing authigenic magnetic minerals in the not yet consolidated (still high-porosity, or high water content) sapropelitic layer.

Apparently, the complicated mechanism of NRM acquisition and especially the timing of remanence acquisition makes it difficult to derive transitonal field behavior from records as presented here. Especially with respect to details of sedimentary reversal records, such as the upper Kaena excursion, we must refrain from interpreting them as true geomagnetic field behavior. Considering the observed delayed NRM in the 100 cm lower grey/sapropelitic layer, it is likely that the same or a similar processes have played a role in the upper Kaena grey/sapropelitic layer. In that case, the recorded "excursion" may be a partial record of the actual reversal, probably smoothed by the authigenic magnetite production interval, while the transition is also recorded slightly deeper (~25 cm; see Figure 9) in the grey/sapropelitic layer, under (slightly?) different redox conditions. These conditions involve either delayed CRM acquisition due to continued growth of magnetic minerals or delayed PDRM acquisition of the already existing and preserved ultra-fine magnetite. The boundary between these two different acquisition conditions is probably located at the 16-17 cm level, at the onset of the "excursion" where the sudden change of some 40° in declination occurs. Additional evidence for this assumed boundary is given by slight but equally sudden changes in H_{cr} and J_{rs}/X_0 (Figure 4), indicating an abrupt change to less fine-grained magnetite. It may be possible that in an environment critical with respect to the balance between porosity (water content) and grain size, a slight change in grain-size determines whether PDRM or CRM acquisition dominates as the remanence lock-in process.

True Geomagnetic Field Behavior?

The complex NRM acquisition process makes it extremely difficult to make a distinction between real features of the geomagnetic field and sedimentary artifacts. As discussed above, we do not interpret the excursion to be a geomagnetic feature. The low ChRM intensities during the "grey layer reversal" show a clear minimum which is unlikely to be caused by a lithological change since upon normalization with J_{rs} the minimum remains, while on the contrary the maximum in the sapropel is eliminated by normalization with J_{rs} (Figure 9). The lower ChRM intensities could therefore be related to an actually lower intensity of the transitional geomagnetic field but could also result from the averaging of normal and reversed directions. During the following "excursion", ChRM intensities are virtually constant, and normalization by J_{rs} does not significantly alter this trend. This could be due to the averaging of a relatively large normal component and small reversed (or intermediate) component, causing no apparent lowering of the intensity, but only "excursional" directions.

An important observation is the fact that both in the "excursion" as well in the "grey layer reversal", VGPs of the HT component are to a considerable extent longitudinally confined. They are constrained to two roughly opposite longitudes: those from the "grey layer reversal" are between India and Australia, while those from the "partial reversal" pass over North America (Figure 10). This corresponds respectively to a clockwise and an anticlockwise swing in declination, a difference which could be explained by a different degree of smoothing [*Hoffman and Slade*, 1986]. The rapid alternation of the LT component VGP path between the Americas path and the Southeast Asia path is not easily explained, but switching between these two paths has also been observed in other transitions, e.g., the Blake event [*Tric et al.*, 1991*b*], the upper Olduvai [*Herrero-Bervera*, 1989] and the lower Kaena [*Linssen*, 1991]. A different degree of smoothing, e.g. due to differential CRM acquisition at sampling scale, in the mottled and gradational transition from grey to brown, could perhaps cause this apparent "rapidly fluctuating" behavior. Considering the clearly delayed NRM acquisition processes that have played a role in these sediments, we find it difficult to believe that these rapid VGP alternations between Asia and the Americas are actually due to rapid fluctuations of the transitional geomagnetic field.

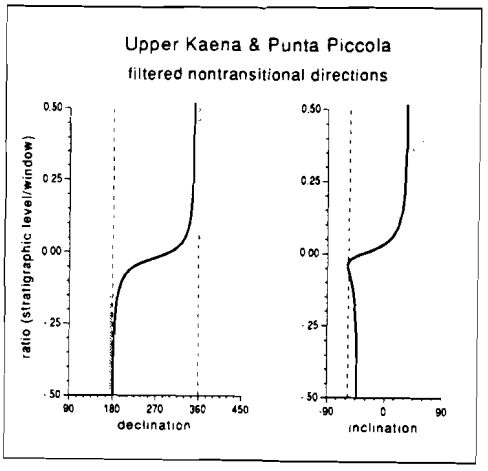
Tric et al. [1991*a*] found that in many reversal records from the past 10 m.y. these same two longitudinal bands are preferred, albeit in a statistical way, and they suggest that this confinement may be due to a significant dipolar component being present during the transition. They argue that smoothing of the geomagnetic signal of widely different lithologies, sedimentation rate and geographical position is unlikely to yield very similar records. On the basis of these results, *Laj et al.* [1991] furthermore suggest that there may be a link between reversal paths, the historical magnetic field and core flow pattern, and temperature distribution in the mantle. The observation that the same reversal produces VGP paths on opposite sides of the globe, however, complicates the dipolar interpretation [*Bogue*, 1991]. Recently, *Clement* [1991] used several records from the Brunhes-Matuyama to derive a reversal model involving an equatorial octupole that can produce the antipodal VGP paths like those observed. similarly, he argues that although systematic errors may result from smoothing or biasing, the geographical grouping of transitional VGPs from widely spaced sites and from different types of paleomagnetic recorders provides a strong argument that it reflects a real feature of the geomagnetic field. He also notes that often VGPs from nearby reversal records, or even within one reversal record, fall within both longitudinally confined peaks.

In order to test the extent to which the confinement of the upper Kaena record is determined by smoothing of nonantipodal directions, we have applied the approach followed by *Rochette* [1990]. We derive two artificially smoothed records by taking

the average nontransitional directions from both the Punta Piccola and the upper Kaena interval (Table 2). The directions from the Punta Piccola sections are almost antipodal with respect to declination but not with respect to inclination. Smoothing results in an anticlockwise progression of the declination, and inclinations first pass through zero and become steeply normal before returning to the average value. The corresponding VGP path shows the expected and typically (near-sided) zonal behavior (Figure 12). If the reversed inclinations had been steeper than the normal inclinations, we would of course have found a far-sided VGP path. The upper Kaena stable directions are mainly nonantipodal in declination, and not surprisingly, the VGP path follows the great circle defined by those directions and passes over North America. In neither case VGPs between Asia and Australia are found.

Comparison with the actually observed VGP paths (Figure 10) shows that the "grey layer reversal" initially proceeds along an Asian path, according to both HT and LT component, in contrast to the artificial records. The HT component remains confined to this path, although a strong eastward deviation is observed. The LT path, on the other hand, starts alternating between Asia and the Americas, while the last part is mainly confined to North America. Most conspicuously, the LT component VGPs on the Americas side of the globe are exactly on the path defined by smoothing of the upper Kaena stable directions (Figures 10 and 12). In addition, all VGPs from the "excursion" are located in North America, both from the HT component (Figure 10) and from the LT component (not shown in Figure 10 for readability, but see Figure 8). Hence it seems as if initially the reversal in the grey layer is a true recording of a path over Asia but becomes increasingly confined to a path over the Americas which is the same as that determined by smoothing of nontransitional directions from the upper Kaena interval. The following "excursion" is then possibly only a partial and in addition strongly smoothed record of the same reversal, since its VGPs are only confined to North America.

