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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

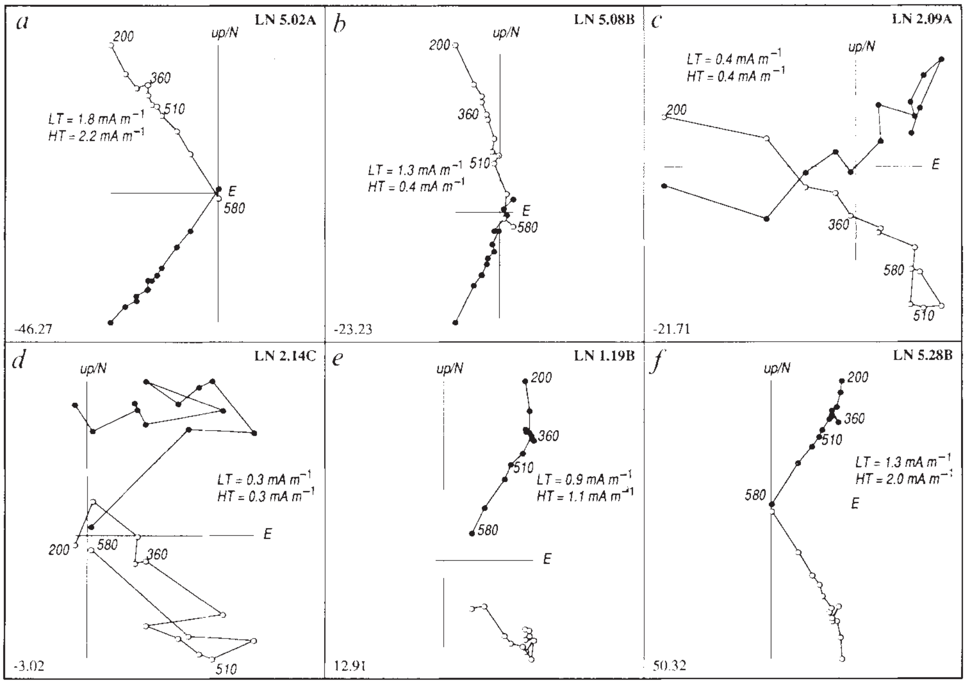

FIG. 1 Representative thermal demagnetization diagrams of the lower Nunivak (LN) reversal. Stratigraphic level in cm (left-hand axis) refers to the stratigraphic column of Fig. 2. Solid (open) symbols are horizontal (vertical) projections. Temperature steps below 200 °C are not shown in order to enhance 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.

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 reversed and the LT component is normal, the latter is easily interpreted to represent a (partial) overprint by secondary (subrecent) 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.

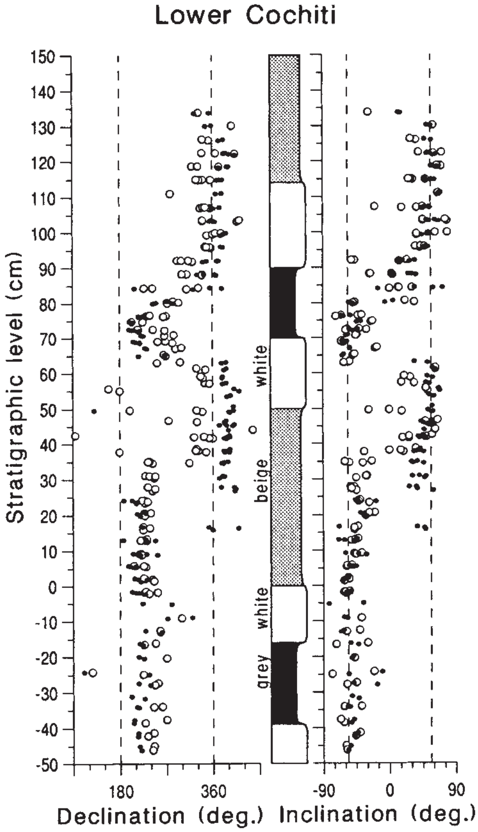

FIG. 3 Declination and inclinations during the lower Cochiti reversal. See also Fig. 2 legend. The record shows three 'polarity reversals', respectively R–N, N–R and N–R, in both the HT and LT directions.

