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Paleoseismic investigations along the Peel Boundary Fault: geological setting, site selection and trenching results

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Abstract

On the basis of a multidisciplinary approach we have unraveled the palaeo-earthquake history of a trenched section across the Peel Boundary Fault. The area shows at present one of the largest contrasts in relative motion on both sides of the fault on the basis of repeated levelling. The geological record for the last 25 thousand years, recovered in the trench, shows evidence of two heavy earthquakes (moment magnitude between 6.0 and 6.6), that occurred in a relatively short timespan around 15 thousands years ago. A third less severe event occurred somewhere in the mid Holocene. The time interval between the two large events is in the order of 1500 years, an interval comparable to that between the last volcanic explosions in the nearby Eifel area. Both records together seem to suggest a relation between large-scale faulting and volcanic activity in the nearby Eifel area, but this interpretation is based on one trench only and should be tested by opening more trenches in the zone that is assumed to be affected by these large events.

Keywords: Peel Boundary Fault, Palaeoseismology, Trenching, Morphotectonics, Recurrence time.

Introduction

Growing awareness, that the expression of earth dynamics is not restricted to specific regions on earth like plate boundaries, but can be observed in intraplate settings as well, drives the research on neotectonics. One of the most important applied aspects of neotectonics forms the research on the timescales at which stress and strain operate and in particularly become expressed in extreme events affecting the society. Within this field the shorter timescales are for obvious reasons most relevant for our society. Documentation by instrumental records are limited to decadal- or at the best centennial times; the geological record forms the only storehouse to explore data beyond the instrumental time series. Palaeoseismic research can document by trenching possible major earthquake events beyond the limits of the historical records. The Dutch contribution to the EC project 'Paleosis' has been focussed to improve the dating possibilities of major prehistoric events. We applied a multidisciplinary approach to constrain the timing of the landmarks of postulated prehistoric large earthquakes and estimated their magnitude and recurrence time. Site selection for exploration of this geological record requires careful preparation. Geomorphic studies, subsidence analysis, structural and seismostratigraphic/tectonic studies were integrated in order to delineate a number of potential sites.

Because the Roer Valley Graben (RVG) forms the most prominent Caenozoic tectonic feature in the Netherlands and is considered an active fault system, it was decided to focus on this tectonic domain (Fig. 1). Two of the most important natural earthquakes in