Finally, an earlier interpretation of the preceeding lower Kaena and both Mammoth reversals [*Linssen*, 1991] suggested that the major directional changes of these three reversals take place in 5-10 cm, or in 1-2 kyr. The intervals of minimum intensities are, on the contrary, very similar (20-30 cm or 4-6 kyr) to the interval found in the upper Kaena record. The character of the complete Mammoth and Kaena reversal succession seems rather complex: the lower Mammoth is recorded as a single reversal, the upper Mammoth is followed by an excursion in inclination, the lower Kaena is followed by a full excursion to normal polarities and high intensities, and the upper Kaena again is followed by an excursion. The VGP paths of the older reversals are generally well confined to great circles: the one from the lower Mammoth passes over India and those from the upper Mammoth and lower Kaena pass roughly over North and South America, while the R-N path of the full excursion follows a path over India again [*Linssen*, 1991]. VGPs of the upper Kaena are confined to a path between India and Australia (HT component) or alternate between this path and a path over the Americas (LT component), while the VGPs of the excursion are confined to North America. However, the fact that these earlier interpreted



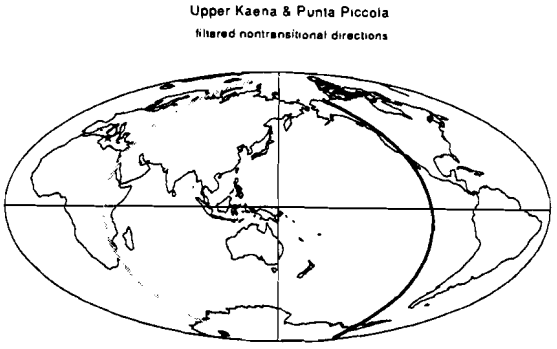


Fig. 12. Filtering of the mean pretransitional and posttransitional directions of Table 2, from the Punta Piccola section (shaded line) and from the upper Kaena interval (solid line), using the approach of *Rochette* [1990]. The reversal occurs - instantaneously - at level 0.00; ordinate gives the fraction of the smoothing window (minus 0.5) that "sees" the normal polarity. The corresponding VGPs on the Aitoff projection show a typically zonal and near-sided transition (Punta Piccola) and a path over western North America (upper Kaena). The final part of the latter coincides with the final part of the LT component VGP path (Figure 10).

records have been established from the same sediments, with the same cyclically fluctuating lithology, warrants a reinterpretation in terms of sedimentary artifacts rather than in terms of true geomagnetic field behavior.

CONCLUSIONS

A detailed paleomagnetic record of the upper Kaena (R-N) polarity reversal from southern Sicily shows that two characteristic components can be determined separately: a low-temperature (LT) component and a high-temperature (HT) component. Both components have been acquired in an early stage following deposition and are probably carried by

authigenically formed and microbially mediated magnetic minerals in a suboxic environment.

A conventional interpretation of the record would suggest a transition which takes places in approximately 3.3 kyr. with lowered intensities during 5.6 kyr. After a stable period of 3.3 kyr, it is followed by a significant "excursion" during 3.6 kyr.

However, based on the lithology, the rock magnetic and paleomagnetic properties, as well as on earlier results from the same marls [*van Hoof and Langereis*, 1991], we conclude that the features observed in the upper Kaena record, and also those in the three preceding records [*Linssen. 1991*], are largely determined by the remanence acquisition process in these sediments. Most likely, the remanence has been smoothed by CRM and/or PDRM acquisition, at least to the extent that finer details such as secular variation and possible rapid geomagnetic fluctuations have been lost from the record. More importantly, we conclude that cyclically fluctuating paleoredox conditions have resulted, first in a delayed paleomagnetic record in a carbonate-poor (and possibly higher-sedimentation rate) grey/sapropelitic layer due to magnetite preservation and second in an apparently posttransitional "excursion" which we believe is a partial, strongly smoothed but less delayed record of the same [upper Kaena) reversal.

It is remarkable that the observed VGP paths are confined either to a longitude between India and Australia, or to a longitude over South and North America. These two geographically confined paths were recently suggested to be related to the reversal process [*Tric et al.*, 1991*a*; *Laj et al.*, 1991; *Clement*, 1991]. The longitudinal confinement of the high temperature component to a path over India/Australia may be a real feature of the geomagnetic field. The major confinement of the LT component and of the "excursional" HT component to a path over the Americas agrees well with the path derived from smoothing the nonantipodal stable directions from the upper Kaena record.

This study emphasizes the importance of including detailed paleomagnetic/rock magnetic, lithological/geochemical data in studies of sedimentary reversal records in order to determine the reliability of the observed "transitional" features. We note that numerous "geomagnetic" features like longitudinal confinement, excursions/rebounds and rapid fluctuations from sedimentary records have been reported in literature which, according to the results of our study, could be due to local biasing effects.

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PART 3

REVIEW OF THE RECORDS

Chapter 10

Data from eight geomagnetic polarity transitions recorded in southern Italy published by Linssen (1988; 1991)

Together with 8 transition records published by Linssen (1988; 1991) the 10 transition records presented in this thesis form a sequence of 13 records of which 5 (upper Thvera, lower and upper Sidufjall, lower and upper Nunivak) have been obtained as double records. These are all recorded in the Pliocene Trubi and Monte Narbone formation in Southern Italy. In chapter 11 of this thesis a review the 18 reversal records will be given. In this review we will also refer to the data published by Linssen (1988; 1991). In order to aid the reader we will show his data in the present chapter. The data are presented as declination, inclination, intensity diagrams and as a virtual geomagnetic pole (VGP) path on an Aitoff world map projection. It concerns the data of five transition records from Calabria as well as the LMS, UMS and LKS transitions from Sicily. In the VGP diagrams, the data from Sicily have been corrected for the 35° clockwise rotation of the sedimentary basin. With exception of the two Mammoth transitions and the lower Kaena transition, the HT as well as the LT component have been determined and are shown as black dots (HT) and circles (LT) in the declination, inclination and intensity plots. The LT component is not always used in the VGP diagrams because the this component is much more scattered than the HT components. Between the declination and inclination diagrams a lithological column is drawn, the different colours representing the white, beige and grey layers. Light grey colours in the columns represent layers with brown-coloured spots. In some sections, the colour differentiation between the layers was less obvious. In that case, the colours in the lithological columns are also less pronounced.

