QUATERNARY DATING METHODS – A USER'S GUIDE

Edited by
P.L. SMART & P.D. FRANCES

Quaternary Research Association

Technical Guide No. 4

CONTENTS

Chapter 9

PALAEOMAGNETIC DATING

R. Thompson

> The key that would eventually unlock the chronology of the Pleistocene (was) found in 1906 in a French Brickyard by Bernard Brunhes, a geophysicist investigating the earth's magnetic field.
>
> *Imbrie and Imbrie (1979).*

INTRODUCTION

Palaeomagnetic stratigraphy involves the measurement of the natural remanent magnetism of sediments or rocks and then the matching or correlating of the measured remanences with other previously dated palaeomagnetic records or geomagnetic field behaviour. The most important remanence changes for Quaternary studies are those associated with polarity reversals of the geomagnetic field.

The Earth's magnetic field is presently of normal polarity while Jupiter's and Saturn's fields are of reversed polarity. The north seeking end of a compass needle on Jupiter and on Saturn would point south rather than to the north as here on Earth. Around 730 ka, the Earth's magnetic field switched from also pointing South, from a state of reversed polarity, to its present state of normal polarity. Seven such polarity switches occurred on Earth during the Pleistocene.

Figure 9.1 illustrates the Pleistocene polarity time scale with its four normal polarity subchrons, four reversed polarity subchrons and seven polarity transitions. The most recent of the polarity switches is referred to as the Matuyama-Brunhes boundary, having been named in honour of two pioneering workers in the study of polarity reversals. The Olduvai, Jaramillo, Matuyama-Brunhes boundary sequence of Pleistocene geomagnetic polarity markers has been recognised in a range of remarkably diverse geological situations including continental lava flows, deep sea sediment sequences, lake sediments, European and Asian loess deposits and sea floor magnetic anomalies from all of the world's oceans. The Cobb Mountain polarity event which took place 1.10 Ma ago is a recently established geomagnetic feature (Mankinen et al., 1978) that only lasted for about 10 ka. The Pleistocene polarity

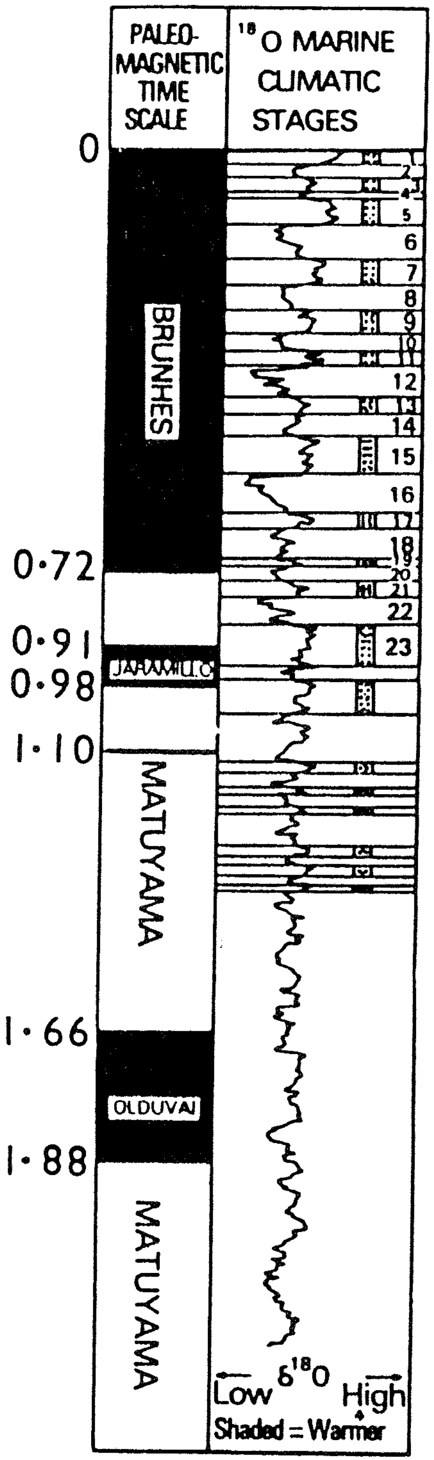

Figure 9.1 *Pleistocene time scales. The polarity time scale after Mankinen and Dalrymple (1979) and Mankinen and Grommé (1982). The 18O record after Shackleton and Opdyke (1976). (Modified from Harland et al., 1982.)*

timescale has been dated through potassium-argon age determinations on some 170 lavas (Mankinen and Dalrymple, 1979, McDougall, 1979). It is unlikely that any further long lasting (more than 20 ka) Pleistocene polarity changes remain to be discovered.

The geomagnetic field varies on all time scales from short pulsations lasting a fraction of a second, through secular changes varying in length from months to hundreds of years, on through large secular changes and aborted polarity reversals taking thousands of years, through full polarity reversals to changes in the average frequency and polarity bias of reversals which take place over periods of hundreds of millions of years. While Pleistocene geomagnetic fluctuations of shorter duration than 10 ka have certainly taken place and while they are of undoubted geomagnetic interest and of potential chronological value, they have proved to be extremely difficult to document unequivocally. Consequently the value of short duration geomagnetic changes for dating purposes is extremely doubtful.

One type of temporary field change falling into this category is commonly referred to as an excursion. Although scores of excursions have been inferred from palaeomagnetic data, duplication between neighbouring sites or real synchronicity have rarely been demonstrated. Perhaps two examples of the likely recognition of Brunhes age excursions, or extremely short reversals, in igneous rocks are the Emperor and Laschamp/Maelifell palaeomagnetic features with ages of around 490 ka and 47 ka respectively (Champion et al., 1981, Bonhommet and Babkine, 1967, Levi et al., 1990).

Secular magnetic direction changes with amplitudes of a few tens of degrees and lifetimes of a few hundred years have been recorded by direct observation from 1500 AD onwards (e.g. Bauer, 1896, Thompson, 1983). They have also been documented in certain regions through archaeomagnetic studies (e.g. Aitken, 1974, Hirooka, 1971). Secular magnetic variations may have some correlation or dating significance at these shorter time scales.

Palaeomagnetic remanence intensities do not directly reflect ancient geomagnetic field intensities, but are predominantly related to magnetic mineralogy, grain size and iron oxide concentration. Nevertheless by using careful, controlled laboratory magnetization techniques, to correct for mineral variations, reliable palaeointensity determinations have been obtained from thermoremanences (e.g. Aitken, 1974). However, attempts to use the depositional remanence of sediments have failed to produce clearly repeatable palaeointensity results. This is because little progress in developing laboratory methods for correcting sedimentological modulations of palaeomagnetic remanence intensity has been made since the pioneering sediment magnetism studies of Johnson et al. (1948) and Ising (1942) half a century ago. In practice, palaeointensity studies have yet to be shown to form the basis of a viable dating method.

The main value of palaeomagnetic studies in Pleistocene dating undoubtedly lies with the polarity reversal time scale of Figure 9.1. The Quaternary reversal time scale has been gainfully employed in many varied and important branches of the Earth sciences. These include the dating of deep sea and lacustrine cores for sedimentological and climatic studies (Opdyke, 1972, Singh et al, 1981), the determination of rates and directions of sea floor spreading, Vine (1966), providing age information about volcanic rock sequences (Einarsson, 1957) and linking the terrestrial and marine climatic records (Kukla, 1970).

