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Paleomagnetic dating and effects of Weichselian periglacial processes on the magnetization of early Pleistocene deposits (southern Netherlands, northern Belgium)

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Abstract

Early Pleistocene deposits of the Dutch-Belgian border area between Breda and Turnhout have been dated with paleomagnetic methods. Two normal polarity subzones in the sedimentary sequence are identified as the Olduvai and Jaramillo subchrons. The magnetozones are compared with the early Pleistocene pollenzones which are based on ecological changes reflecting Quaternary climatic variations. The Olduvai subchron was identified in pollenzones Tiglian C4 and C5 and the Jaramillo subchron in pollenzone Bavelian Bv3b. Late Pleistocene Weichselian periglacial processes clearly influenced the remanent magnetization of the early Pleistocene deposits. Melting of the ice-rich topzone of the Weichselian permafrost apparently modified the original remanence and led to complete remagnetization of the upper part of early Pleistocene clay beds.

Introduction

The study area is situated between Breda in the southern Netherlands and Turnhout in northern Belgium (Figure 1). During the early Pleistocene this region was part of the southern North Sea Basin. Deposition occurred in fluvial and tidal environments during glacial and interglacial periods (Zagwijn 1985; Kasse 1988, 1993). Because of uplift of the Belgian hinterland (Brabant Massif) and subsidence of the Netherlands during the middle and late Pleistocene, the early Pleistocene units, which are covered only by a thin layer of Weichselian periglacial sediments, are found close to the surface.

The chronostratigraphy of the Quaternary in the Netherlands and Belgium is based mainly on the repeated changes in vegetation caused by glacial and interglacial climates (Doppert & Zonneveld 1955; Dricot 1961; Paepe & Vanhoorne 1970; Zagwijn & De Jong 1984; Zagwijn 1960, 1985). The paleomagnetic method, which proved to be useful for correlation and chronostratigraphic dating of sedimentary sequences, was used less frequently (Van Montfrans 1971; Hus et al. 1976).

Paleomagnetism is based on the property of rocks to record past geomagnetic field variations by acquisition of a natural remanent magnetization. Magnetostratigraphy relies on the global synchronous polarity changes or reversals of the geomagnetic field recorded in rock strata, and magnetic polarity time scales are well-established (Cande & Kent 1992, 1995). As polarity transitions last less than 10,000 years, the polarity boundaries provide excellent time markers. The global nature of field reversals allows to correlate marine and continental deposits, and provides the chronological framework for biostratigraphic units (pollenzones).

In contrast to radiometric dating methods, the magnetostratigraphic units are based on inherently time-independent physical properties. Often, one of the major problems in the magnetostratigraphic study of continental deposits is the lack of a continuous sedimentary sequence. On the continents, changes in the accumulation rate and erosion may result in a fragmentary record in which, compared to deep sea records, only part of the polarity transitions or boundaries are

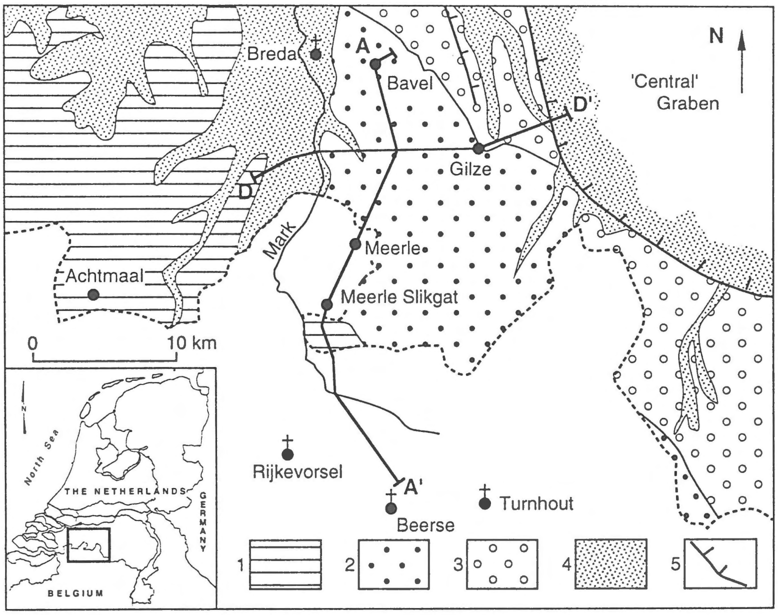

Figure 1. Generalized geological map of the study area (after Zagwijn & Van Staalduinen 1975) showing the locations of the two cross-sections of Figures 3 and 4 and the investigated sites (black circles). 1 = Tegelen Fm; 2 = Kedichem Fm; 3 = Sterksel Fm (1, 2 and 3 are covered by a thin unit of mostly Weichselian sands); 4 = younger, mostly Weichselian and/or Holocene deposits; 5 = fault.

found. Therefore, independent information about the lithostratigraphic position, the depositional environment and the biostratigraphic age of the sedimentary units is needed to be able to correlate the magnetozones encountered with the geomagnetic polarity time scale (GPTS).

Until the recent investigations by Kasse (1988, 1990, 1993) and Vandenberghe & Kasse (1989), the litho- and biostratigraphy of the early Pleistocene in the southern Netherlands and northern Belgium was a matter of debate. Furthermore, the influence of Weichselian periglacial processes on the remanent magnetization of these deposits had not been clearly evaluated. Indications in this study suggest complete remagnetization of the upper part of some early Pleistocene deposits. Thus it is necessary to re-evaluate and reinterpret earlier paleomagnetic results (Van Montfrans 1971).

Methods

Samples for paleomagnetic investigations have been taken from silt and clay layers in exposures and boreholes. Oriented samples were obtained in the exposures of Meerle, Bavel and Gilze, in the latter two with the aid of plastic tubes. In the boreholes only the inclination can be obtained because the azimuths of these boreholes are unknown.

The natural remanent magnetization (NRM) of the samples has been measured with a three-axis SCT superconducting magnetometer with a sample access port of 6.3 cm. Prism and cylindre-shaped samples were measured four times and cube-shaped samples six times, corresponding to different orientations of the sample. Hence the average components along three perpendicular sample axes were obtained and the average intensity and direction (inclination and declination) calculated.