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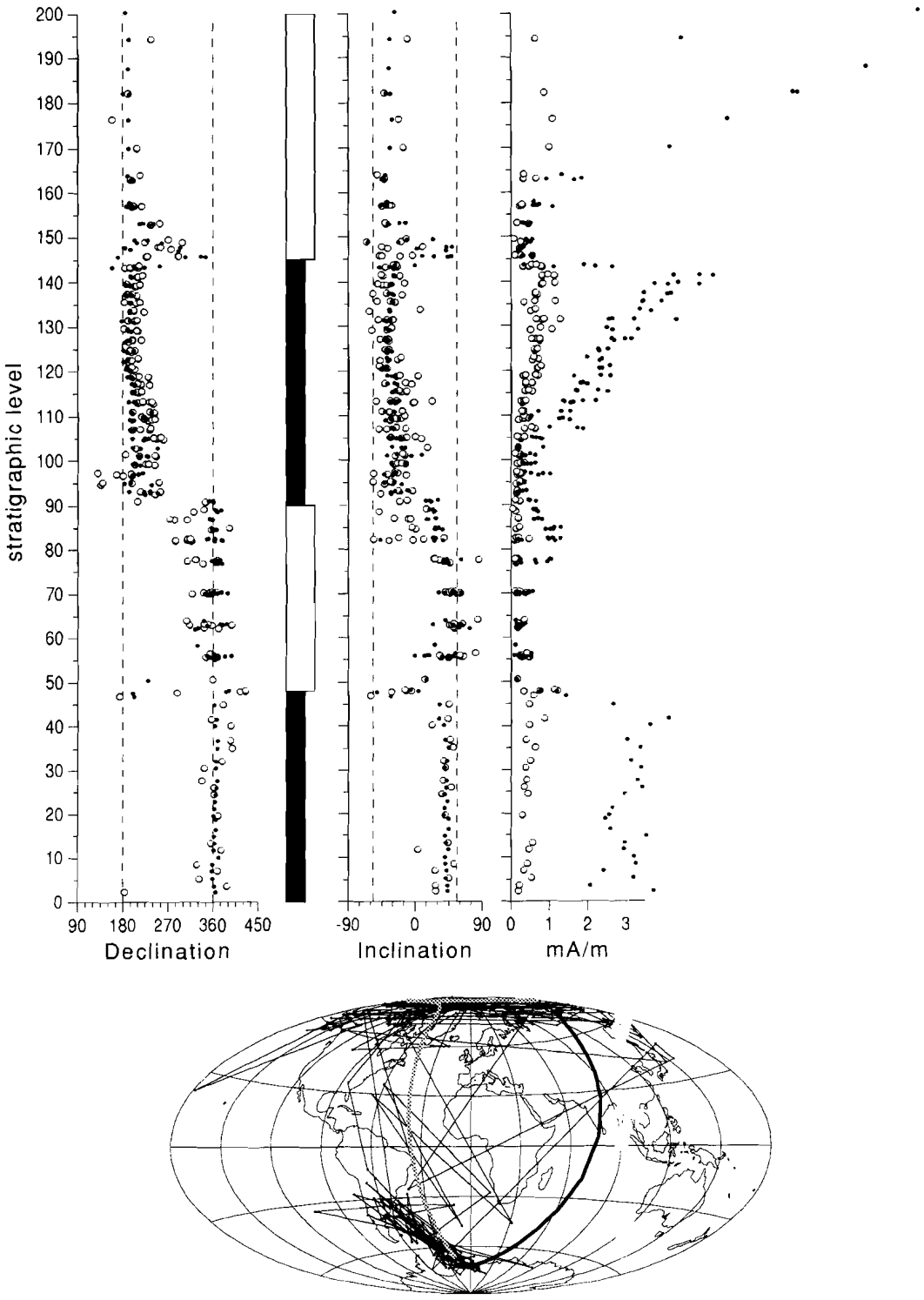


Figure 1: Upper Thvera record from Calabria (UTC) (Linssen, 1988, 1991). Closed (open) circles: HT (LT) component. Lithological column: black represents a grey layer, white a white layer. VGP paths: black (grey) line is smoothed non(near)-transitional directions. See text for discussion

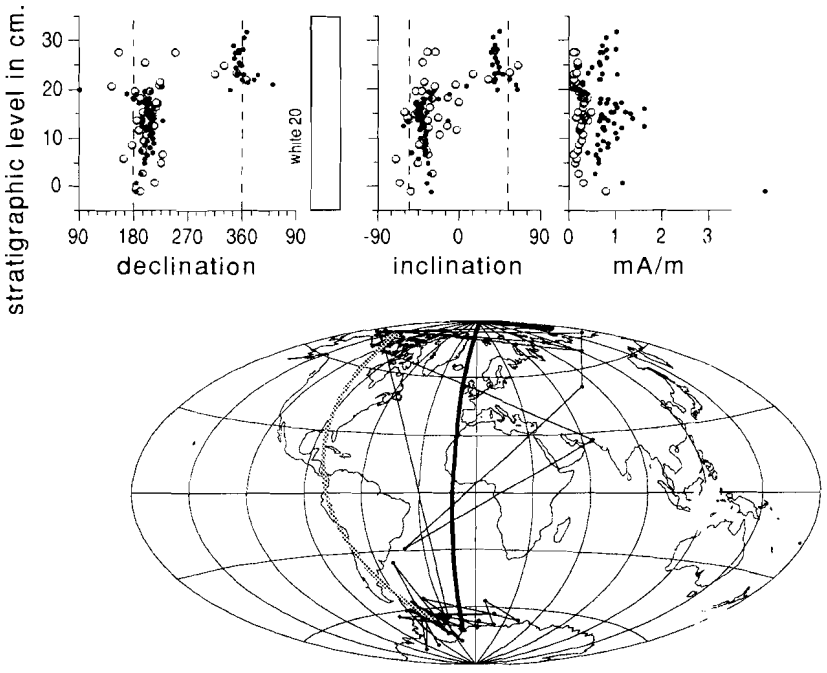
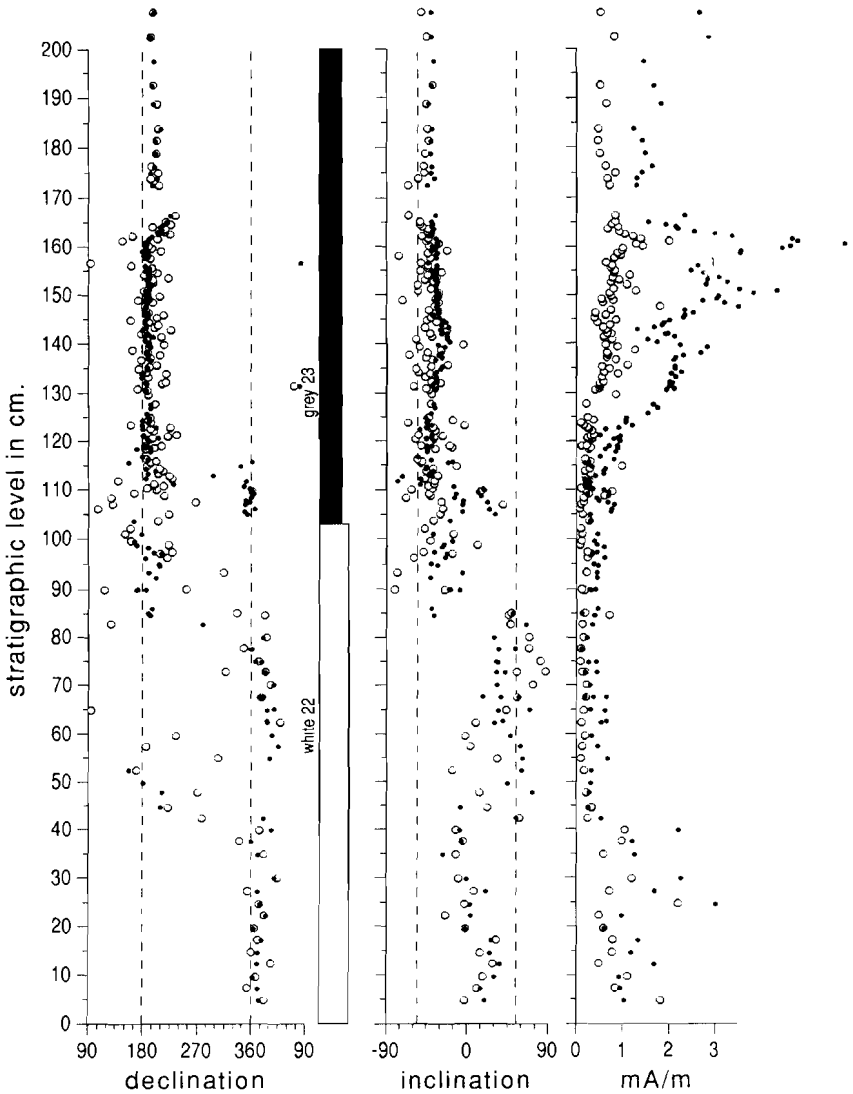


Figure 2: Lower Sidufjall record from Calabria (LSC) (Linssen, 1988, 1991). See caption to figure 1