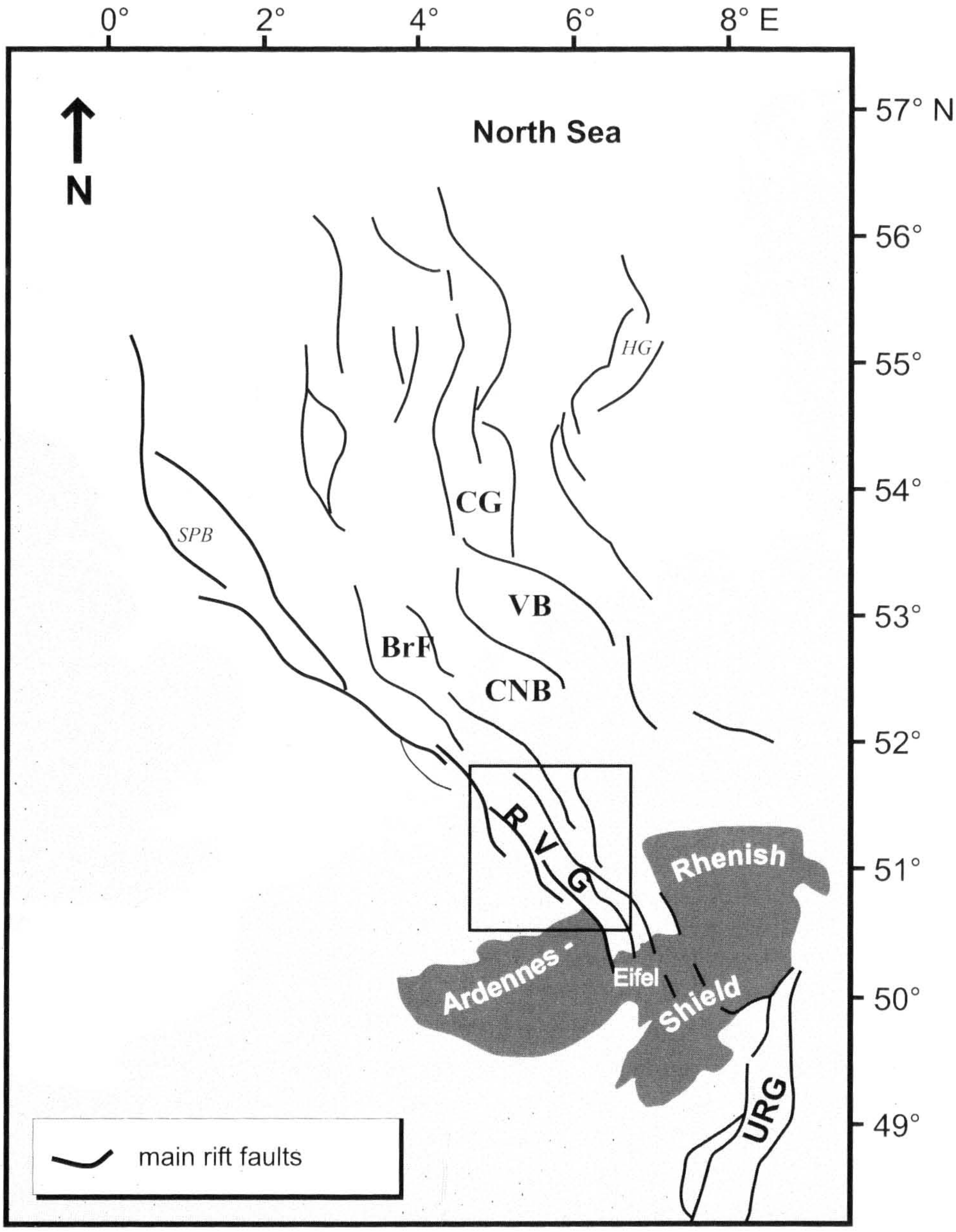

Fig. 1. Structural elements within the North Sea Basin. Main depocentres are indicated: SPB = Sole Pit Basin, CG = Central North Sea Graben, BrF = Broad Fourteen Basin, VB = Vlieland Basin, HG = Horn Graben, CNB = Central Netherlands Basin, URG = Upper Rhine Graben, RVG = Roer Valley Graben. SPB and HG are of subordinate significance during the Late Neogene, whereas the URG, RVG, CNB, VB, and CG show anomalously thick Quaternary sediment thicknesses (Caston, 1977). This pattern marks the dogleg structure. Note that the fracture zones bounding the RVG are fanning towards the northwest. The recent crustal stressfield has in this area a NW-SE orientation (Gruenthal and Stromeyer, 1994).

the Netherlands this century (Uden, 1932; Ml=5.5 and Roermond, 13-4-1992; Ml=5.8) occurred in this domain, but several earlier events have been documented within the RVG (Berz, 1994).

Earlier paleoseismic studies (Camelbeeck et al., 1998) focussed on the Bree fault, a part of the southern principal displacement zone, known as the Feldbiss fault zone, of this RVG (Fig. 2). The Dutch part of the project concentrates on the northern principal fault of the RVG, known as the Peel Boundary Fault (PBF).

As event dating was an important aim in this study, we selected potential sites along the PBF close to the Maas river (which crosses the PBF), where river terraces are present that can be accurately dated. The outcome suggests a dynamical link between high magnitude faulting events and volcanic events in the nearby Eifel region, an aspect that is discussed.

The geohistory and structural setting of the Roer Valley Graben

Rock, when deformed in direct shear develops narrow shear zones in which most of the deformation takes place. Tchalenko (1970) has shown that the great similarities in structural deformation at various scales ranging from microscopic to regional (earthquake faults) can be interpreted in terms of similar mechanisms. This observation stresses the value to take the regional setting into account when evaluating local observations made in a trench.