To isolate the most stable magnetization component or so-called characteristic remanence (ChRM), several pilot samples were progressively demagnetized in increasing alternating fields from 0 to 70 mT in 13 steps. The NRM decay curves of the pilot samples are represented in Figure 2. After deposition of the sediment and during storage in the laboratory, the samples may have acquired a viscous remanent magnetization (VRM) in the present magnetic field. The demagnetization tests indicated the presence of a VRM which could be removed in moderate alternating fields (15–30 mT). The paleomagnetic results are plotted in diagrams representing the inclination, declination (only

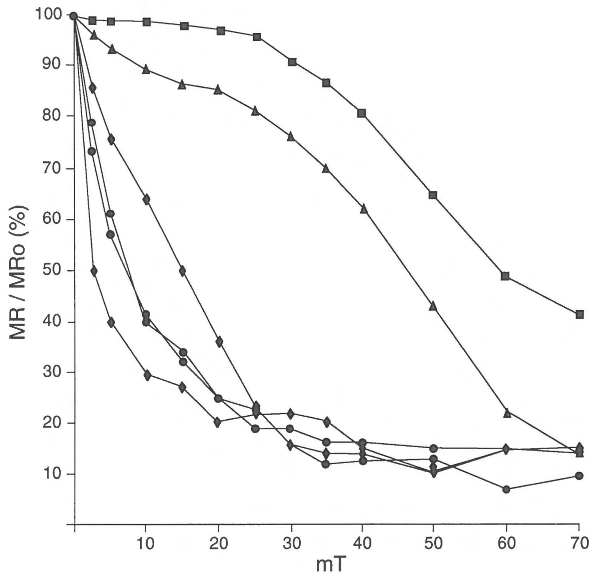

Figure 2. Alternating field demagnetization curves or ratio of residual remanence to original remanence (MR/MRo) versus alternating field. ● = Turnhout Mbr at Meerle (2 samples); ▲ = Turnhout Mbr at Achtmaal; ■ = Turnhout Mbr at Meerle Slikgat; ◆ = Gilze Mbr at Gilze (2 samples).

Table 1. Chronostratigraphy, lithostratigraphy and provenance of the early Pleistocene units in the Breda–Turnhout area. Modified after Kasse (1990). R = Rhine, M = Meuse, S = Scheldt, L = Local provenance. The numbers between brackets refer to the unit numbers in the text.

CHRONOSTRATIGRAPHY			LITHOSTRATIGRAPHIC UNITS IN BREDA - TURNHOUT AREA				
LATE PLEISTOCENE			Twente Formation			(11)	L
MIDDLE PLEISTOCENE			Eindhoven Formation			(10)	L
MIDDLE PLEISTOCENE			Sterksel Formation			(9)	R+M (+S)
EARLY PLEISTOCENE	BAVELIAN		Kedichem Formation	Bavel Member		(8)	R+M
EARLY PLEISTOCENE	MENAPIAN		Kedichem Formation	Gilze Member (7)	Spruitenstroom Clay		S+M+R
EARLY PLEISTOCENE	MENAPIAN		Kedichem Formation	Gilze Member (7)	Alphen Sand		S+M
EARLY PLEISTOCENE	WAALIAN		Kedichem Formation	Gilze Member (7)	Gilze Clay		S (+R)
EARLY PLEISTOCENE	EBURONIAN		Kedichem Formation	Gilze Member (7)	Appelenberg Sand		S
EARLY PLEISTOCENE	TIGLIAN	C5	Tegelen Formation	Turnhout Member		(6)	R (+S)
EARLY PLEISTOCENE	TIGLIAN	C4	Tegelen Formation	Beerse Member		(5)	S
EARLY PLEISTOCENE	TIGLIAN	C3	Tegelen Formation	Rijkevorsel Member		(4)	R (+S)
EARLY PLEISTOCENE	PRAETIGLIAN		Kieseloölite Formation	hiatus			
PLIOCENE			Kieseloölite Formation	Merksplas Sand		(3)	

for the exposures) and residual magnetic intensity. For each sample also the latitude of the virtual geomagnetic pole (VGP) is given. Northern (southern) VGP latitudes represent normal (reversed) polarities. During a transitional period intermediate field directions may occur.

Lithostratigraphy

The Pleistocene lithostratigraphic sequence of the Breda–Turnhout area is illustrated by two cross-sections (Figures 3, 4) in which the sampling locations for paleomagnetic investigations are indicated. Detailed information concerning the lithostratigraphy, nomenclature, correlation and provenance of this sequence can be found in Kasse (1988, 1990, 1993).

Cross-section AA' (Figure 3), between Beerse and Bavel is approximately at right angles to the depth contours of the North Sea Basin (Zagwijn & Doppert 1978). The northern dip of the lithostratigraphic units diminishes from 2.2‰ for the top of the Kieseloölite Formation (unit 3 in Table 1) to 1.1‰ for the top of the late Tiglian Turnhout Member (unit 6).

Cross-section DD' (Figure 4) is situated south of Breda. The unit boundaries are more or less horizontal, because the section is parallel to the depth contours of the basin. East of Gilze, in the Central or Roer Valley Graben, the early Pleistocene units 4, 6 and 7 are covered by middle Pleistocene sandy sediments of the Sterksel Formation (unit 9).

Eleven lithostratigraphic units have been distinguished. In Table 1 these local units are correlated with the formal units in the Netherlands (Zagwijn & Van Staalduinen 1975) and Belgium (Paepe & Vanhoorne 1976). Unit 1 consists of glauconiferous, shell-bearing sands of the Oosterhout Formation. Unit 2 is characterized by medium (250–500 μm), shell-bearing sands of the nearshore Maassluis Formation. It is covered by and also laterally equivalent to the medium-grained, non-calcareous Merksplas Sand of the Kieseloölite Formation (unit 3).

The early Pleistocene units 4 to 8 are predominantly fine-grained. The Rijkevorsel Member (unit 4), Beerse Member (unit 5) and Turnhout Member (unit 6) have all been incorporated in the Tegelen Formation (Kasse 1988, 1990). Unit 4 (Rijkevorsel Mbr) consists of micaceous medium to fine sands (100–250 μm), with a clay layer at the top. The unit is characterized by a fining-upward sequence, which was formed in a brackish-