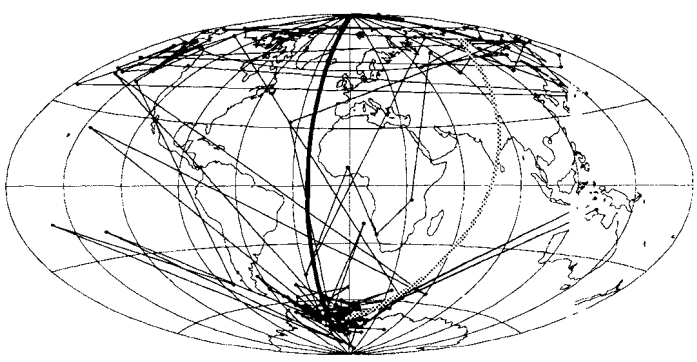


Figure 3: Upper Sidufjall record from Calabria (USC) (Linssen, 1988, 1991). See caption to figure 1

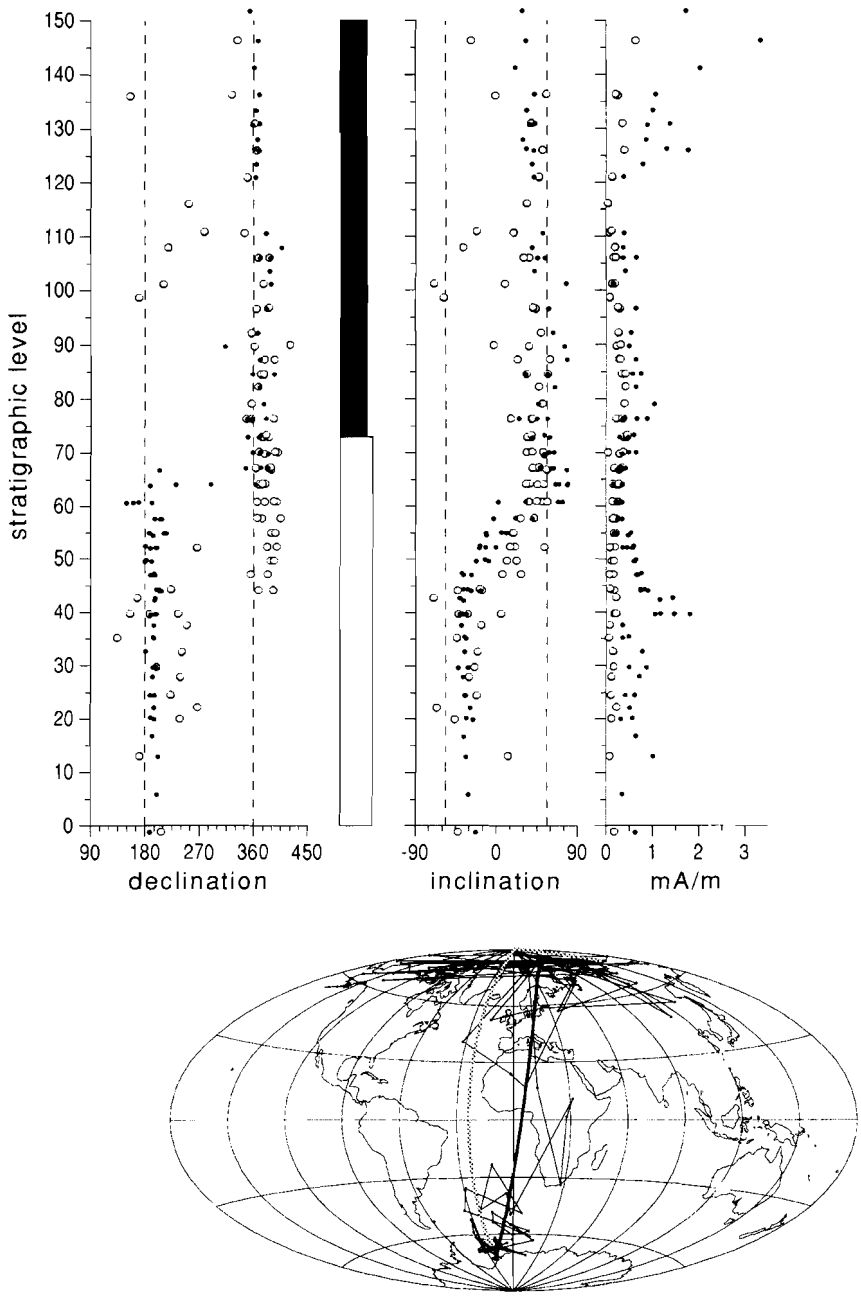


Figure 4: Lower Nunivak record from Calabria (LNC) (Linssen, 1988, 1991). See caption to figure 1

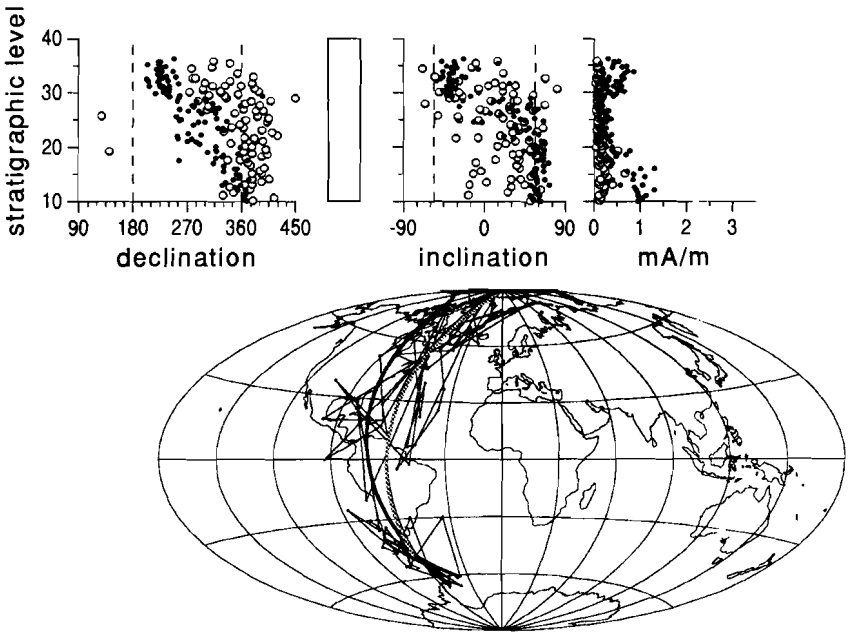


figure 5: Upper Nunivak record from Calabria (UNC) (Linssen, 1991). See caption to figure 1.