The Roer Valley Graben (RVG) came into existence in Late Permian to Early Triassic times during a rifting phase as part of the transition from Late Variscan to Early Alpine orogenies (Ziegler, 1990, 1994; Winstanley, 1993; Geluk et al., 1994). The southern bounding

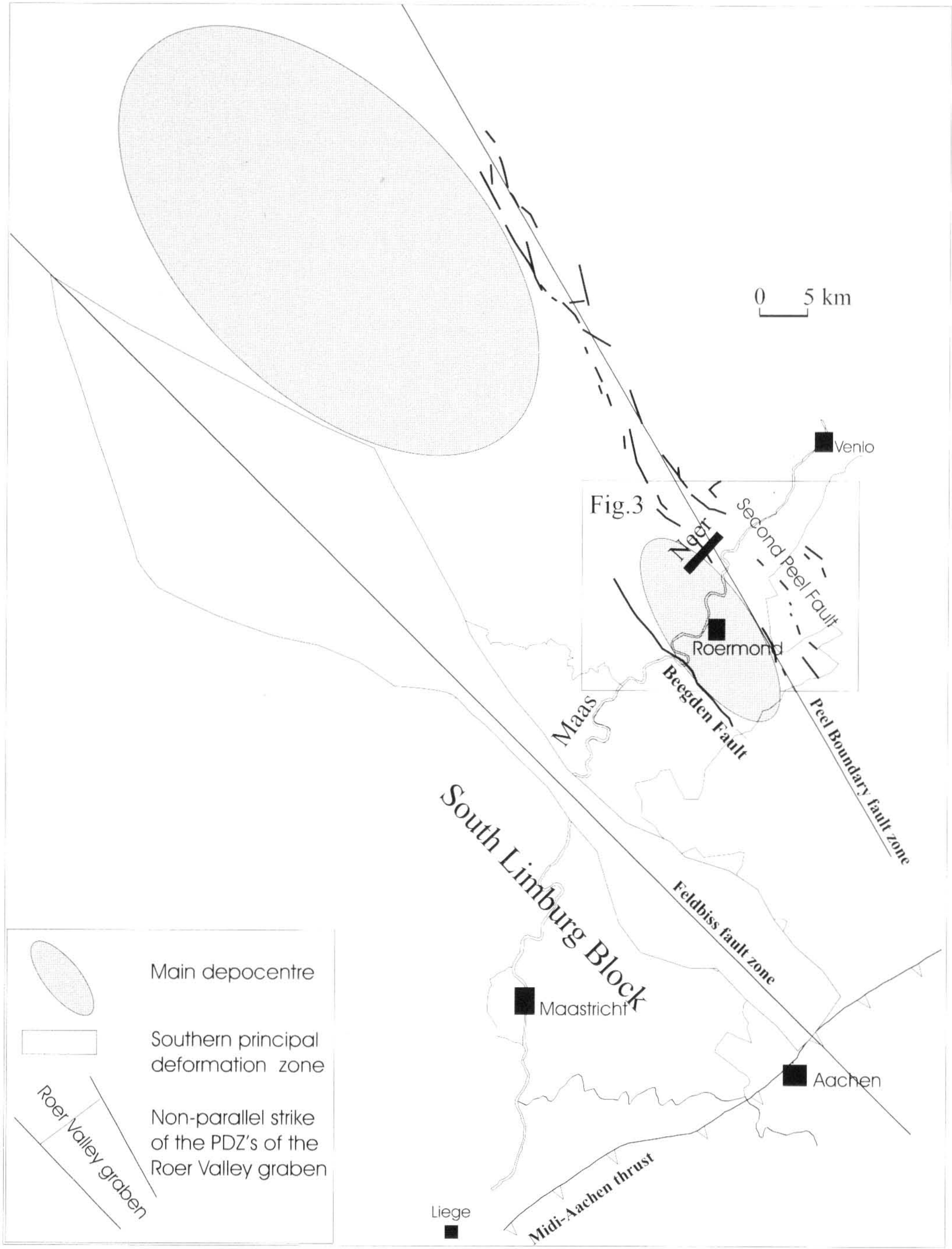

Fig. 2. The Roer Valley Graben bounding principal displacement zones (PDZ) have a different surface expression: the northern PDZ is much narrower than the southern PDZ. Two Neogene depocentres can be recognised from isopach maps. Position of this figure is marked in Fig. 1.

lineament of the RVG can be traced from the east coast of England into the Eifel (Fig.1). Deep seismic data show that this is a crustal-scale feature affecting the upper mantle (Chadwick and Pharaoh, 1998). The northern bounding lineament is of similar length in map view; the deep seismic reflection profile crossing this structure in the studied area (Remmelts and Duin, 1990) is not as clear in depth as the one mentioned above. The similarity in dimensions suggests that this also is a crustal scale feature.