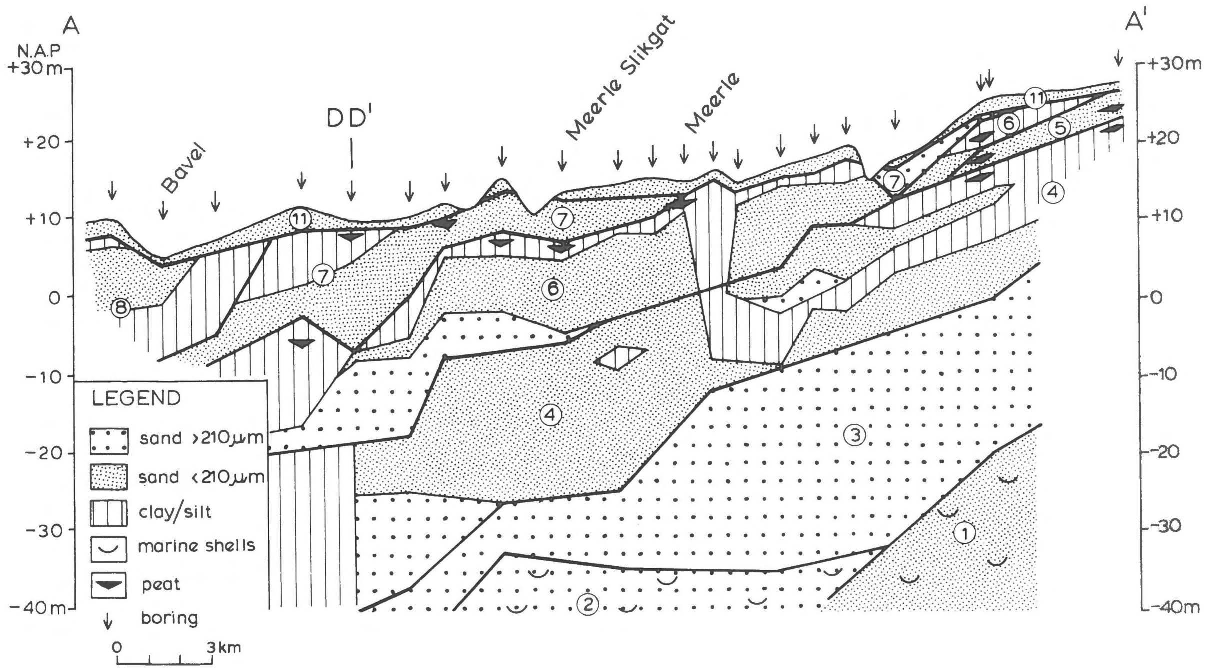

Figure 3. Lithostratigraphic cross-section AA', perpendicular to the depth contours of the North Sea Basin, between Beerse and Bavel (location in Figure 1). 1 = Oosterhout Fm; 2 = Maassluis Fm; 3 = Merksplas Sand, Kieseloölite Fm; 4–6 = Tegelen Fm; 4 = Rijkevorsel Mbr; 5 = Beerse Mbr; 6 = Turnhout Mbr; 7, 8 = Kedichem Fm; 7 = Gilze Mbr; 8 = Bavel Mbr; 9 = Sterksel Fm; 10 = Eindhoven/Twente Fm; 11 = Twente Fm. (Modified after Kasse 1988).

water tidal environment (Kasse 1988). To the north, fragments of marine molluscs are found in the unit (Figure 4). This indicates a lateral transition between the Rijkevorsel Member and the shell-bearing Maassluis Formation (Zagwijn & Van Staalduinen 1975).

Unit 4 is overlain by fine eolian sands and cryoturbated peat-beds (Figure 3: unit 5) of the Beerse Member (Kasse 1993). During the deposition of the Turnhout Member (unit 6), the Beerse Member was preserved only along the southern margin of the study area. Similar to the Rijkevorsel Member, the Turnhout Member is a brackish to fresh-water tidal deposit, characterized by a clear fining-upward sequence with peat beds in the upper part of the clay (Kasse 1988, 1990). This clay has been sampled for paleomagnetic analysis in clay pit Meerle and in several boreholes (Figure 3).

The Turnhout Member is covered by the Gilze Member (unit 7) of the Kedichem Formation. The thickness of the latter unit increases to the north (Figure 3). The generally fine, fluvial sands of Scheldt provenance have predominantly been deposited by braided river systems and locally by meandering rivers (Kasse 1988, 1990). The deposits of the meandering systems show a fining-upward sequence with backswamp clays at the top. Such clays have been investigated paleomagnetically at Gilze (Figure 4).

In the neighbourhood of Bavel, the Gilze Member has been eroded by the Rhine during the Bavelian (Figure 3). The valley was filled with Rhine-provenance sediments by a meandering style river (Bavel Mbr: unit 8). An 8-m-thick clay, probably formed in a meander cut-off, was analysed (Figure 3: Bavel).

In the Central or Roer Valley Graben, east of the faults in Figure 4, the middle Pleistocene Sterksel Formation (unit 9) directly overlies the Gilze Member (unit 7). The large thickness of the Sterksel Formation in the graben indicates considerable subsidence during the middle Pleistocene. Since unit 7, of early Pleistocene age, has an equal thickness across the faults, it has been concluded that the subsidence increased strongly at the end of the early Pleistocene and during the middle Pleistocene (Kasse 1988).

After the early middle Pleistocene, periglacial erosion was dominated by local, northward draining fluvial systems such as the Mark river (Figure 4: unit 10). During the Weichselian Pleniglacial and late Glacial, a thin eolian cover (1–2 m) was deposited over the landscape (unit 11: Twente Fm). It rests often directly