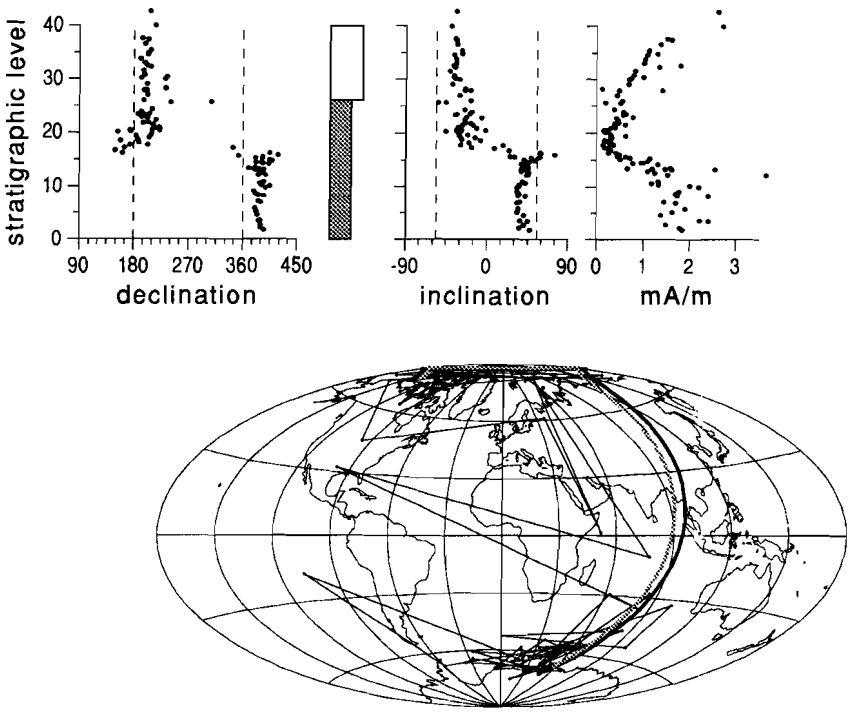


figure 6: Lower Mammoth record from Sicily (LMS) (Linssen, 1991). See caption to figure 1.

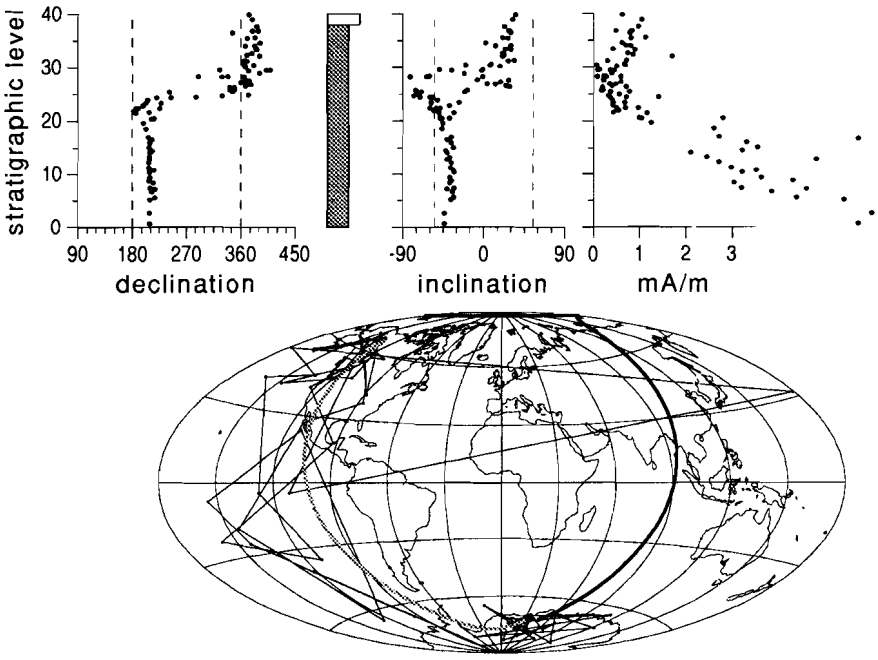


figure 7: Upper Mammoth record from Sicily (UMS) (Linssen, 1991). See caption to figure 1.

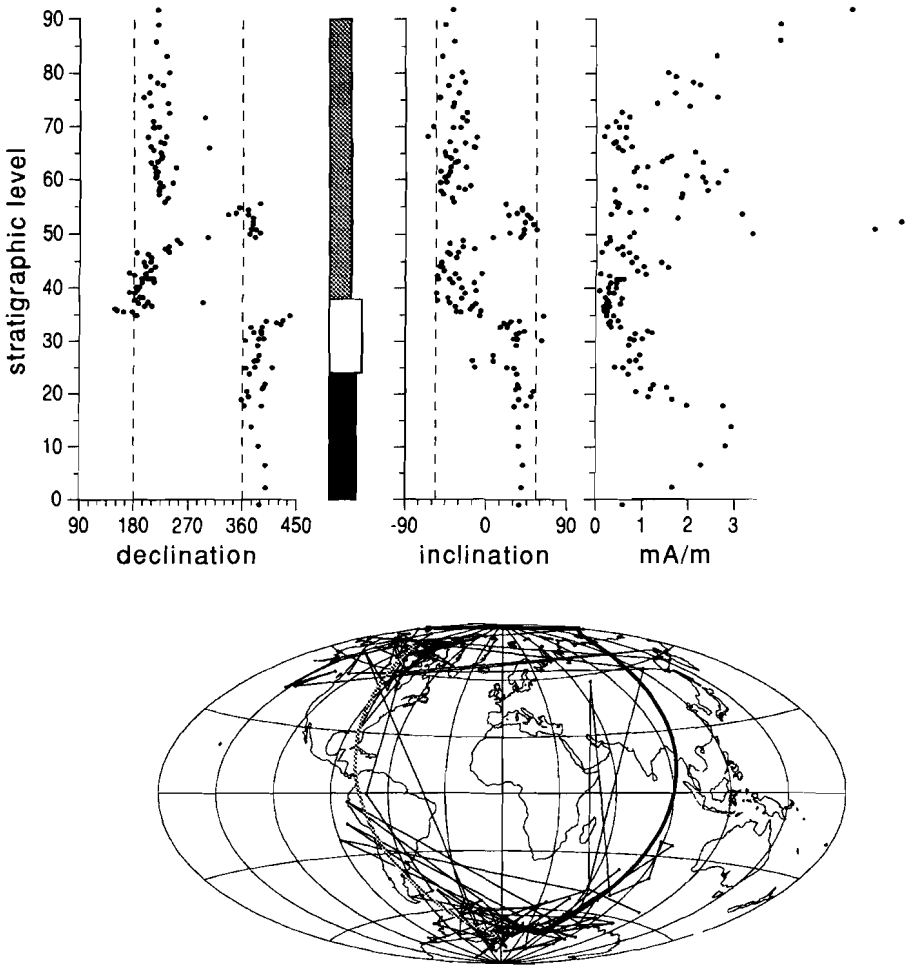


figure 8: Lower Kaena record from Sicily (LKS) (Linssen, 1991). See caption to figure 1.

Chapter 11

Eighteen geomagnetic polarity transitions recorded in southern Italy: a review

Introduction

The Pliocene polarity transitions in the Gilbert and Gauss Chrons, unambiguously identified by magnetostratigraphic studies in marine marls the Trubi and Narbone formations in Southern Italy, have been used to study the transitional behaviour of the geomagnetic field during polarity transitions. The sediments show a layering with a 21-22 kyr cyclicity, that can be correlated via orbital (climate) forcing to the precession of the Earth's rotational axis. The cycles range from number 1 at the Mio-Pliocene boundary up to number 119 at the Gauss/Matuyama boundary (Hilgen, 1991). On Sicily, a single cycle is expressed as a colour layering of grey-white$_{(1)}$-beige-white$_{(2)}$ marls, whereas in Calabria the white$_{(1)}$/beige/white$_{(2)}$ layering is represented by one single white layer. In many cases, it appeared that the study was not as straightforward as the magnetostratigraphic studies. The first order geomagnetic changes that are studied in magnetostratigraphy ('normal or reversed polarity') are very well recorded in sediment as is proven by the excellent fit of the NRM polarity zonation with the geomagnetic polarity time scale. The study of higher-order geomagnetic changes, necessary to investigate the morphology of geomagnetic polarity transitions, however, is a field full of pitfalls and traps. The remanence of the sediments under study consists of two major components: A low temperature (LT) component demagnetized between temperatures of 250° and 350° (sometimes up to 500° C), and a high temperature (HT) component demagnetised between 510° and 580-600°C. The remanence of these components is not acquired at a constant depth below the sediment/water interface but this depth may change under changing environmental circumstances (van Hoof and Langereis, 1991; chapter 1).