THE PALAEOMAGNETIC METHOD

The palaeomagnetic method involves:- (1) collecting orientated samples, (2) determining the direction of any remanent magnetization held by the samples, (3) confirming the stability of the remanence, (4) checking that the measured remanences relate to the geomagnetic field direction at sample formation time and (5) matching the measured remanences with a previously-dated pattern of geomagnetic behaviour. The first three steps of collection, measurement and partial demagnetization are relatively straightforward. The fourth step, in particular, is a vital part of the palaeomagnetic method and is to be ignored only at peril.

Collection

The more sample orientation data obtainable, the safer the palaeomagnetic method will be. Complete orientation information of way-up and azimuth is best, although often with core materials way-up alone or just relative azimuth have to suffice. Way-up allows magnetic inclination to be measured while azimuth allows declination to be measured. Sediment cores can either be sub-sampled while fresh using small (approximately 10 ml volume), thin-walled plastic boxes or sawn into suitably sized samples if dry. Sediment sections can also be sub-sampled with plastic boxes, using a clean vertical face. Igneous rocks can be sampled either by collecting blocks some 10 to 20 cm across, or else by drilling cores, some 25 mm in diameter and several centimetres in length, using a portable rock drill with a diamond impregnated bit. Full orientation information can be obtained for a block by noting the strike and dip of a convenient flat face and for a core by noting the direction and dip of its long axis and its way-up.

There are no fixed rules about which sediment and igneous rock types yield useful palaeomagnetic information. Most lavas carry a usable remanence whereas most sediments do not. Less than one sediment sequence in five turns out to be amenable to palaeomagnetic dating (Stupavsky and Gravenor, 1984, Thompson and Oldfield, 1986). In general, any sediments that were deposited in high energy environments, have been distorted since deposition, are entirely minerogenic, are almost completely organic or contain an appreciable coarse silt or sand fraction are very unlikely to yield dependable palaeomagnetic results. Eminently usable palaeomagnetic data have

been obtained, however, from a whole range of materials including deep sea clays, loess deposits, moderately organic lake sediments, red siltstones and sandstones, basalts and andesites. The most persistently troublesome materials have been minerogenic varves and shallow marine deposits.

Measurement

Easy to use, robust, portable, sensitive, fluxgate magnetometers are to be found in dozens of laboratories around the world. These instruments are ideal for Pleistocene palaeomagnetic work. Their operation can be mastered in seconds and samples measured at rates of around twenty to sixty per hour. The more specialised superconducting magnetometers can also be employed. They have recently become very reliable and easy to use, and can be extremely valuable for specialised studies of small or very weakly magnetized samples. Remanence directions can be measured to within a couple of degrees with most types of magnetometer, so that instrumental errors are rarely of concern in palaeomagnetic studies.

Cleaning

The natural remanence (NRM) of rocks and sediments may be made up of several magnetic components lying in different directions. These components can be any combination of primary remanence plus secondary chemical (CRM), viscous (VRM), partial thermal (PTRM) or isothermal (IRM) remanences. Alternating field and thermal partial demagnetization or "magnetic cleaning" techniques have been developed in order to separate out the more stable magnetizations in multicomponent remanences. In many cases these laboratory based treatments have been found to work extremely well and allow stable, characteristic remanences (ChRM) to be isolated. Usually, the most stable component distinguished is taken to be the primary remanence direction, reflecting the geomagnetic field direction at the time of origin of the material under investigation. With Brunhes age sediments, storage in zero field for a few days often provides an effective cleaning method, but for Matuyama or older materials, active demagnetization methods are most desirable. Demagnetization studies should form an integral part of any palaeomagnetic study.

Checking

Although there is no single, simple, definitive method of proving that a palaeomagnetic remanence is a true record of the ancient field, there are a variety of tests and criteria that can demonstrate the converse — ie. that palaeomagnetic data are unreliable. Irving (1964) succinctly summarizes the use of minimum criteria of reliability in palaeomagnetic studies. Thompson (1984) lists ten reliability criteria for magnetostratigraphic, secular variation studies of sediments.

Igneous Rocks The main problem with igneous samples concerns the effect of self reversal, in which the thermoremanence acquired by a rock on cooling grows in the opposite direction to that of the ambient field. Graham (1949) originally suggested the need for a self reversal mechanism to explain combinations of normal and reverse directions in rocks. Neel (1955) described seven theoretical self reversal mechanisms and Nagata et al. (1951) discovered the first natural example of self reversal in a dacitic pumice. Natural self-reversal now appears to be a comparatively rare phenomenon, occurring in less than 1% of all igneous rocks. The best check that self reversal has not taken place in a sample is to monitor remanence changes at elevated temperatures during both demagnetization and remagnetization in a laboratory field. Heller (1980), using this elevated temperature method, has demonstrated self-reversal processes in Olby-Laschamp lavas.

Igneous rocks may be bodily moved following their magnetization, for example through the rolling of blocks at the leading edge of an advancing lava flow. Such transposed remanences are best avoided at the collection stage by sampling the interior of massive lavas, but can also be recognised after collection, through lack of consistency of palaeomagnetic directions. Secondary magnetization may originate in igneous rocks in several ways, for example through the chemical growth of new iron oxide minerals or through viscous effects. The complications of secondary magnetizations are much less in Pleistocene rocks than in older materials. Their effects can generally be satisfactorily removed by thermal or alternating field demagnetization techniques.

Sediments As the majority of sediments do not carry a true record of the ancient field, many checks and reliability criteria need to be employed in palaeomagnetic investigations of sediments. The most diagnostic and most straightforward check is the demonstration of a repeatable signal in duplicate sediment sequences or cores. The further apart the repeatable sequences, while still lithologically or biologically firmly correlatable, the better. Another useful check is to see whether unusual palaeomagnetic directions occur at, or close to, lithological boundaries or close to the ends of core sections. Such features are almost always connected with sedimentological or coring disturbance effects and not with ancient field variations.

Bioturbation can cause remanence acquisition to be delayed until sediments are buried beneath a surface bioturbation zone. Bioturbation depth can be estimated from sediment mixing studies (e.g. Ruddiman and Glover, 1972, Guinasso and Schink, 1975) and from remanence intensity variations through polarity transition records (e.g. Denham and Chave, 1982, Hyodo, 1984) and may amount to tens of centimetres.

Other difficulties associated with sediments are bedding and current effects, compaction, recent weathering, and distortion during subsampling and coring. Bedding errors arise from deposition on sloping surfaces, while current errors are caused by grain rotations connected with water flow-induced shear stresses. Both of the latter effects lead to systematically low inclinations and can lead to errors in declination. Sediment compaction after remanence acquisition can also lead to mechanical flattening of the magnetic inclination.

One of the most difficult problems in palaeomagnetic studies of Pleistocene sediments is recognition of the inclination error of depositional remanence. The effect of inclinational error, caused by particles tending to come to rest with their long axes parallel to the bedding plane, was clearly recognised in one of the first studies of the magnetic remanence of sediments (Ising, 1942) and has since been well documented in laboratory deposition studies (King, 1955). The widespread nature of the effect is illustrated by the low inclinations to be found in many Pleistocene sediments. In comparison, the inclinations found in Pleistocene igneous rocks much more closely resemble those of the present day field. Indeed so many instances of incorrect attribution of low inclinations in sediments to ancient field behaviour (particularly excursions) can now be demonstrated (e.g. Opdyke, 1976, Verosub and Banerjee, 1977 and Banerjee et al. 1979) that serious stratigraphers are strongly advised to disregard low magnetic inclination results in sediments for dating and correlation purposes.