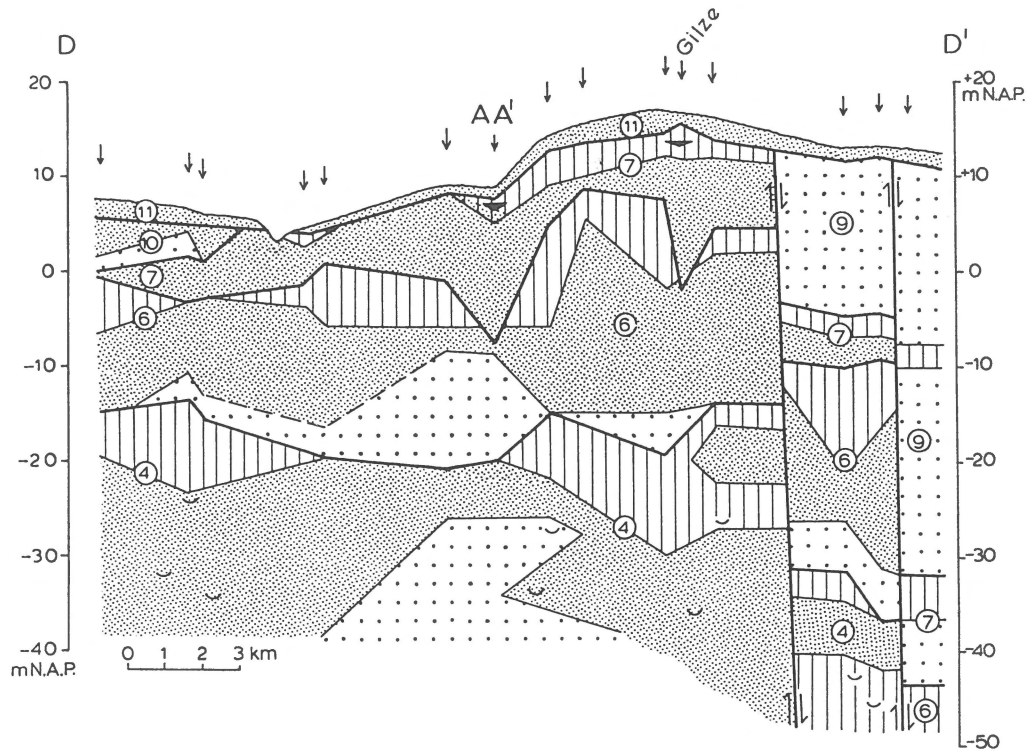

Figure 4. Lithostratigraphic cross-section DD' south of Breda parallel to the depth contours of the North Sea Basin (location in Figure 1, legend in Figure 3). (Modified after Kasse 1988).

on middle and early Pleistocene sediments, indicating a large hiatus. The duration of this hiatus increases to the south, where successively older units subcrop below the late Pleistocene unconformity.

Magnetostratigraphy

Magnetostratigraphy is based on the recognition of geomagnetic polarity reversals which are well-established in the polarity time scale of the last 110 Myr (Cande & Kent 1992, 1995). During the Quaternary, two major polarity zones occurred: the Matuyama reversed chron (2.58–0.78 Ma) and the Brunhes normal chron (0.78–0 Ma). Within the Matuyama zone, three polarity subchrons of shorter duration have been found with certainty (Réunion: 2.14–2.15 Ma; Olduvai: 1.77–1.95 Ma; Jaramillo: 0.99–1.07 Ma; Cande & Kent 1995). In the Netherlands, following Zagwijn (1985, 1989), the Pliocene-Pleistocene boundary is conventionally placed as early as 2.3 Ma and the early to middle Pleistocene transition at approximately 0.7 Ma. The correlation of the three polarity subchrons to specific bio(chrono)stratigraphic stages of the early Pleistocene in the Netherlands is often difficult. For instance, the Jaramillo subzone was first correlated with the Waalian (Van Montfrans 1971), and later with the Bavelian stage (Zagwijn & De Jong 1984).

The paleomagnetic results obtained from the early Pleistocene deposits in the Breda–Turnhout area are presented below.

Bavel

In clay pit Bavel (Figure 3), an up to 8 m thick clay layer belonging to the Bavel Member of the Kedichem Formation, was investigated (Figure 5). The clay has been correlated on pollen-analytical grounds with the Bavel Interglacial of the Bavelian stage (Zagwijn & De Jong 1984; Kasse 1988).

The basal part has a negative inclination and a southward pointing declination and is thus clearly reversed, while the upper part from 1.5 to 4.75 m with a positive inclination and northerly declination is normal. The transition from normal to reversed polarity is situated between 5.25 and 9.0 m. The scatter of

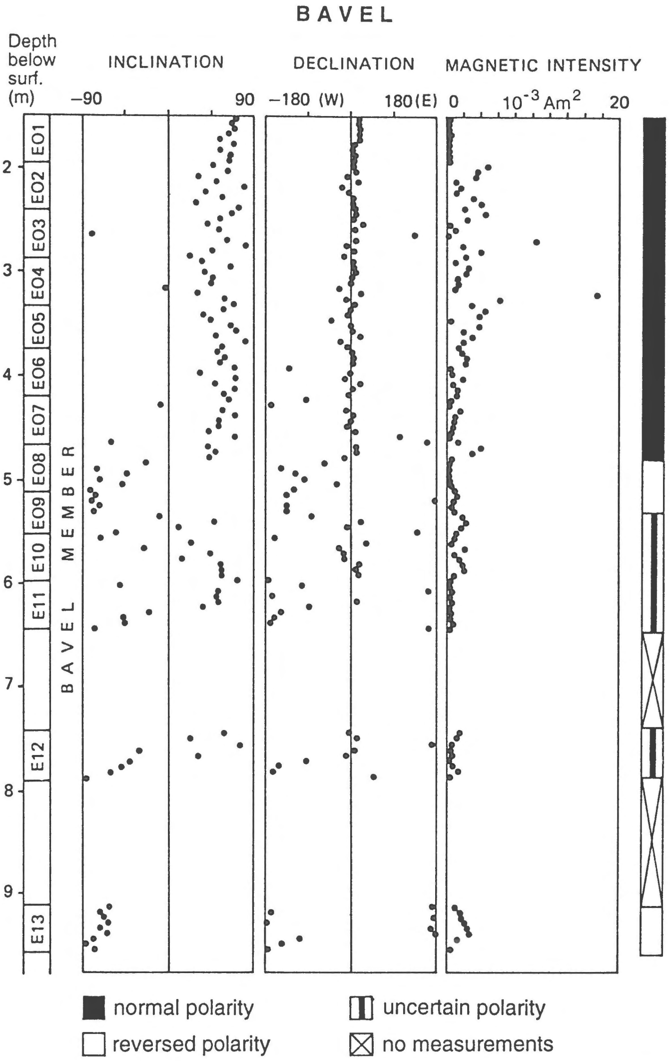

Figure 5. Characteristic remanent magnetization (ChRM) of the Bavel Mbr (upper part Kedichem Fm) at Bavel after alternating field demagnetization (at 15 mT) (location in Figures 1 and 3). E01 to E13 refer to the sampling device used in the field. The magnetic reversal from reversed to normal between 5.25 and 9 m is attributed to the lower boundary of the Jaramillo subchron.

the magnetization in this interval is high and thus the polarity uncertain. The reason could be the low magnetization intensity observed at this level. This low intensity could point to a transitional behaviour of the geomagnetic field or to a delayed acquisition of NRM during the reversal (Van Hoof & Langereis 1991).

The clastic nature of the investigated section and the absence of organic material, soils or roots point to a relatively rapid infilling of the abandoned channel. The presence of current ripple and climbing ripple lamination also indicates a high sedimentation rate. According to Thompson & Oldfield (1986), a reversal may occur in less than 5000 years. The lithology and sedimentology of the investigated section seem to indicate that the polarity reversal occurred within a few thousand years.