As a matter of course, a change in depth of acquisition during a geomagnetic polarity transition will introduce artefacts in the direction of the remanence. Another feature that may obscure the transitional geomagnetic information in the sediment is filtering of (geomagnetic) signal through process of NRM acquisition by the sediment. Indications for this mechanism were found by Langereis et al., (1992; chapter 2) who averaged the stable directions of the magnetostratigraphic subchronozones (mean *non*-transitional directions) and the stable directions just before/after the polarity transitions (mean *near*-transitional directions). The simple filtering of mean non-transitional directions as well as the mean near-transitional directions resulted in synthetic transitions with virtual geomagnetic pole (VGP) paths that show a striking correlation with the observed VGP paths.

Van Hoof et al. (1993; chapter 3) then developed a geochemically constrained model that tries to clarify the difficulties that were encountered during the study of the polarity transitions. This diagenetic magnetite model explains the directional changes as a function of the stratigraphical level, as well as the process of smoothing and changes in the lock-in depth of the remanence. In this chapter a review, of all 18 polarity transitions is given with an emphasis on the extent to which the data are consistent with the model. Finally, the possible geomagnetic information retained in these records is discussed.

18 transitions

The 18 reversal records have either been published earlier, or are (to be) submitted or in press; all are presented in this thesis (table 1). The data are presented as declination, inclination, intensity plots and as a virtual geomagnetic pole (VGP) path on an Aitoff world map projection. Ten of the Sicilian transitions (UTS to GGS, and UKS) have been presented in chapters 3 to 9 of this thesis, the data of five transitions from Calabria as well as the LMS, UMS and LKS transitions from Sicily were published by Linssen (1988; 1991) and are presented in chapter 10 of this thesis (table 1). In the VGP diagrams, the data from Sicily have been corrected for the 35° clockwise rotation of

transition		abbrev.	section	publication
Upper Nunivak	Calabria	UNC	Singa	Linssen, 1991, figure 5 chapter 10
Lower Nunivak	Calabria	LNC	Singa	Linssen, 1988; 1991, figure 4 chapter 10
Upper Sidufjall	Calabria	USC	Singa	Linssen, 1988; 1991, figure 3 chapter 10
Lower Sidufjall	Calabria	LSC	Singa	Linssen, 1988; 1991, figure 2 chapter 10
Upper Thvera	Calabria	UTC	Singa	Linssen, 1988; 1991, figure 1 chapter 10
Upper Kaena	Sicily	UKS	PP	van Hoof and Langereis, 1992 / chapter 9
Lower Kaena	Sicily	LKS	PP	Linssen, 1991, figure 8 chapter 10
Upper Mammoth	Sicily	UMS	PP	Linssen, 1991, figure 7 chapter 10
Lower Mammoth	Sicily	LMS	PP	Linssen, 1991, figure 6 chapter 10
Gilbert/Gauss	Sicily	GGS	CB	Van Hoof, 1993 / chapter 8
Upper Cochiti	Sicily	UCS	CB	van Hoof, this thesis / chapter 7
Lower Cochiti	Sicily	LCS	CB	van Hoof, this thesis / chapter 7
Upper Nunivak	Sicily	UNS	CB	van Hoof et al., 1993 / chapter 6
Lower Nunivak	Sicily	LNS	CB	van Hoof et al., 1993 / chapter 6
Upper Sidufjall	Sicily	USS	CB	van Hoof, this thesis / chapter 5
Lower Sidufjall	Sicily	LSS	CB	van Hoof, this thesis / chapter 5
Upper Thvera	Sicily	UTS	EM	van Hoof and Langereis, 1993 / chapter 4
Lower Thvera	Sicily	LTS	EM	van Hoof and Langereis, 1993 / chapter 4

Table 1: Transitional records in the chapters and/or references PP: Punta Piccola, B: Capo Bianco, EM: Eraclea Minoa

the sedimentary basin. With exception of the two Mammoth transitions and the lower Kaena transition, the HT as well as the LT component have been determined and are shown as black dots (HT) and circles (LT) in the declination, inclination and intensity plots. The LT component is not always used in the VGP diagrams because the this component is much more scattered than the HT components. Between the declination and inclination diagrams a lithological column is drawn, the different colours representing the white, beige and grey layers. In some columns layers with brown-coloured spots have been indicated if these were present. In some sections, the colour differentiation between the layers was less obvious. In that case, the colours in the lithological columns are also less pronounced.

Changes in direction, intensity and lithology.

The directional changes that are related to lithological boundaries are summarized in table 2. In total, 39 major directional changes have been observed in the records; 9 of them are clearly related to lithological boundaries, another 11 can most likely be related to such boundaries. 19 of the 39 major directional changes are not evidently related to changes in lithology. Apart from the changes in the directions, there are also changes in intensity without corresponding changes in directions. For example, in the LSS record (fig. 1, chapter 5), the intensity maximum is just below the top of grey 20, in the LCS (fig. 9, chapter 7) there is a decrease at the bottom of grey 43, and an increase above the top of the brown oxidised layer of white$_{(1)}$44. Similarly in the UTC record the intensity collapses at the onset of grey15 (fig. 1, chapter 10).

Smoothing

Tric et al et al. (1991) found that in many sedimentary records of polarity transitions the VGP paths were confined to a great circle passing over South and North America (or its antipode) and Laj et al. (1992) suggested that these bands of longitude were related to the pattern of fluid motion in the outer core and to higher seismic velocities in the lower mantle. Ten polarity transitions out of the group of 18 from this review also have their VGP paths clearly confined to one of these great circles, two paths show more or less this confinement (table 3). Langereis et al. (1992) have shown that this confinement may equally well be explained by smoothing related to the remanence acquisition mechanism of the sediment. The non-transitional directions are the average stable directions of the sub-chronozones resulting from the magnetostratigraphic studies of the sections.

UNC	N-R	-
LNC	R-N	-
USC	N-R	?
	R-N	+
	N-R	-
	excursion	-
LSC	R-N	-
UTC	excursion	+
	N-R	?
	excursion	+
UKS	excursion	?
	R-N	?
LKS	N-R	?
	R-N	-
	N-R	-
UMS	R-N	-
LMS	N-R	-
GGS	R-N	?
UCS	excursion	?
	N-R	?
	R-N	-
	N-R	-
	R-N	-
	N-R	+
LCS	excursion	-
	R-N	?
	N-R	?
	R-N	-
	excursion	+
UNS	N-R	-
LNS	R-N	+
USS	N-R	-
LSS	R-N	-
	excursion	+
UTS	N-R	?
	R-N	+
	N-R	-
	excursion	-
	excursion	+
LTS	R-N	-

table 2: Correlation of directional changes with lithological boundaries. N: normal, R: reversed, +: good correlation, ?: probable correlation, -: no evident correlation.

The near-transitional directions are the average stable directions just before and after the intermediate directions in the transitional records. Synthetic VGP paths were produced by smoothing both the non-transitional and near-transitional directions. In the VGP projections also the synthetic non- and near-transitional VGP paths shown in black and grey, respectively. The observed VGP paths are then compared with the synthetic paths. In the case of the USS record (fig. 3, chapter 5), no synthetic near-transitional path could be constructed because already in the lowermost part of the record there appear to be only intermediate directions. In the LNS (fig. 8, chapter 6) and the LSC records (fig. 2, chapter 10) too few intermediate directions are present to make a comparison with the synthetic paths. Comparison of the observations (table 4) with the non-transitional paths shows that from the 18 polarity transitions, two are undetermined, four do not fit the synthetic paths, five have a fair fit, and seven show a good fit. Comparison with the near-transitional data result in 3 out of 18 are undeterminate, one does not fit the data, four show a fair fit and 10 show a good or excellent fit.