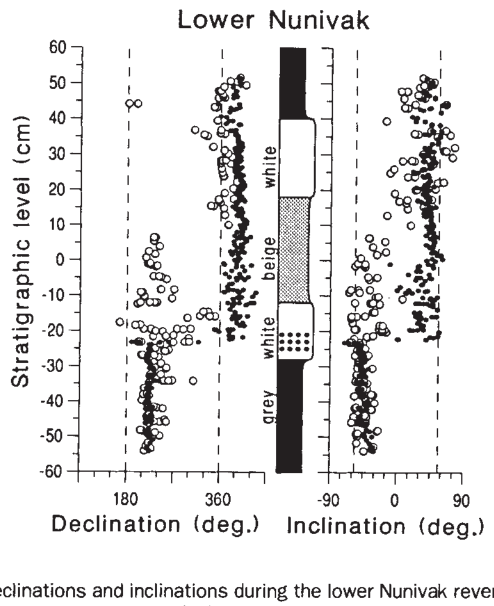

FIG. 2 Declinations and inclinations during the lower Nunivak reversal. Open (closed) symbols denote LT (HT) directions. The LT component reverses polarity ~35 cm higher than the HT component and shows more scatter. The weathering colours (carbonate contents[4]) of each sedimentary cycle are grey (70%), 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.

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 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 magnetomineralogy 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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New early Cambrian animal and onychophoran affinities of enigmatic metazoans

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THERE is much interest in the early evolution of metazoans with the restudy of the Middle Cambrian 'soft-bodied' fauna of the Burgess Shale[1]. Several other, newly discovered Cambrian 'soft-bodied' faunas[2] provide a wealth of new data. One of the oldest and best-preserved faunas was discovered in 1984 in Chengjiang in southern China[3]. This fauna is of early Cambrian age, about late Atdabanian[4] (~520–530 Myr BP)[2]. We now describe a new 'armoured lobopod' from the Chengjiang fauna. This animal shows close affinity with the enigmatic *Microdictyon*[5]. The conundrum *Hallucigenia*[6] is reinterpreted as another 'armoured lobopod', as are *Xenusion*[7] and *Luolishania*[8]. The large plates set in pairs along the trunk are a synapomorphy of this group, which flourished soon after the 'Cambrian explosion'. Soft-part anatomy suggests that the group has affinities with the Burgess Shale 'lobopod' *Aysheaia*[9]. All these marine, Cambrian forms are here grouped with the extant, terrestrial velvet worms in the phylum Onychophora.

The 'soft-bodied' Chengjiang fauna in the early Cambrian Chiungchussu (Qiongzhusi) Formation occurs at several localities around Kunming in Yunnan, People's Republic of China[10,11]. It is recognized as the most important early Phanerozoic *Lagerstätte* beside the Burgess Shale Phyllopod bed[12,13]. Over 3,000 remarkably preserved, 'soft-bodied' specimens collected to date have revealed a diverse fauna, now partly under study in a Chinese-Swedish project.

In 1989, two laterally compressed specimens of a caterpillar-like animal (Fig. 1) were collected. The 5–6-cm-long trunk bears 11 pairs of legs, above which are rounded, outwardly facing plates. These immediately indicate affinity with the Lower Cambrian problematicum *Microdictyon*[14]. Plates of this form known from isolated phosphatic material, were recently found pair-wise above the legs of a worm-like Chengjiang animal[5] (Fig. 3*b*). Despite this discovery, the affinities of *Microdictyon* have remained enigmatic[11,12,15–19]. We regard the paired plates, not noted elsewhere in the animal kingdom, as homologous in the two taxa.

The rounded head of the new form is poorly differentiated from the trunk. There are several thin, elongated structures anteroventrally. At least four of these radiate from a central void, giving the impression of papillae surrounding a mouth. Two elongate structures may be antennae. Three distinct annuli separate neighbouring plate- and leg-bearing segments. Each annulus has a vertical row of about four 'tubercles'. Long, tentacle-like papillae protrude dorsally from the trunk and head. Some annuli carry two such papillae, bilaterally placed, possibly indicating two longitudinal rows. Long, finely annulated protrusions extend ventrally from the trunk, two or possibly more per annulus. The distinctly annulated legs carry large, paired and curved, three-dimensionally preserved claws (Fig. 3*a*). The trunk merges smoothly with the last leg pair, other legs attach at a circular area beneath each plate. A longitudinal dark band centrally in the trunk apparently represents the gut.

The 10 plate pairs gradually increase in size from the anterior to the fifth pair. Each plate is rounded, dorsoventrally elongate, convex outward, with a ridge near its edge, and a surface pattern of fine pits and nodes. The posteriormost plates are more subrectangular. Each plate is drawn out dorsocentrally (Fig. 2) into a broad-based, pointed spine, directed obliquely upwards and out.