Gilze

The upper 3 m of the Gilze Member (Kedichem Fm) have been investigated in a clay pit at Gilze (Figures 4, 6). This exposure displays a heterogeneous lithology with two backswamp clay units separated by a fluvial sand layer. The upper clay layer (1.3–2.2 m below the surface) is predominantly massive and contains one intercalated peat layer. This clay, which was pollen-analytically dated as Waalian A and Waalian B (Kasse 1988), has been intensively cryoturbated during the Weichselian. A fine (100–250 μm) sand layer (2.2 to 2.6 m below the surface) separates the cryoturbated clay from an underlying non-cryoturbated, compact, crumbly clay. This lower clay contains humic to peaty soil horizons and it has been interpreted as a backswamp deposit on top of a fluvial fining-upward sequence. Pollen analysis revealed that the lower clay was deposited in a backswamp environment during the Waalian interglacial (Kasse 1988). The relatively high content of thermophilous trees of dry habitats (*Carpinus, Quercus*) points to a Waalian A age (Zagwijn 1963; Zagwijn & De Jong 1984: boring Leerdam).

Because of the scatter of the magnetization directions from the VGP latitudes, we infer that the cryoturbated upper clay, the sand layer and the top samples from the lower clay possess an uncertain polarity (Figure 6). The lower part of the section is reversed. The somewhat scattered results in the lower part may be explained by the presence of peaty soil horizons. The change in polarity occurs approximately 60 cm below the base of the cryoturbations. The influence of the latter on the remanent magnetization will be discussed further on.

Meerle

The investigated 3-m-thick clay in clay pit Desta south of Meerle (Figure 3) belongs to the upper part of the brackish to fresh-water tidal, estuarine Turnhout Member (Tegelen Fm). The high content of thermophilous trees (especially *Alnus*) and the presence of the megasporangia of *Azolla tegeliensis* unmistakably point to a Tiglian age for the Turnhout Member (Kasse 1988). Since the underlying Beerse Member was most probably formed during the Tiglian C4 cold period, the (upper part of the) Turnhout Member was deposited during the interglacial Tiglian C5.

Well-developed periglacial load casts (cryoturbations) of Weichselian age have been found at the sampling site to a depth of 1.1 m below the top of the Turnhout Member. The section shows a normal polarity in the upper, and a reversed polarity in the lower part (Figure 7). The change in polarity in the 50 cm transition zone is just below the base of the cryoturbation structures. To appraise the influence of the cryoturbations on the remanence, a second profile (J. Hus, unpublished) was examined at Meerle close to a large sand-filled periglacial load cast. The same trend was noticed in this new profile: the inclination appears to be reversed in the bottom part and becomes positive in the upper part.

Meerle Slikgat

Approximately 4 km N of Meerle (Figure 3), the Turnhout Member could be sampled more completely in borehole Meerle Slikgat (Figure 8). Not only the upper part, but also the lower part of the member is fine-grained here, so that changes in the magnetic field during the deposition of the member have been more faithfully recorded. In contrast to clay pit Meerle (Figure 7), the Turnhout Member in Meerle Slikgat is covered by the early Pleistocene Gilze Member (Kedichem Fm). Therefore, the Weichselian periglacial processes did not influence the magnetic polarity of the member. Linking up with the lower part of Meerle (Figure 7), the uppermost sample of the member at Meerle Slikgat has a negative inclination, indicating a reversed geomagnetic field (the declination is unknown). The lower part of the member exhibits low-angle and positive inclinations (uncertain to normal polarity). The member is separated from the underlying Rijkevorsel Member (Tegelen Fm) by a clear erosional boundary, expressed by a thin layer of medium sand (250–500 μm). Pollen analysis indicated that the Turnhout and Rijkevorsel Members were formed during two periods with a high sea level and a warm temperate climate, probably the Tiglian C5 and C3 respectively (Kasse 1988, 1990).

In contrast to the lower part of the Turnhout Member, the two samples from the Rijkevorsel Member have a negative inclination, which points to a reversed polarity (Figure 8). The period during which the transition in the magnetic field from reversed to normal occurred is not registered in the sedimentary sequence, because of erosion at the base of the Turnhout Member. This erosion must have occurred after the deposition of the Rijkevorsel Member but before the sedimentation of the Turnhout Member, that is during the Tiglian C4 or during the start of the Tiglian C5. Deposits of the cold Tiglian C4 period (Beerse Mbr) are missing in this borehole. More to the south, such deposits are locally preserved (Figure 3), but they could not be analysed paleomagnetically, because of their predominantly sandy character.

Achtmaal

A sequence comparable to Meerle Slikgat, consisting of the Turnhout Member erosively overlying the Rijkevorsel Member, has been investigated in bore-

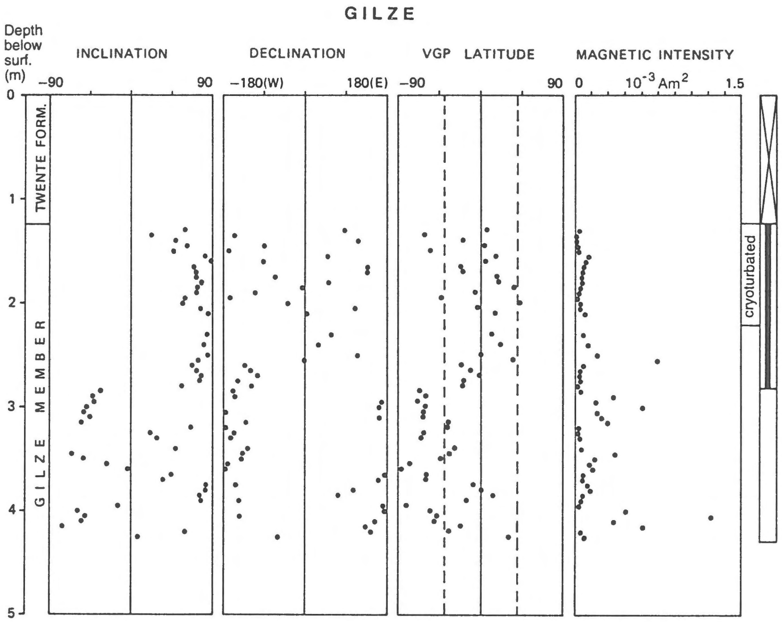

Figure 6. ChRM of the Gilze Mbr (Kedichem Fm) at Gilze after alternating field demagnetization (20–30 mT) (location in Figures 1 and 4, legend in Figure 5). The positive and uncertain inclinations in the cryoturbated upper part of the Gilze Mbr are caused by Weichselian remagnetization.