The post-rift phase after the initial Early Triassic rifting resulted in a thick unit of Triassic, Jurassic and Early Cretaceous, mainly clastic, sediments, which were partially removed during a Late Cretaceous inversion. Upper Cretaceous chalks are present on the shoulders of the graben. At the end of the Cretaceous the RVG started to subside again. Uplift of the RVG during Upper Eocene times is reported by Zijderveld et. al. (1992). The subsidence of the Roer Valley Graben resumed in Late Oligocene times, it accelerated strongly in the Middle Miocene and lasts until the present day (Van den Berg, 1994). E.g long-term average subsidence values, observed along the Peel Boundary Fault increase from 0.01 mm/yr during the Late Oligocene via 0.05 mm/yr during the Miocene and Pliocene to 0.08 mm/yr during the Quaternary (Geluk et al., 1994).

The graben is one of a suite of depocentres underlying the southern North Sea basin with proven strong Late Neogene activity. These form an assemblage with all characteristics indicative for large-scale transtensional deformation within the European plate. E.g. the sub-parallel Upper Rhine Graben and Central North Sea Graben, separated by three overstepping elements below the Dutch onshore area: i.c. the RVG, the Central Netherlands Basin (Zuidersea Basin) and the Vlieland Basin show a dogleg-like arrangement. Such an angular relationship of these major lineaments strongly resembles the angular relations between structures that tend to develop in left lateral shear under ideal conditions (Fig. 1).

This indicates that the major fractures form a strain pattern related to NNW oriented first order stress, being perpendicular to the Alpine thrust front and therefore probably dominated by this front within this region. Currently the stressfield orientation parallels the overall NW-SE orientation of the Peel Boundary Fault (Grunthal & Stromeyer, 1994). This orientation is not favourable for the postulated strike-slip motion along the Peel Boundary Fault but could generate such a type of motion along the Feldbiss fault system that differs in overall average orientation. There is strong paleogeographic evidence that the stressfield experienced a few, time-dependant significant fluctuations over the last 4 Ma. (Van Dijke and Van den Berg, in prep.); if these fluctuations are cyclical with fractal characteristics, thus occurring both at short-, intermediate-, and long time scales, is yet unknown. Morphological analyses of lineaments which have developed over the last few hundred thousands years do suggest strike slip motion did occur along the Peel Boundary Fault also (Van den Berg et al., 1994) which would imply at least young Pleistocene stressfield re-orientations; such motion is not observed by instrumental record in modern times.

The surface area of the RVG is in the order of 100X40 km and isopachs of the Neogene infill of the RVG show that its geometry is controlled by two (sub)depocentres (Fig. 2).

The morphology of the fault system

The PBF fault system consists of a large number of en echelon faults and towards the northwest the PBF becomes extremely complex because of wrench faulting. The southeastern part (near Roermond) is the less complex part of the structure. There the PBF is composed of some subsidiary faults (Fig. 2): the main fault of the PBF with the antithetic Beegden Fault and the synthetic Second Peel Fault, in addition to some minor antithetic faults observed in cross section. Of these faults only the Second Peel Fault and the Peel Boundary Fault have stepped surface expression, whereas the Beegden Fault markedly affects the (Late) Pleistocene to Holocene geology but without a traceable terrain step. The position of these fault elements next to the Roermond- (sub)depocentre suggests a mechanical relationship. The surface expression of the fault zone along its length is remarkably discontinuous. It shows a large number of elements arranged into a braided pattern, the nodes of this pattern have typically a spacing of about 10 km, the surface length-width ratio of the embraced blocks amounts to about 4 to 1 (Fig. 2).

Trench-site selection

We considered site selection as a very important step in the whole of the project. Several aspects were considered.

1) A simple fault system is considered as important, because seismic slip is concentrated on a small number of faults displaying the most evident traces of this slip. Also, if traces of large earthquakes are found, a minimum fault dimension is required. For local magnitudes higher than 6.3, which is the present estimate of the maximum magnitude in seismic risk studies in the region (De Crook, 1996), the lateral dimension required is at least 10-20 km.
2) The historic record (up to 117 yrs) of surface deformation of the RVG shows a significant variability between various sub-units (Van den Berg et al., 1994). The largest contrast in estimated vertical velocities along the PBF is found around the Roermond depocentre. Seismic profiles confirm that the maximum displacement is shown at the PBF and less at the Beegden Fault and Second Peel Fault. In 1992 the Roermond earthquake with a moment magnitude of 5.42 occurred at the PBF in this region. Based on the above-mentioned criteria, this region was selected for further study, which concentrated on the geomorphologic parameters.
3) Geomorphic terrain analysis basically comes down to the identification of terrain steps or -zones of knicks, together with identifying the geologic process responsible for shaping of these

earth surface irregularities. Earth surface processes that significantly affect the landscape in this study area are related to: climatic conditions, subsurface processes, fluviatile processes, lithology and anthropogenic processes.