Lovlie (1989), despite all the above difficulties, remains optimistic that additional laboratory based techniques, e.g. magnetic fabric analyses, will one day be sufficiently developed to allow discrimination between sediment records distorted by the various biasing effects, described above, and genuine ancient field signals. He suggests that more extensive determinations of anisotropies of magnetic susceptibility and remanence should allow the establishment of credible excursion and secular variation time scales based on sediment data.

Matching

Johnson and McGee (1983) have discussed limitations of the reversal magnetostratigraphic method when unconformities or hiatuses are present. They have shown a uniform sample spacing through time to be the most effective strategy for promoting good matching of magnetic records. Errors in palaeomagnetic data and gaps in rock or sediment successions can at times obscure correlations to such a degree as to make matching untenable. Seven examples of matching Pleistocene magnetic records with previously dated geomagnetic behaviour patterns are discussed in the following sections.

MAGNETOSTRATIGRAPHIC APPLICATION OF THE PLEISTOCENE POLARITY TIME SCALE

Figure 9.2 illustrates the use of the Olduvai, Jaramillo and Matuyama-Brunhes boundary sequence of geomagnetic markers in dating a sequence of loesses and soils at Lochuan in China. The Lochuan inclination data are of high quality and can be matched with the polarity sequence of Figure 9.1 without difficulty, to yield the five ages listed in Figure 9.2. By contrast the inclination data of the North Sea sediments of Figure 9.3, although spanning a broadly similar time span to the Lochuan sediments and although containing both normal and reverse inclinations, are too

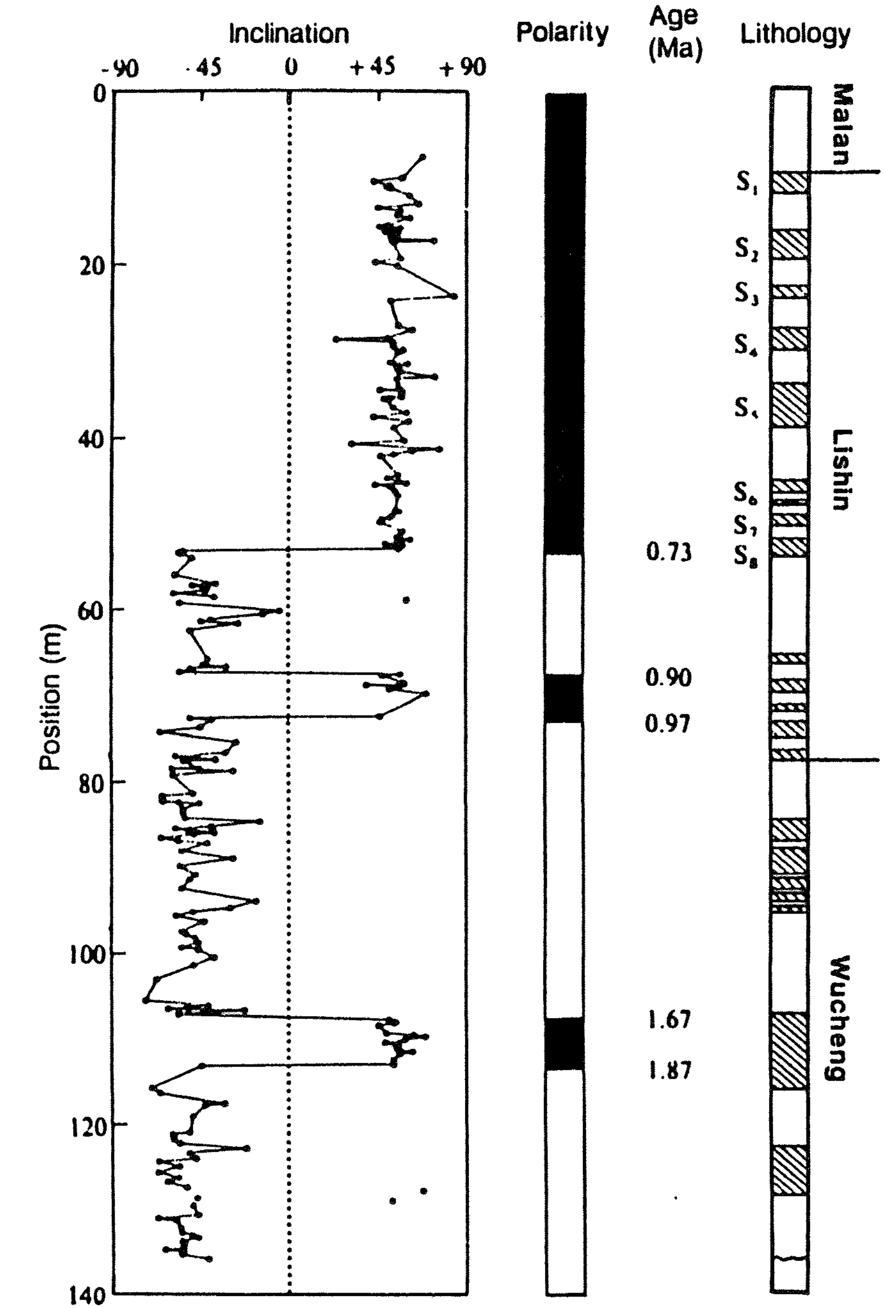

Figure 9.2 *Variation of magnetic inclination with depth in a sequence of loess (white) and soil layers (hatched) in the Lochuan borehole. Ages of the polarity chron boundaries after Mankinen and Dalrymple (1979). (Modified from Heller and Liu 1982.)*

scattered to be matched with any certainty with the polarity sequence of Figure 9.1. The most that can reasonably be gleaned from such poor quality data is that the older sediments with negative, reverse inclinations are likely to be at least 700 ka in age.