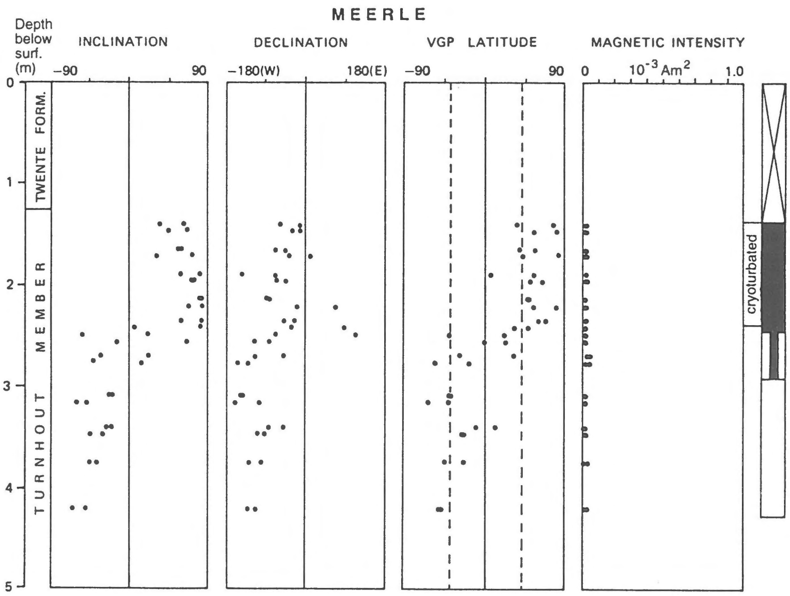

Figure 7. ChRM of the Turnhout Mbr (upper part Tegelen Fm) at Meerle after alternating field demagnetization (30 mT) (location in Figures 1 and 3, legend in Figure 5). The normal polarity in the cryoturbated upper part of the member is an effect of Weichselian remagnetization.

hole Achtmaal (Figures 1, 9). The Turnhout Member is characterized by predominantly positive inclinations (normal polarity). The negative inclinations, characteristic for the upper part of the member in Meerle and Meerle Slikgat, are missing, because 2 to 3 m of the member have been eroded here. Like in Meerle Slik-

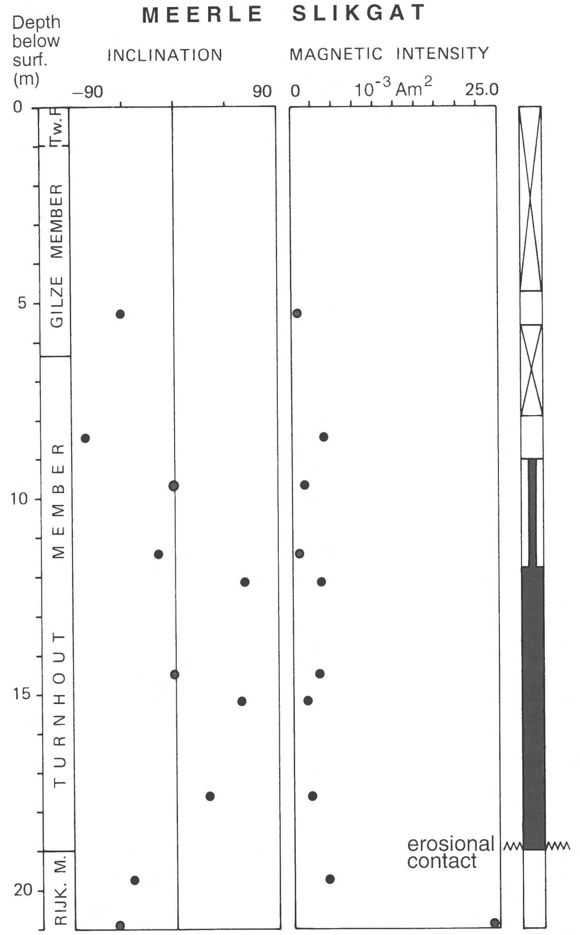

Figure 8. Inclinations of the Turnhout Mbr and Rijkevorsel Mbr (Rijk. M.) (Tegelen Fm) at Meerle Slikgat after alternating field demagnetization (20 mT) (location in Figures 1 and 3, legend in Figure 5, Tw.F. = Twente Fm). The low-angle and positive inclinations in the lower part of the Turnhout Mbr are attributed to the Olduvai subchron.

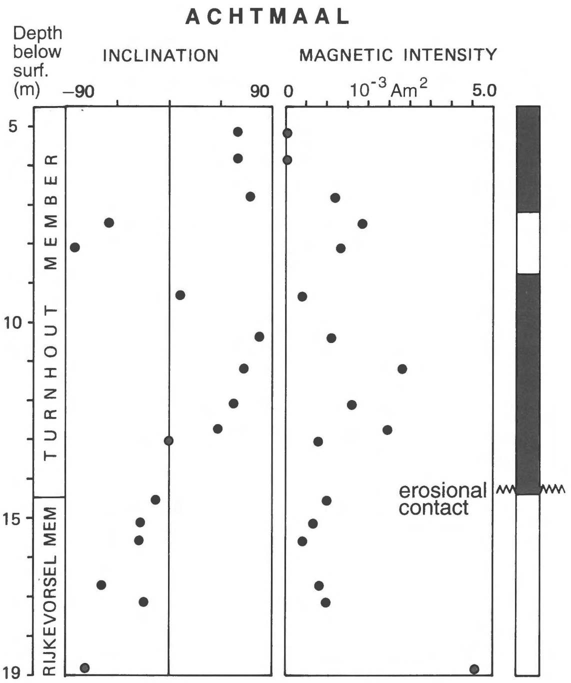

Figure 9. Inclinations of the Turnhout and Rijkevorsel Members (Tegelen Fm) at Achtmaal after alternating field demagnetization (30 mT) (location in Figure 1, legend in Figure 5). The positive inclinations in the Turnhout Mbr are attributed to the Olduvai subchron. The change from reversed in the Rijkevorsel Mbr to normal in the Turnhout Mbr is situated at the erosional contact between the two members.

gat, the Rijkevorsel Member is clearly characterized by negative inclinations (reversed polarity).