The length of the transition in the sediment is a measure of the width of the smoothing window. The lengths of trajectories with intermediate directions in the record of the LNS is less than 5 cm and in the subsequent UNS transition it is some 15 cm (fig. 8, chapter 6). Probably, the width of the filter between the LNS record abruptly changes at the lithological boundary of the transition.

	cluster	path
UNC	(SA)	+
LNC	-	-
USC	-	-
LSC	-	-
UTC	SA	+/-
UKS	A, NA	+
LKS	-	+
UMS	-	+
LMS	-	-
GGS	NA	+
UCS	-	-
LCS	-	+
UNS	SA,(NA)	+
LNS	-	-
USS	SA	+/-
LSS	NA	+
UTS	SA	+
LTS	SA,NA	+

Table 3: Coincidence of VGP clusters with the inclined dipole clusters (Hoffman, 1992) and VGP paths with longitudinal bands over these clusters. NA: North America, SA: South America, A: Australia. (A) plus (+) denotes that the observed VGP path is confined to one of the great circles found by Tric et al. (1991). (NA), (SA), +/- denotes that the cluster or the confinement is not clear.

If the geomagnetic signal is smoothed with a narrow filter then some of the information of the geomagnetic field will be preserved. Hoffman (1992), in a study of several volcanic records of polarity transitions from the last 10 Myr, speculated that there are long-lived intermediate inclined dipolar stages with VGP

positions that cluster on South America and close to Australia and, more speculative, in the northern Atlantic close to North America. These clusters happen to be localised on the same longitudinal bands as found by Laj et al. (1992) in their study to the confinement of the VGP paths of sedimentary transition records. If these clusters are indeed characteristic for the geomagnetic transitional field, then in sedimentary records clusters of VGPs should be observed at these same locations. If the filter width is too large with respect to the duration of the inclined dipolar stages, the VGP path should only pass the location of the cluster, whereas if the duration of these long-lived phenomenon is longer than the filter width, it will be seen even in the case of extended smoothing. From the 18 polarity transitions of the present study, six transitions have VGP clusters on one or two locations as found by Hoffman and 11 transitions have their VGP paths over one or two of the clusters (table 3). Although this is a reasonable correlation, we still feel that most of the clusters in our sedimentary records may equally be caused by sedimentary artefacts as by true transitional geomagnetic field behaviour.

	non	near
UNC	good	good
LNC	good	good
USC	fair	bad
LSC	undet	undet
UTC	bad	fair
UKS	good	good
LKS	fair	fair
UMS	bad	good
LMS	fair	fair
GGS	good	good
UCS	fair	fair
LCS	bad	good
UNS	bad	good
LNS	undet	undet
USS	fair	undet
LSS	good	good
UTS	good	good
LTS	good	good

Table 4: Correlation of the observed VGP paths with the synthetic paths from the smothed non transitional directions (non) and the smoothed near transitional directions (near). Undetermined (undet) means that too few intermediate VGPs were present to determine the correlation.

Diagenetic magnetite model

The diagenetic magnetite model (Van Hoof et al., 1993; chapter 3) was developed to explain the directional changes observed in the remanence of sediments. It is partly based on geochemical measurements and its describes the redox conditions in the sediment shortly after deposition, and the timing of the formation of the remanence bearing magnetites. In this model, the grey layers always carry the primary magnetites with a remanence that is acquired very shortly after deposition. The white and beige layers also contain the primary magnetites but superposed on it, there may be secondary magnetites that have acquired their remanences several thousands of years after deposition. These secondary magnetites are formed in a moving front of oxidation that is initiated at the bottom of a beige/white sequence and/or at the top of a white/beige sequence. Migration of iron takes place from the grey layers, depleting the iron content of these layers. If a geomagnetic transition takes place between the formation of the primary and secondary magnetites, the remanences of the two magnetites have different directions. If the remanences of the primary and secondary magnetite have a comparative intensity, the sum vector may have an intermediate direction, unless there is a distinction between their respective unblocking temperature spectra. This model was used to explain the observations in one transition record (Van Hoof et al., 1993; chapter 3). The validity of the model is now tested in the other transition records.

The diagenetic magnetite model predicts that the white and beige layers above a grey layer with the post-transitional direction should not show directional changes, but three records contradict this prediction.

In the UTS record (chapter 4, fig. 10, page 31), the directional changes are recorded in the uppermost white layer at level 125-130 cm. The top of the grey layer below this swing has clearly a post-transitional direction, while the grey layer above the UTS section has also post-transitional directions (Hilgen and Langereis, 1988). According to the model, the layers in between two grey layers with (primary) post transitional directions should also have post-transitional directions and the directional swing in the uppermost white layer is therefore in contradiction with the model. Van Hoof and Langereis (1992) related the uppermost directional changes in both upper Thvera records to a present-day field overprint, since in the UTS record the swing coincided with a particular

brown layer with rock magnetic parameters that indicated weathering products. But this brown layer is of primary (near-depositional) origin as is evident from other records. Hence, in the UTS record the topmost grey layer has recorded secondary magnetite directions.

In the LCS record (fig. 9, chapter 7, page 78) the anomalous swing in the beige layer at level 140 can be explained by the model only if the restriction that the oxidation front cannot penetrate the grey layers is slightly adjusted. In that case, the top part of the middle grey layer has post-transitional directions due to secondary magnetites and the gradual polarity transition recorded in this layer is caused by a gradual decrease in secondary magnetite formation. The results of several other records also indicate that the conclusion from the model - the remanence of the grey layers shows the real geomagnetic field - needs some adjustment. For instance: in the UTC record (fig 1, chapter 10, page 110) at stratigraphically the same level (topmost grey to topmost white, above level 145 cm) as in the transition from the UTS record, level 140 cm (fig. 10, chapter 4 page 31) a similar directional change is observed, and this is also, on the same grounds as above, in contradiction with the model.

The upper Thvera transition is in Calabria (UTC, fig 1, chapter 10, page 110, between levels 85 and 95 cm) as well as in Sicily (UTS, fig 10, chapter 4, between levels 85 and 105 cm) recorded in stratigraphically the same grey layer. If the grey layers contain only primary magnetites, then the morphologies of both transition records should be the same. But they are not.

A similar reasoning can be used if the upper Sidufjall records from Sicily and Calabria are compared: the directional changes in the grey layer of the USC record (fig 3) are not observed in the USS record. Apparently, the present early diagenetic magnetite model needs some refinement, making it slightly more complex than the first version presented in chapter 3.