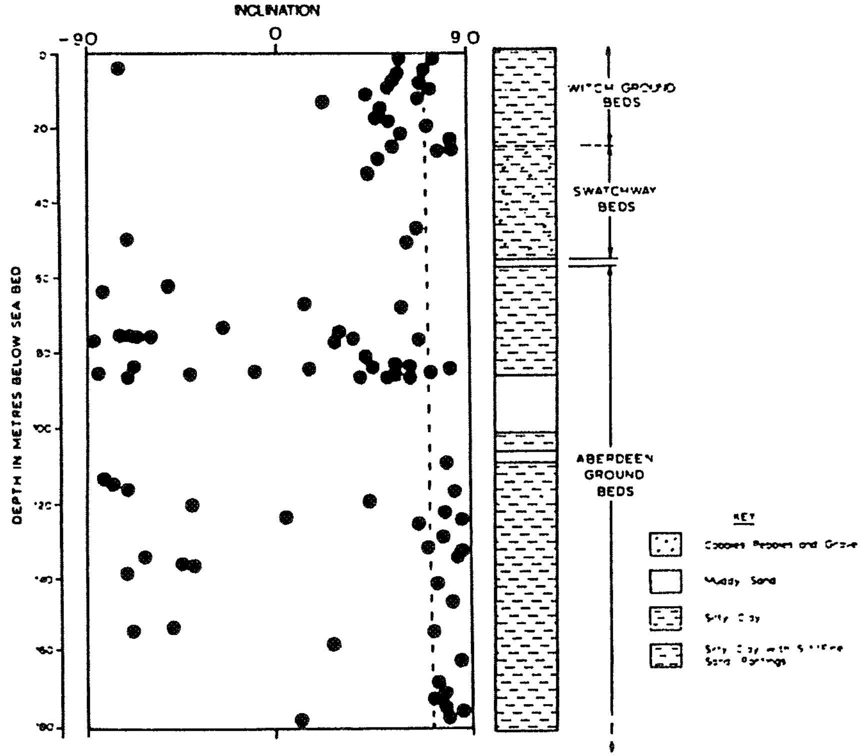

Figure 9.3 *Variation of lithology and magnetic inclination with depth in the North Sea borehole SRN33. No clear sequence of polarity reversals can be distinguished. The negative inclinations in the Aberdeen Ground beds are taken to indicate that the sediments are of Matuyama age or older. The boundary between the normally magnetized Swatchway beds and mixed polarity of the Aberdeen Ground beds is probably a hiatus. The occasional reversed inclination to be found in the Witch Ground beds is interpreted to have been caused either by inadvertent overturning of core segments, by sedimentological effects or by mechanical disturbance. (Modified from Begg 1979.)*

Another example of good quality palaeomagnetic data is provided by the declination logs of two deep sea cores used in oxygen isotope studies, as shown in Figure 9.4. In core Vema 28-239 the Olduvai, Jaramillo and Matuyama-Brunhes boundary sequence can be recognised in the declination changes. The scattered declinations at the top of the core are most probably to be interpreted as deriving from coring disturbances. In core Vema 28-238, the Matuyama-Brunhes boundary is found at

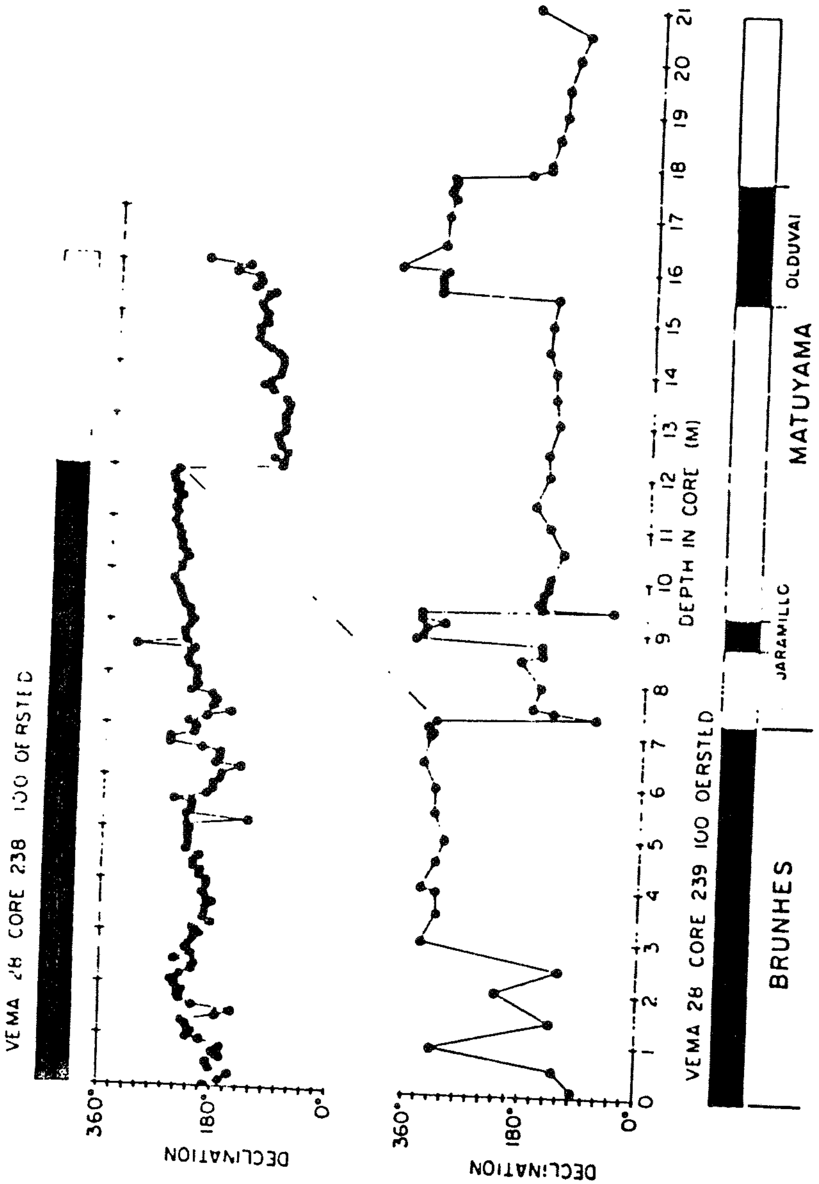

Figure 9.4 Variation of magnetic declination with depth in cores V28-238 and V28-239. The cores are not orientated with respect to true north. (From Shackleton and Opdyke, 1973.)

a greater depth than in core 28-239 implying a higher deposition rate for these sediments.

The age of Pleistocene materials may also be determined through use of the polarity time scale when remanence directions have been determined *in situ*. Such *in situ* studies which avoid the need for sample collection are sometimes possible when dealing with the relatively strong magnetization of basalts. *In situ* normally magnetized materials tend to increase the strength of the geomagnetic field while reversely magnetized materials tend to reduce the geomagnetic field strength. The precise effect depends on the shape and orientation of the magnetized material and on the latitude of the site under investigation. Einarsson (1957) demonstrated that the remanence polarity of many Icelandic lavas can be detected *in situ* with the use of an ordinary compass. As an example of this approach, Piper (1971) has produced a palaeomagnetic map of the polarity chrons of the basalts of south-west Iceland by measuring the magnetic anomalies associated with the lavas.

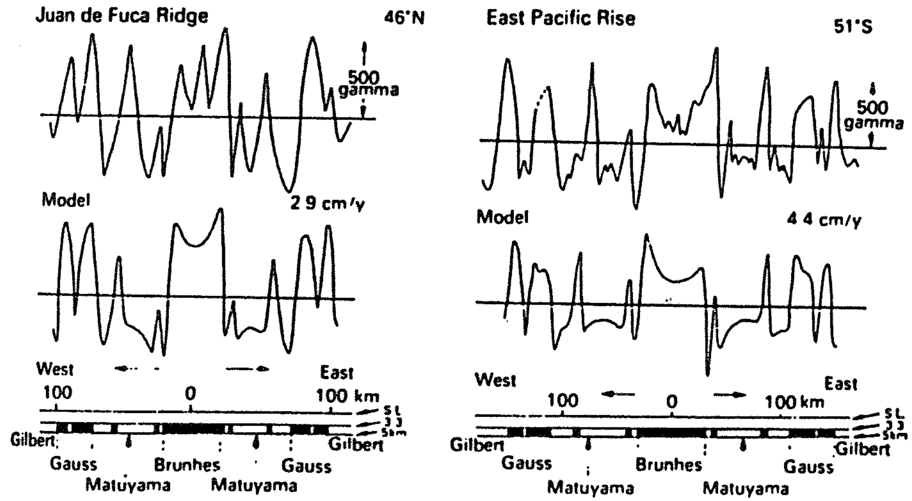

Figure 9.5 *Observed magnetic anomaly profiles across the Juan de Fuca ridge and the East Pacific rise compared with the theoretical magnetic anomaly according to the Vine-Matthews hypothesis assuming the palaeomagnetic polarity time-scale of Figure 9.1. It is assumed that the magnetic blocks are confined to oceanic layer 2. Black denotes normal magnetization, unshaded denotes reverse. (After Bott, 1971 redrawn from Vine, 1966.)*