Influence of Weichselian periglacial processes on the polarity

The magnetic zonation in the Turnhout Member is far from clear and it is therefore important to know if periglacial processes have influenced the original remanence. Hus et al. (1993) established that the deformed strata adjacent to an ice-wedge cast reveal large directional magnetization changes, not induced by the geomagnetic field, but caused by periglacial disturbances of the sediment. Periglacial phenomena are common in Weichselian deposits in the Netherlands and Belgium (Vandenberghe 1983, 1985). In places where the net accumulation was low during the Weichselian, the periglacial structures have formed in the underlying pre-Weichselian units. Such a situation is often present in western Noord-Brabant and northern Belgium, where the Weichselian Twente Formation directly overlies early Pleistocene sediments, separated by an unconformity which formed during the middle and late Pleistocene.

The influence of the Weichselian periglacial processes and resulting cryoturbation structures (ice-wedge casts, convolutions) on the magnetic polarity is clearly demonstrated in the sections Gilze (Figure 6) and Meerle (Figure 7). Detailed field drawings of these periglacial structures in Gilze, Meerle and additional sites have been presented in Kasse (1988). In Meerle and Gilze, where the clay layers of the Turnhout Member and the Gilze Member are directly overlain by the Twente Formation, the change from reversed to normal polarity is accompanied by the presence of cryoturbations in the upper part of the clays. Such a normal polarity in the cryoturbated upper part of the Turnhout Member has also been reported by Van Montfrans (1971: 44, 45) in the clay pits St. Franciscus and De Toekomst near Turnhout, and later by Hus et al. (1976) in several clay pits near Beerse. In the other sampling locations (Bavel, Meerle Slikgat, Achtmaal), a sand layer is present between the investigated clay bed and the Weichselian deformations, so the deformations could not affect the top of the clay beds. In these latter locations a change in polarity did not occur.

The sections Meerle and Gilze, which both are strongly influenced by cryoturbation, reveal changes from reversed to normal or uncertain polarity in different lithostratigraphic units, viz. in the Turnhout Member and Gilze Member respectively. If these polarity changes are attributed to changes of the geomagnetic field during the early Pleistocene, then they could correspond to short-lived reversals, such as the normal polarity subchron Cobb Mountain (1.21 Ma). However, their short duration makes them difficult do detect, especially if they occur in cryoturbated beds. On the other hand, recent investigations (J. Hus, unpublished) of a clay pit at Beerse (investigated previously by Kasse 1988: Beerse Blak) yielded a clear normal polarity in the Turnhout Member from its top down to a peaty horizon, and a reversed polarity below the latter.

The discussion as to whether cryoturbation can affect the earlier recorded polarity of clay layers depends to a large extent on the plasticity of the clay during the cryoturbation. If the involution process was only a rigid displacement and overturning of layers, then part of the original remanence, in particular its polarity, may have survived, but a large scatter in the magnetization directions would result (Hus et al. 1993). If cryoturbation involved the complete liquefaction of the clay, as is the case in Meerle and Gilze, then a reorientation of the magnetic carriers will have occurred according to the then existing magnetic field (remagnetization). Experiments by Vandenberghe & Van den Broek (1982) indicated that oversaturation and excess pore water pressure must have been present during the formation of the Weichselian periglacial load casts (involutions). Under such circumstances, a loss of intergranular contacts in the liquefied layer occurred and a complete remagnetization of the sediment is feasible. In the case of Weichselian cryoturbations of the Tiglian clay layer in Meerle, the reversed polarity acquired during the Matuyama magnetozone was replaced by the normal polarity during the Brunhes period. The polarity reversals in cryoturbated sites like Meerle and Gilze occur some distance (15–60 cm) below the macroscopically visible cryoturbations. It is assumed that remagnetization occurred shortly after the periglacial involution during the melting of the ice-rich top zone of the permafrost. Continued melting of the underlying ice-poor permafrost zone may have resulted in partial remagnetization without load casting.

Discussion and conclusions

Paleomagnetic measurements of the early Pleistocene deposits in the Breda–Turnhout area have given new results concerning the magnetic polarity of the lithostratigraphic units and the correlation of the polarity time-scale with the bio(chrono)zones based on Quaternary climatic changes.

Previously, Van Montfrans (1971), following the ideas of Paepe & Vanhoorne (1970), stated that the Rijkevorsel, Beerse and Turnhout Members are of Tiglian, Eburonian and Waalian age, respectively. The sediments of the later established Bavel Member (pit Bavel) were, like the Turnhout Member, also correlated at that time with the Waalian. In both the Turnhout and the Bavel Member, a polarity transition was observed from reversed to normal, which was interpreted as the base of the Jaramillo subchron in the Waalian. Later, Zagwijn & De Jong (1984) demonstrated that the Bavel Member was not formed during the Waalian, but during the younger Bavelian, which had not been distinguished before. More recently, Kasse (1988, 1993) dated the Rijkevorsel, Beerse and Turnhout Members all as exclusively Tiglian, instead of Tiglian, Eburonian and Waalian, on the basis of paleobotanical results.

These new results clearly demonstrate the problems of geomagnetic dating of discontinuous continental deposits. The puzzling stratigraphic relations have led to erroneous paleomagnetic correlations. For instance, the Turnhout and Bavel Members were regarded previously as Waalian, and the normal magnetozone in these units was thought to correspond with the Jaramillo subchron. However, the new lithostratigraphic and paleoecological results make it probable that different magnetic reversals are involved. In contrast to Van Montfrans (1971), who correlated the reversal in the cryoturbated upper part of the Turnhout Member with the base of the Jaramillo subchron, this reversal is explained in this study (exposure Meerle) by Weich-