Offsets

We applied the mean *near*-transitional directions to the reversal tests which were published by McFadden and McElhinny (1990). All mean *near*-transitional directions have negative reversal tests, with the exception of the LKS transition, which has a class B reversal test. Similarly all mean *non*-transitional directions have negative reversal tests, with the exception of the subchronozones enclosing the LTS, UMS and UTC polarity transitions (table 5). The diagenetic magnetite model can account for differences between the two localities on Sicily and Calabria over (a maximum of) two sedimentary cycles, but not for differences between entire subchronozes, as is observed in the parallel sections (table 5). An overprint by the present-day field direction (with an inclination of 52°) may have caused the observed offset. In order to investigate this possibility we have subtracted a present day field direction from the mean *non*-transitional directions. The present-day field direction was corrected for the bedding tilt of each section. The corrected present day field direction was stepwise subtracted from the reversed and normal mean non-transitional directions before and after each transition. After each step the angle between the reversed and normal directions was calculated. This subtraction continued until the angle between the mean normal and mean reversed direction was at a maximum. In the majority of the cases this leads to a decrease of the offset between the N and R directions. In the most ideal case this maximum is 180°. After this subtraction procedure, again the reversal tests were applied. This resulted in positive reversal tests of the mean non-transitional directions in all sections, except the Punta Piccola section and the non-transitional directions enclosing UNS, LCS and UCS transitions. Nevertheless, also in the cases of UNS, LCS and UCS the non-antipodality angle γ is decreased (table 6). This suggests that the offsets could well be due to a present-day field overprint. Contrary to these sections the mean non-transitional directions of the PP section would require subtraction of a reversed overprint to decrease the non-antipodality angle (γ). This is quite unlikely, suggesting that at least here stable directions are truly non-antipodal. Investigation of (Sicilian) volcanics from the Gauss subchron are needed to determine more reliably whether this curious offset in this chron is due to the geomagnetic field or due to an unlikely reversed overprint.

	PRE				POST				reversaltest		
	D	I	N	Rsum	D	I	N	Rsum	γ	γ*	Class
UNC	11.4	47.2	29	28.811	205.5	-43.8	16	15.128	10.4	9.4	neg
LNC	191.7	-43.1	20	19.888	11.4	47.2	29	28.811	4.1	3.2	neg
USC	7.4	50.1	14	13.782	191.7	-43.1	20	19.888	7.6	5.8	neg
LSC	190.4	-40.1	3	2.991	7.4	50.1	14	13.782	10.2	8.3	neg
UTC	13.7	42	12	11.696	190.4	-40.1	3	2.991	3.1	9.9	C
UKS	203.9	-37.6	12	11.950	34.5	42.2	32	31.750	9.3	3.6	neg
LKS	34.7	41.4	8	7.860	203.9	-37.6	12	11.950	9.2	8.2	neg
UMS	204.2	-37.9	9	8.922	34.7	41.4	8	7.860	8.8	9.2	B
LMS	32.7	40.1	19	18.923	204.2	-37.9	9	8.922	7.0	5.7	neg
GGS	219.2	-39.4	49	48.280	32.7	40.1	19	18.920	5.0	3.2	neg
UCS	29.5	49.6	8	7.950	219.2	-39.4	49	48.280	12.3	5.9	neg
LCS	214.7	-41	17	16.780	29.5	49.6	8	7.950	9.3	6.0	neg
UNS	31.5	50.9	12	11.930	214.7	-41	17	16.780	10.1	5.2	neg
LNS	219.8	-42.5	13	12.571	31.5	50.9	12	11.930	10.4	8.5	neg
USS	25.8	45.3	4	3.980	223.2	-44.9	13	12.786	12.3	8.7	neg
LSS	222.1	-41.4	7	6.930	25.8	45.3	4	3.980	12.5	8.6	neg
UTS	30.4	47.4	11	10.860	222.1	-41.4	7	6.930	10.3	8.0	neg
LTS	221.5	-50.3	4	3.960	30.4	47.4	11	10.860	7.8	10.7	C

Table 5. Average directions of the magnetostratigraphic subchronozones (non-transitional directions) before (PRE) and after (POST) the polarity transitions. D= declination, I= inclination, N= number of data, Rsum= unit vector sums. γ=angle between N and R, γ=critical angle, Class=results of the reversal test, (C=marginal, B=reasonable, A=good, neg=negative (McFadden and McElhinny, 1990)*

	PRE				POST				reversaltest		
	D	I	N	Rsum	D	I	N	Rsum	γ	γ*	Class
UNC	13.8	43.6	29	28.811	201.7	-46.9	16	15.128	6.5	9.4	B
LNC	190.8	-44.9	20	19.888	12.6	45.4	29	28.811	1.4	3.2	A
USC	9.4	46.5	14	13.782	189.8	-46.7	20	19.888	0.3	5.8	B
LSC	188.6	45	3	2.991	10	45.3	14	13.782	1.0	8.3	B
UTC	14	41.5	12	11.696	190.2	40.6	3	2.991	3.1	9.9	B
UKS	207.0	-36.1	12	11.950	30.6	42.8	32	31.750	8.1	3.6	neg
LKS	30.6	42.1	8	7.860	207.2	-36.0	12	11.950	6.6	8.2	B
UMS	206.7	-36.3	9	8.922	29.7	41.0	8	7.860	6.7	9.2	B
LMS	29.7	41.0	19	18.923	206.7	-36.6	9	8.922	5.0	5.7	B
GGS	217.1	40.7	49	48.280	34.5	38.6	19	18.920	2.9	3.2	A
UCS	37.1	48.5	8	7.950	213.8	-42.8	49	48.280	6.1	5.9	neg
LCS	211.1	-43.2	17	16.780	34.6	48.9	8	7.950	6.2	6.0	neg
UNS	35.8	50.7	12	11.930	211.8	-42.8	17	16.780	8.4	5.2	neg
LNS	215.1	-44.6	13	12.571	37.6	50.5	12	11.930	6.1	8.5	B
USS	34.5	46.2	4	3.980	214.5	-44.9	13	12.786	1.3	8.7	B
LSS	214.3	-42	7	6.930	34.1	46.2	4	3.980	4.2	8.6	B
UTS	36.4	48.1	11	10.860	216.7	-41.9	7	6.930	6.2	8.0	B
LTS	215.6	-49.7	4	3.960	36.1	48	11	10.860	1.7	10.7	C

table 6. Average directions of the magnetostratigraphic subchronozones (non transitional directions) before (PRE) and after (POST) the polarity transitions after correction for a present day field overprint corrected for the bedding plane. The directions from the Punta Piccola section (LMS to UKS) have been corrected with a reversed overprint. See caption to table 5.

Conclusion

The first-order geomagnetic signal (polarity) has been correctly recorded in the marls of Southern Sicily, as is shown in magnetostratigraphic studies. But in the investigated sedimentary records of polarity transitions, the higher-order geomagnetic changes are to a large extent obscured. Many changes in direction and in intensity observed in these records are strongly related to lithological boundaries. In many cases transitional fields can be explained by a complex filtering process due to early diagenetic magnetite formation. The width (length or duration) and the morphology of the filter(s) is highly variable. Some information of the transitional field, like longi-

tudinal confinement of VGPs (Laj et al., 1992) and/or the long-lived VGP clusters (Hoffman, 1992) is probably still present in the records. The preliminary diagenetic magnetite formation model cannot explain all observations, and requires the adjustment of some of its conclusions. The mean directions of the subchronozones are found not to be antipodal. The difference in offset in the parallel sections from Sicily and Calabria suggests, however, that these offsets are not caused by the geomagnetic field. It can be quantitatively shown that the offset is probably caused by a present-day overprint in all sections, except in Punta Piccola. If the offsets from this section is caused by an overprint, the direction of the overprint must have been reversed, which is unlikely.

Although the reliability of sediments to record polarity transitions is certainly far from ideal, some important geomagnetic information may still be present in the sediments. For instance, there seems to be evidence for inclined dipole directions (Hoffman, 1992) in our records. If true, the nature of the filtering process suggests that these are long-lived features of the transitional field. Parallel records from both volcanics and sediments must provide the key to solve the problem of remanence acquisition. Another very interesting question that emerges from this study is the reason of the anomalous offsets of the mean stable directions in the Gauss Chron. Also in this case, a paleomagnetic study in volcanic rocks should give a more definite answer.

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Curriculum Vitae

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