The greatest use of palaeomagnetic dating of magnetic anomalies has been in marine surveys, through the application of the Vine-Matthews hypothesis of sea-floor spreading. Figure 9.5 plots magnetic anomaly profiles, measured by towing a proton precession magnetometer at the sea surface, across the Juan de Fuca and East Pacific ridges. High field intensities are to be interpreted as being caused by the remanence of normally magnetized ocean floor beneath the magnetometer. The central positive anomalies thus overlie Brunhes age sea floor. The sequences of anomalies can be neatly accounted for by symmetrical blocks of normally and reversely magnetized oceanic crust as illustrated at the bottom of Figure 9.5. The observed and modelled anomaly profiles can be seen to correlate well. Matching of marine anomaly profiles with the palaeomagnetic polarity time scale allows oceanic crust to be dated all around the world in this most elegant manner.

OTHER MAGNETIC CORRELATION METHODS

Secular Variation

Some sediment sequences hold records of palaeomagnetic direction fluctuations which resemble secular geomagnetic changes which have taken place over a time scale of centuries. A potential magnetostratigraphic method is to match these fluctuations with a previously dated, secular variation record (Mackereth, 1971, Thompson 1973). This secular variation approach to magnetostratigraphy is much more difficult to apply than that of polarity reversals because:- (1) secular geomagnetic changes are not global features, (2) type secular variation records are generally based on lake sediments which have proved to be exceedingly difficult to date with better than 10% accuracy, whereas the polarity time scale is based on results from basalts which have proved to be ideal rocks for potassium-argon dating, and (3) secular variation patterns with amplitudes of only some 20° are more susceptible to noise than polarity changes of 180°.

Figure 9.6 plots Holocene palaeomagnetic direction fluctuations from three lakes in Iceland as an illustration of the variation in quality of palaeomagnetic secular variation data. The pre 1500 BP Vatnsdalsvatn record of Figure 9.6a from Thompson and Turner (1985) has been found in four other cores from lake Vatnsdalsvatn and is interpreted as a useful palaeomagnetic signal, largely reflecting direction changes of the ancient geomagnetic field. The Draghalsvatn record of Figure 9.6b is an example of a disappointing site for palaeomagnetic studies at which the noise, particularly in inclination, totally overwhelms any geomagnetic secular variation signal. The Svinavatn records of Figure 9.6c and d are further examples of unusable palaeomagnetic records. Although the Svinavatn sediments have good magnetic stability, strong intensity, reasonable within-core consistency, sensible mean directions and typical amplitudes of variation, they do not repeat between cores and consequently cannot be interpreted as having correctly recorded the ancient geomagnetic field. Coring disturbance, core orientation difficulties, compaction, bioturbation, remagnetization, micro-slumping, grain reorientation, sampling errors, instrumental effects and, in particular, the inclination error of depositional remanence (Ising, 1942) all serve to degrade and distort the palaeomagnetic signal of sediments.

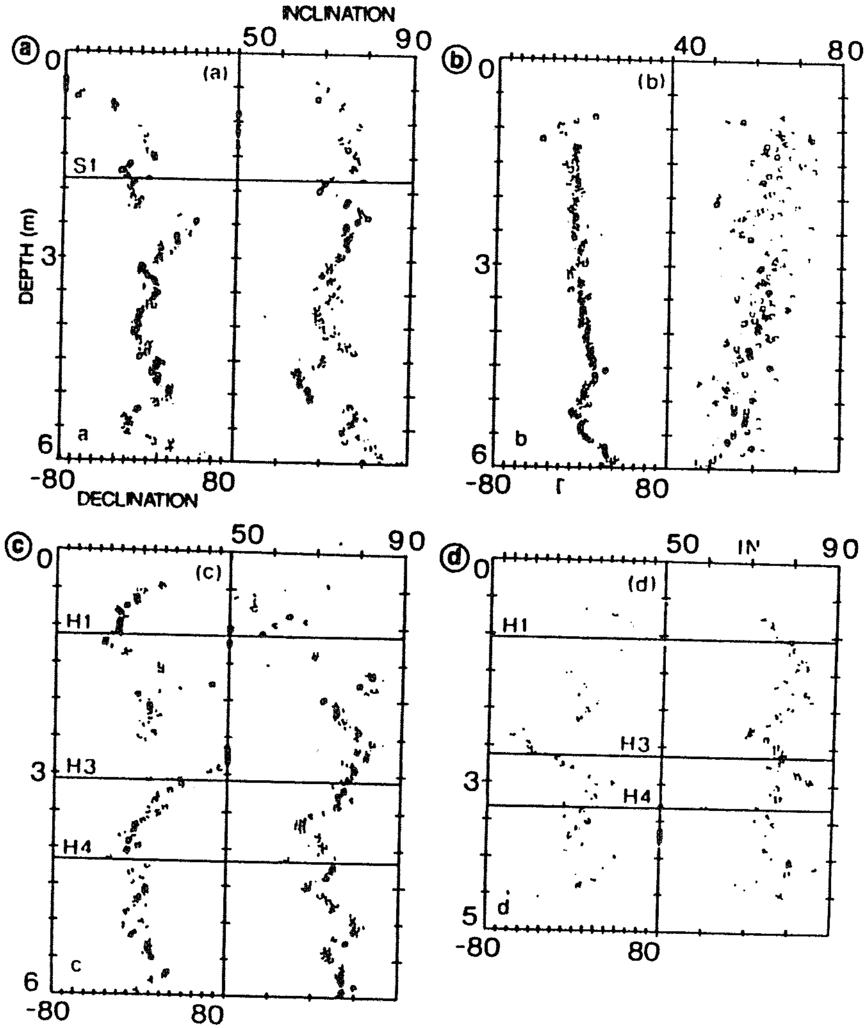

Figure 9.6 *Variation of relative declination and relative inclination with depth in sediment cores from three lakes in Iceland. Tephra horizons, marked as horizontal lines: H-1 Hekla 846 years BP, H-3 Hekla 2980 years BP, H-4 Hekla 4545 years BP, S1 Snaefells 1705 years BP. (a) Vatnsdalsvatn. The palaeomagnetic records are scattered at the top but may reflect the ancient field in the sediments below 1.5 m depth. (b) Draghalsvatn. The inclination record is too scattered for magnetostratigraphic work. (c) and (d) Two Svinavatn cores. The Svinavatn palaeomagnetic directions show some internal coherence, but are not repeatable between cores (compare the records between tephras H-4 and H-3 and between H-3 and H-2). The Svinavatn sediments are also of no magnetostratigraphic value, despite their internal consistency, stability under alternating field demagnetization and mean inclination being close to that of the axial dipole field.*

Not surprisingly, for the great majority of sediments a usable palaeomagnetic record of secular variation is not recoverable with present sampling and measuring techniques.