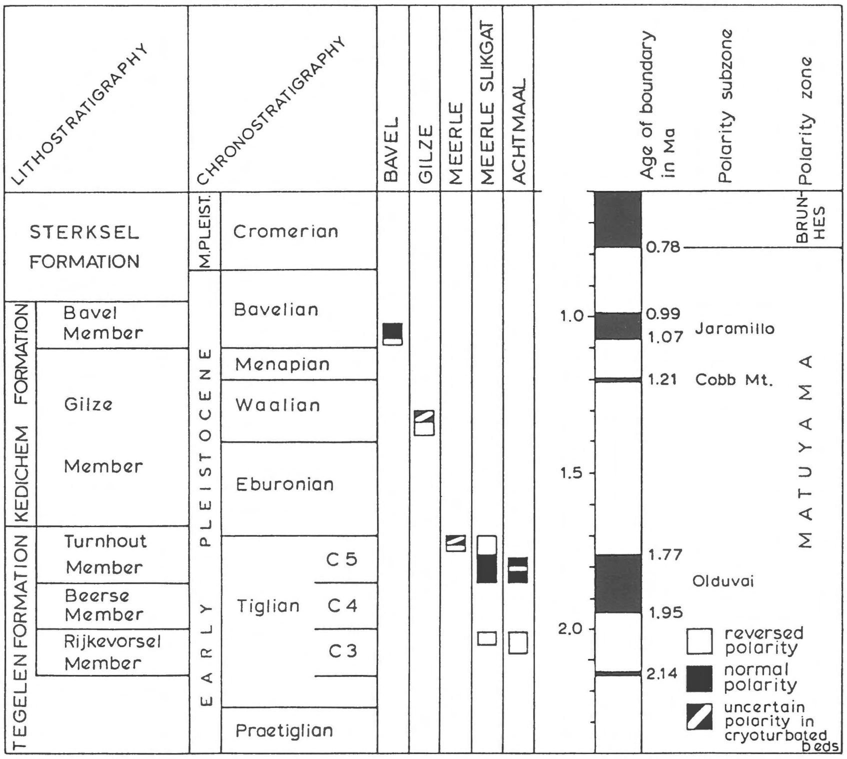

Figure 10. Litho-, chrono- and magnetostratigraphy of the early Pleistocene deposits in the Breda–Turnhout area (locations in Figure 1). Pliocene–Pleistocene boundary following Zagwijn (1985, 1989); magnetostratigraphy after Cande & Kent (1995). Normal polarity at 2.14 Ma is Réunion subchron.

selian remagnetization. The results of the present geomagnetic investigations are summarized in Figure 10.

The Bavel Member (upper part Kedichem Fm) was formed during the Bavel Interglacial (pollenzone Bv3b) of the Bavelian (Zagwijn & De Jong 1984; Kasse 1988). The polarity reversal from reversed in the lower part to normal in the upper part (Figure 5) is interpreted as the lower boundary of the Jaramillo subchron. The transitional directions have been recorded in the sedimentary sequence, because of the high sedimentation rate in the abandoned river channel. Normally, transitional or intermediate directions are rarely found in continental deposits because of the short duration of the reversals (2–10 kyr) and the fragmentary preservation of the deposits.

These results concerning the Bavel Member are generally in agreement with previous investigations by Van Montfrans (1971) and Zagwijn & De Jong (1984). Van Montfrans (1971) described a 5.9-m-thick clay bed with a normal polarity. Later, Zagwijn & De Jong (1984) correlated this clay bed with the Bv3b pollenzone of the Bavel Interglacial. According to the latter authors the somewhat older Bavelian pollenzone Bv2 contains the transition from reversed to normal at the lower boundary of the Jaramillo subchron. The new results presented above (Figure 5) seem to indicate, however, that the lower boundary of the Jaramillo (1.07 Ma) is equivalent to pollenzone Bv3b.

The Gilze Member (Kedichem Fm) was deposited during the Eburonian, Waalian and Menapian (Kasse 1988). Only the Waalian part could be investigated magnetically, because of its clayey texture. It shows the reversed polarity of the Matuyama chron (Figure 6). The normal or uncertain polarities in the cryoturbated top of the clay bed are explained by Weichselian remagnetization.

The Turnhout Member (upper part Tegelen Fm) reveals a normal polarity in its lower and middle parts (Figures 8, 9), a reversed polarity in its upper part (Figures 7, 8) and again, due to Weichselian cryoturbation, a normal polarity near its top (Figure 7). These results partly confirm earlier work of Van Montfrans (1971: pits De Toekomst, St. Franciscus) and Hus et al. (1976: St. Franciscus, Dekt). However, these authors did not find the normal polarity in the lower and middle parts, probably because these parts are missing in the Belgian clay pits, as the member wedges out to the south.

The change from normal to reversed in the Turnhout Member is interpreted as the upper boundary of the Olduvai subchron (1.77 Ma). Since the member was deposited during the warm temperate Tiglian C5 (Kasse 1988), it is concluded that the Olduvai upper boundary occurs within the (lower part of the) Tiglian C5. This position is different from earlier results obtained in Netherlands Limburg (Van Montfrans 1971), where the boundary seems to be present in clay beds of late Tiglian or even early Eburonian age. In the Lower Rhine Embayment, Arias et al. (1984) also observed the Olduvai subchron in the youngest Tiglian (Tiglian C5–6). Our results, however, are in good agreement with the recent results of Hallam & Maher (1994), who found a reversed polarity in clay beds of Pastonian (= late Tiglian) age. Their conclusion is that the late Tiglian sediments fall within the Matuyama chron.

The polarity change from reversed to normal in the cryoturbated uppermost part of the Turnhout Member (Figure 7) is explained by remagnetization of the clay during the Weichselian. The melting of the ice-rich topzone of the permafrost probably led to a complete liquefaction of the sediment and to associated periglacial involutions. The original Matuyama reversed polarity was destroyed and replaced by the normal polarity of the Brunhes chron. This polarity change was also described by Van Montfrans (1971: pit St. Franciscus). He regarded this reversal as an original magnetic signal and interpreted it as the lower boundary of the Jaramillo subchron. However, it is striking that the depth of the reversal, approximately 1.2 m below the top of the clay, coincides with the Weichselian cryoturbation depth of approximately 1 m reported by Paepe & Vanhoorne (1970). Therefore, in our opinion this reversal must be attributed to remagnetization during the Brunhes period.

The Beerse Member (Tegelen Fm) has not been investigated here because of its sandy nature in the pits which are open now. Pollen analysis of the peat beds in the member indicates a cold climate, which was correlated with the Tiglian C4 (Kasse 1988, 1993; Vandenberghe & Kasse 1989). According to Van Montfrans (1971: pit De Toekomst), the lower part of the member shows a reversed, and the upper part a normal polarity. In our opinion this reversal may represent the base of the Olduvai subchron (1.95 Ma), which base is then time-equivalent to part of the Tiglian C4 (Kasse 1993).

The upper part of the Rijkevorsel Member (Tegelen Fm) is of reversed polarity (Figures 8, 9). This result was also obtained by Van Montfrans (1971), who investigated the Rijkevorsel Member in two Belgian clay pits (De Toekomst and St. Franciscus). Kasse (1988) dated the Rijkevorsel Member on paleoecological grounds as Tiglian C3. According to Zagwijn (1963), fluvial sedimens at the type locality of the Tegelen Formation in the Dutch province Limburg (pit Russel-Tiglia-Egypte) were also deposited during the Tiglian C3. These sediments, however, revealed a normal polarity (Van Montfrans 1971), which was interpreted as the Réunion subchron (Zagwijn 1989: his Figure 20). It is concluded that the reversed upper part of the Rijkevorsel Member was probably deposited during the late Tiglian C3, after the Réunion and before the Olduvai subchron.

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