Mineral Magnetism

Magnetic investigations can provide lithostratigraphies in addition to chronostratigraphies when mineral magnetic as well as palaeomagnetic parameters are measured. As mineral magnetic investigations involve laboratory-induced artificial magnetizations and remanences, they are quite distinct from palaeomagnetic studies of the ancient geomagnetic field. Mineral magnetic stratigraphy is largely related to variations in the concentrations, compositions and grain sizes of the constituent iron oxide minerals (Thompson and Oldfield, 1986). An example of the use of mineral magnetic lithostratigraphy in Pleistocene studies (Oldfield and Robinson, 1985) is shown in Figure 9.7, where the mineral magnetic parameter 'S' is plotted against depth for a sediment core from the North Atlantic. The 'S' parameter of Stober and Thompson (1979) is chosen to emphasise variation in iron oxide composition and grain size. The 'S' ratio is defined as the ratio of the laboratory induced isothermal remanences in fields of 100 mT and 1 T. The mineral magnetic fluctuations in 'S' ratio in the North Atlantic can be correlated with the oxygen isotope variations of the last 20 ka (Imbrie et al., 1984). High 'S' ratios occur in the interglacial deposits. These high ratio values indicate that higher proportions of 'haematite' as opposed to 'magnetite' minerals occur in the interglacials. Oldfield and Robinson (1985) interpret the variations in magnetic ratio as being related to changing atmospheric circulation patterns between glacial and interglacial periods with proportionally more haematite-rich Saharan dust contributing to the sediments at the coring site during interglacials. Another good example of mineral magnetic stratigraphy in ocean sediments is the work of Bloemendal and deMenocal (1989) in the Arabian Sea and the eastern tropical Atlantic. Magnetic susceptibility measurements also offer considerable potential for correlation between marine and terrestrial sequences, as for instance in Kukla et al. (1988) for the Chinese loess.

BRITISH PLEISTOCENE MAGNETOSTRATIGRAPHY

The full complement of Pleistocene polarity changes of the Olduvai, Cobb Mountain, Jaramillo, Matuyama-Brunhes boundary sequence are to be found in the sediments on the west flank of the Rockall bank (Shackleton et al., 1984).

On land, in Britain, reversely magnetized Pleistocene sediments are rare (van Montfrans, 1971 and Table 9.1). In North Sea sediments, reverse magnetizations, presumably of Matuyama age (Begg, 1979, Stoker et al. 1983), are to be found, although as described above the quality of much of the North Sea palaeomagnetic data has been very poor.

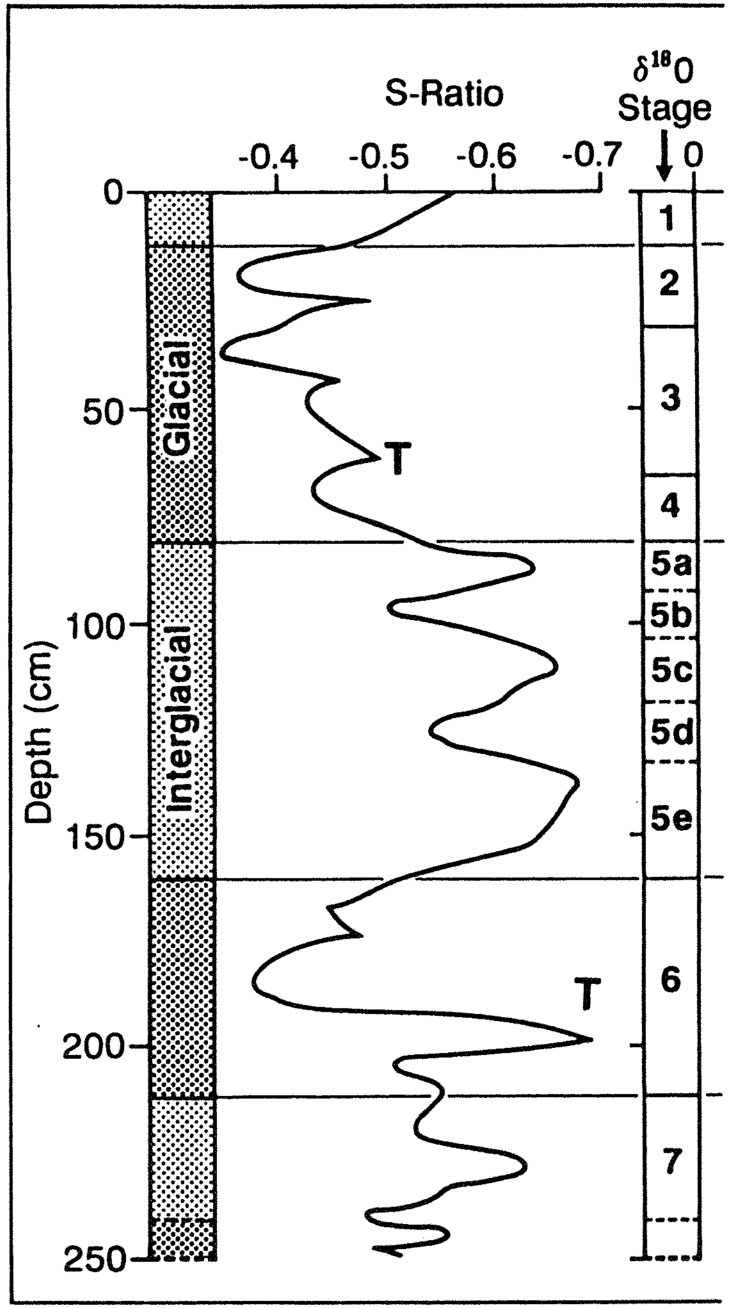

Figure 9.7 *Variation of mineral magnetic ratios with depth in a North Atlantic core. High magnetic ratios corresponding to high 'haematite to magnetite' ratios are found in the interglacial sediments. Tephra layers indicated by the letter T. (After Oldfield and Robinson, 1985.)*

Table 9.1 *Intensity, stability and polarity of some Pleistocene sediments from the British Isles. Polarity N = normal, R = reversed, ? = indeterminate.*

Locality	Lithology or zone	Stage	Collector	Number of Samples	Intensity of NRM mA/m	Median Destructive Field mT	Polarity
Gallowflat	clay	Devensian	Thompson	15	33	—	N
Maidenhall	silts and clays	Ipswichian	Wymer	4	2	15	N?
Burtle beds	Underlying silts and clays	Ipswichian	Heyworth	14	0.5	15	N
Hoxne	Gipping silts	Early Wolstonian	Gibbard	4	1	—	N
Marks Tey	Lacustrine	Hoxnian	Turner/Thompson	12	16	—	N
	laminated clay			14	10	—	N
Hitchin	lacustrine zone I	Hoxnian	Gibbard	15	0.7	—	N
Gort	lacustrine	Hoxnian	Turner	4	0.2	20	N
Happisburgh	clay	Anglian	Thompson	2	—	—	?
	lower till	Anglian		21	—	—	?
Sugworth	clay	Cromerian	Osmaston	27	470	—	N
	sands and silts	Cromerian		15	6	—	N
Westbury-sub-Mendip	silty clays	Middle Pleistocene	Stringer	5	<1	—	N?
Nettlebed	lacustrine	Pre-Cromerian	Turner	4	0.4	15	N?
Chillesford	blue clay	Pastonian	Thompson	13	1.8	—	N
	fine-grained band in Crag	Waltonian		2	2	—	?
Aldeburgh	blue clay	Pastonian	Thompson	15	1.5	—	N
Eastern Bavents	brown clay	Baventian	Collins/Thompson	28	0.8	5	N
	dark grey clay	Baventian		25	18	10	N
Cove Bottom	grey clay	Baventian	Thompson	10	0.5	6	N
	dark grey clay	Baventian		20	47	10	N
Ludham	L4	Baventian	Roy. Soc. Borehole	10	1	10	?
	L2	Thurnian		20	2	<10	N?
	L1	Ludhamian		6	1	20	N
Tattingstone Hall	silt layers in Crag	Ludhamian	Thompson	7	0.8	—	?
Bawdsey	silt band in Crag	Waltonian	Thompson	7	0.7	—	N
North Sea (Fladen)							
SLN33	(0-20 m)	Witchground beds	BGS Borehole	16	20	35	N
	(20-58 m)	Swatchway beds		10	8	30	N
	(58-182 m)	Aberdeen ground beds		70	10	50	R+N
77/2	(0-22 m)	Witch Ground Beds	BGS Borehole	21	6	—	N
	(22-35 m)	Swatchway beds		19	4	—	N
	(35-124 m)	Lower Channel deposits		19	10	—	N
	(124-200 m)	Aberdeen Ground Beds		216	10	>20	R+N
North Sea (Forties)							
77/3	(0-140 m)	Channel deposits	BGS Borehole	51	8	—	N

The lack of reversals in the Pleistocene sediments of East Anglia compared with those of the Netherlands was one of the main reasons that led Zagwijn (1975) to propose that there were long stratigraphic gaps covering the upper part of the Lower Pleistocene and the lower part of the Middle Pleistocene in the known East Anglian successions.

SUMMARY

The state-of-the-art of the five magnetic dating and correlation techniques of Pleistocene materials discussed above is well summarized by the following comments from recent review articles:-

1) Reversal magnetostratigraphy

> Palaeomagnetic stratigraphy offers one of the most promising methods for establishing world-wide correlations of (terrestrial) Quaternary events ... (and) ... of correlating the marine and terrestrial records.
>
> *Lowe and Walker (1984).*

2) Excursions

> Although it seems clear that excursions exist ... it seems premature to use these changes as a dating method.
>
> *Tarling (1983).*

3) Secular variation

> Master curves of palaeomagnetic (secular variations) may provide a rapid means of dating sediment sequences in lakes ... the precision which may be achieved by palaeomagnetically based (secular variation) dating cannot be estimated accurately from results obtained so far ... it will inevitably vary from lake to lake.
>
> *Oldfield (1977).*

4) Mineral magnetism

> This method would appear to offer considerable potential as a means of correlation in a range of sediments and ... time scales where variations in mineral magnetic properties reflect synchronous environmental properties.
>
> *Lowe and Walker (1984).*

REFERENCES

Aitken, M.J. 1974. *Physics and Archaeology*.Clarendon Press, Oxford, pp. 291.

Banerjee, S.K., Lund, S.P. and Levi, S. 1979. Geomagnetic record in Minnesota sediments – Absence of the Gothenburg and Erieau excursions. *Geology*, **7**, 588-591.

Bauer, L.A. 1896. On the secular motion of a free magnetic needle. II. *Physics Review*, **3**, 34-48.

Begg, P. 1979. Magneto-stratigraphy. Unpublished B.Sc. dissertation, University of Edinburgh.

Bloemendal, J. and DeMenocal, P. 1989. Evidence for a change in the periodicity of tropical climate cycles at 2.4 Myr from whole-core magnetic susceptibility measurements. *Nature*, **342**, 897-900.

Bonhommet, N. and Babkine, J. 1967. Sur la presence d'aimantations oversees dans la Chaine des Puys. *Comes Rendus des Academie des Sciences Paris*, **264**, 92-94.

Bott, M.H.P. 1971. *The Interior of the Earth*.Edward Arnold, London, pp. 316.

Champion, D.E., Dalrymple, G.B. and Kuntz, M.A. 1981. Radiometric and palaeomagnetic evidence for the Emperor reversed polarity event at 0.46 ± 0.05 M.y. in basalt lava flows from the Eastern Snake River Plain, Idaho. *Geophysical Research Letters*, **8**, 1055-1058.

Denham, C.R. and Chave, A. 1982. Detrital remanent magnetization: viscosity theory of the lock-in zone. *Journal of Geophysical Research*, **87**, 7126-7130.

Einarsson, T. 1957. Magneto-geological mapping in Iceland with the use of a compass. *Philiosphical Magazine Supplement Advances in Physics*, **6**, 232-239.

Graham, J.W. 1949. The stability and significance of magnetism in sedimentary rocks. *Journal of Geophysical Research*, **54**, 131-167.

Guinasso, N.L. and Schink, D.R. 1975. Quantitative estimates of biological mixing rates in abyssal sediments. *Journal of Geophysical Research*, **80**, 3032-3043.

Harland, W.B., Cox, A.V., Llewellyn, P.G., Picton, C.A.G., Smith, A.G. and Walters, R. 1982. *A Geologic Time Scale*. Cambridge University Press, New York, pp. 128.

Heller, F. 1980. Self-reversal of natural remanent magnetisation in the Olby-Laschamp lavas. *Nature*, **284**, 334-335.

Heller, F. and Liu, T. 1982. Magnetostratigraphical dating of loess deposits in China. *Nature*, **300**, 431-433.

Hirooka, K. 1971. Archaeomagnetic study for the past 2000 years in Southwest Japan. *Memoirs of the Faculty of Science Kyoto University. Geology and Mineralogy Series*, **XXXVIII**, 167-207.

Hyodo, M. 1984. Possibility of reconstruction of the past geomagnetic field from homogeneous sediments. *Journal of Geomagnetism and Geoelectricity*, **36**, 45-62.

Imbrie, J., Hays, J.D., Martinson, D.G., McIntyre, A., Mix, A.C., Morley, J.J., Pisias, N.G., Prell, W.L. and Shackleton, N.J. 1984. The orbital theory of Pleistocene climate: support from a revised chronology of the marine and ¹⁸O record. In *Milankovitch and Climate* (ed Berger, A.L.), Part I, Reidel, Boston, 269-305.

Imbrie, J.I. and Imbrie, K.P. 1979. *Ice Ages*, Macmillan, London, pp. 224.

Irving, E. 1964. *Paleomagnetism and its Application to Geological Problems*, Wiley, New York, pp. 399.

Ising, G. 1942. On the magnetic properties of varved clay. *Arkiv for Matematik, Astronomi och Fysik*, **29**, 1-37.

Johnson, E.A., Murphy, T. and Torrenson, O.W. 1948. Prehistory of the Earth's magnetic field. **53**, 349-372.

Johnson, N.M. and McGee, V.E. 1983. Magnetic polarity stratigraphy: stochastic properties of data, sampling problems and the evolution of interpretations. *Journal of Geophysical Research*, **88**, 1213-1221.

King, R.F. 1955. The remanent magnetisation in artificially deposited sediments. *Monthly Notes of the Royal Astronomical Society*, **7**, 115-134.

Kukla, G., Heller, F., Liu, X.M., X.U., T.C., Liu, T.S. and An, Z.S. 1988. Pleistocene climates in China dated by magnetic susceptibility. *Geology*, **16**, 811-814.

Kukla, G.J. 1970. Correlations between loesses and deep-sea sediments. *Geologiska Foreningens i Stockholm Forhandlinghar*, **92**, 148-180.

Levi, S., Audunsson, H., Duncan, R.A., Kristjansson, L., Gillot, P-Y. and Jakobsson, S.P. 1990. Late Pleistocene geomagnetic excursion in Icelandic lavas: confirmation of the Laschamp excursion. *Earth and Planetary Science Letters*, **96**, 443-457.

Lovlie, R. 1989. Palaeomagnetic stratigraphy: a correlation method. *Quaternary International*, **1**, 129-149.

Lowe, J.J. and Walker, M.J. 1984. *Reconstructing Quaternary Environments*, Longman, London, pp. 389.

Mackereth, F.J. 1971. On the variation in direction of the horizontal component of remanent magnetisation in lake sediments. *Earth and Planetary Science Letters*, **12**, 332-338.

Mankinen, E.A. and Dairymple, G.B. 1979. Revised geomagnetic polarity time scale for the interval 0-5 M.y. B.P. Journal of *Geophysical Research*, **84**, 615-626.

Mankinen, E.A. and Gromme, C.S. 1982. Paleomagnetic data from the Coso range, California and current status of the Cobb Mountain normal geomagnetic polarity event. *Geophysics Research Letters*, **9**, 1279-1282.

Mankinen, E.A., Donnelyy, J.M. and Gromme, C.S. 1978. Geomagnetic polarity event recorded at 1.1 M.y. B.P. on Cobb Mountain, Clear Lake volcanic field, California. *Geology*, **6**, 653-656.

McDougall, I. 1979. The present status of the geomagnetic polarity timescale. In *The Earth: Its Origin, Structure and Evolution*, (ed. McElhinny, M.W.), Academic Press, New York.

Montfrans, H.M. Van. 1971. Palaeomagnetic dating in the North Sea Basin, unpublished Ph.D. Thesis. Amsterdam University.

Nagata, T., Atkimoto, S. and Uyeda, S. 1951. Reverse thermoremanent magnetism. *Proceedings of the Japan Academy*, **27**, 643-645.

Neel, L. 1955. Some theoretical aspects of rock magnetism. *Advances in Physics*, **4**, 191-242.

Oldfield, F. 1977. Lakes and their drainage basins as units of sediment-based ecological study. *Progress in Physical Geography*, **1**, 460-504.

Oldfield, F. and Robinson, S.G. 1985. Geomagnetism and palaeoclimate. In *The Climatic Scene*, (eds. Tuley, M.J. and Sheail, G.M.). Allen Unwin: London, 186-205.

Opdyke, N.D. 1972. Palaeomagnetism of Deep-Sea cores. *Reviews in Geophysics and Space Physics*, **10**, 213-249.

Opdyke, N.D. 1976. Discussion of paper by Morner and Lanser concerning the palaeomagnetism of Deep-Sea core A179-15. *Earth and Planetary Science Letters* **29**, 238-239.

Piper, J.D. 1971. Ground magnetic studies of crustal growth in Iceland. *Earth and Planetary Science Letters* **12**, 199-207.

Ruddiman, W.F. and Glover, L.K. 1972. Vertical mixing of ice-rafted volcanic ash in North Atlantic sediments. *Geological Society of America Bulletin*, **83**, 2817-2836.

Shackelton, N.J. and Opdyke, N.D. 1973. Oxygen isotope and palaeomagnetic stratigraphy of equatorial Pacific core V28-238. *Quaternary Research*, **1**, 39-55.

Shacketon, N.J. and Opdyke, N.D. 1976. Oxygen isotope and palaeomagnetic stratigraphy of Equatorial Pacific core V28-239. *Geological Society of American Memoir*, **145**, 449-464.

Shackleton, N.J., Backman, J., Zimmerman, H., Kent, D.V., Hall, M.A., Roberts, D.G., Schitker, D., Baldauf, J.G., Desprairies, A., Homrighausen, R., Huddlestun, P., Keene, J.B., Kaltenback, A.J., Krumsiek, K.A.O., Morton, A.C., Murray, J.W. and Westberg-Smith, J. 1984. Oxygen-isotope calibration of the onset of ice-rafting and history of glaciation in the North Atlantic region. *Nature*, **307**, 620-623.

Singh, G., Opdyke, N.D. and Bowler, J.M. 1981. Late Cainozoic stratigraphy, palaeomagnetic chronology and vegetational history from Lake George, N.S.W. *Journal of the Geological Society of Australia*, **28**, 435-452.

Stober, J.C. and Thompson, R. 1979. An investigation into the source of magnetic minerals in some Finnish lake sediments. *Earth and Planetary Science Letters*, **45**, 464-474.

Stoker, M.S., Skinner, A.C., Fyfe, J.A. and Long, D. 1983. Palaeomagnetic evidence for early Pleistocene in the central and northern North Sea. *Nature*, **304**, 332-334.

Stupavsky, M. and Gravenor, C.P. 1984. Paleomagnetic dating of Quaternary sediments: a review. In *Quaternary Dating Methods*, (ed. Mahaney, W.C.), Elsevier, Amsterdam, 123-140.

Tarling, D.H. 1983. *Palaeomagnetism*, Chapman and Hall, London, pp. 379.

Thompson, R. 1973. Palaeolimnology and palaeomagnetism. *Nature*, **242**, 182-184.

Thompson, R. 1983. ^{14}C dating and magnetostratigraphy. *Radiocarbon*, **25**, 229-238.

Thompson, R. 1984. A global review of palaeomagnetic results from wet lakes sediments. In *Lake Sediments and Environmental History*, (eds. Haworth, E.Y. and Lund, J.W.G.), Leicester University Press, Leicester, 145-164.

Thompson, R. 1984. Geomagnetic evolution: 400 years of change on planet Earth. *Physics of the Earth and Planetary Interiors*, **36**, 61-77.

Thompson, R. and Oldfield, F. 1986. *Environmental Magnetism*, Allen and Unwin, London, pp. 227.

Thompson, R. and Turner, G.M. 1985. Icelandic Holocene palaeomagnetism. *Physics of the Earth and Planetary Interiors*, **38**, 250-261.

Verosub, K.L. and Banerjee, S.K., 1977. Geomagnetic excursions and their palaeomagnetic record. *Reviews of Geophysics and Space Physics*, **15**, 145-155.

Vine, F.J. 1966. Spreading of the Ocean Floor: New evidence. *Science*, **154**, 1405-1415.

Zagwijn, W.H. 1975. Variations in climate as shown by pollen analysis, especially in the Lower Pleistocene of Europe. In *Ice Ages, Ancient and Modern*. (eds. Wright, A.E. and Moseley, F.), *Geophysical Journal Special Issue*, **6**, 137-152.