On the basis of the comparison of morphological lineaments, coinciding with the seismically derived fault map and the expression on the aerial photographs, a number of sites could be chosen along the eastern part of the PBF. These sites are situated on river Maas terraces. The formation and abandonment of these terraces is constrained well in time (Fig. 3), (Van den Berg, 1996).

One may assume that a river terrace forms a continuous surface during its formation. Any vertical displacement of this surface post-dates the age of the terrace. This provides us with a maximum value for the time-lapse of fault-scarp formation. In this case Maas terraces of various ages and lithology (gravel-rich sands or sand facies) are potentially suitable. Although the gravel terraces are among the youngest available terraces, the sandy facies is preferred, because any post-depositional deformation of sedimentary structures will be easier to detect in fine-textured deposits.

Important climate-related processes, like the deposition of aeolian sandsheets, are restricted to the glacial periods of the Pleistocene. This holds that Pleistocene landscapes can be regarded as being 'frozen' for a period of the last 11.500 years at least. During the Holocene only the riverine (and coastal) areas experience large-scale changes in the landscape.

4) Shallow drillholes, over the most promising site (based on the presence of a small fault scarp), suggested a 1.5 meter offset in the younger sediments. However stationary or upwelling groundwater occurred at some sites. Groundwater distribution maps revealed that most selected sites would suffer from this problem. In fact only one location would probably be dry enough to avoid additional pumping. The latter would have raised the costs of trenching tremendously.
5) Geophysical investigations on this site (see: Demanet et al., this issue) showed clearly distinguishable discontinuities across the presumed fault plane. This determined the position and the length of the trench.

Taking all these criteria and evidence together, a site near the village of Neer along the PBF was selected. The site is located on a Maas river terrace (Fig.3). This terrace was abandoned by the river at the end of the Weichselian Late Pleniglacial. The surface shows along the fault-trace a vertical offset of about 1.3 m (Camelbeeck, 2000)

Local stratigraphy and chronology

An important aspect of this study was to constrain as closely as possible the timing of paleoseismic events indicated by the displacement of stratigraphic units, this requires a good understanding of the local stratigraphy. To this we first present here an extended description of the sequence of events leading to this stratigraphy. All chronology of the discussed features have been summarised below in figure 8. The local stratigraphy reflects the palaeo-landscape development within a Late Pleistocene sedimentary setting. This is intimately connected with the climate evolution. This climate evolution is characterised by repeated episodes of climatic deterioration at millenium timescales and even higher frequencies (c.f. lake core records). The millenial cycles can be recognized in oxygen isotope records from the Greenland ice-cores (Johnsen et al., 1992; Grootes et al., 1993; Bond et al., 1993), in organic-bearing sediments by palynological records (a.o. Hoek, 1997) and in sedimentary records by sequence stratigraphy (Van den Berg and Schwan, 1996). Although a very detailed comparison can be made between the various climate signals, questions remain about absolute chronology and causal relations between the various events (see also Bjorck et al., 1996). We believe however, that for the relevant time slice and the purposes of this study, we can safely apply the ice-core timescale which is based on annual layer counting. The amplitude of the climate fluctuations is so large and so steep that possible lagging effects are in the order of decades, which is far beyond our possible resolution.

The age of the various stratigraphic units can be inferred from different techniques. We distinguish between age approximations based on physical techniques (radio-carbon dating , and Optically Stimulated Luminescence (OSL)) and those that are based on correlation. Traditionally the physical techniques are referred to as 'absolute dating' and the other as 'relative'. Within the relevant time-slice of the exposed series (the Late Pleniglacial through the Holocene), both approximations are applicable but, providing the right correlation, the relative dating may be more accurate than the absolute dating. This is due to the fact that the reference records for the relative dating, being the Greenland ice-cores, allow for annual layer counting. This in contrast to the 'absolute' dating tools, the results of which suffer from confidence limits covering time slices in the order of several thousands of years.

The local floodplain-landscape underwent profound changes during the transition of a glacial to an interglacial. During a glacial the floodplain in general

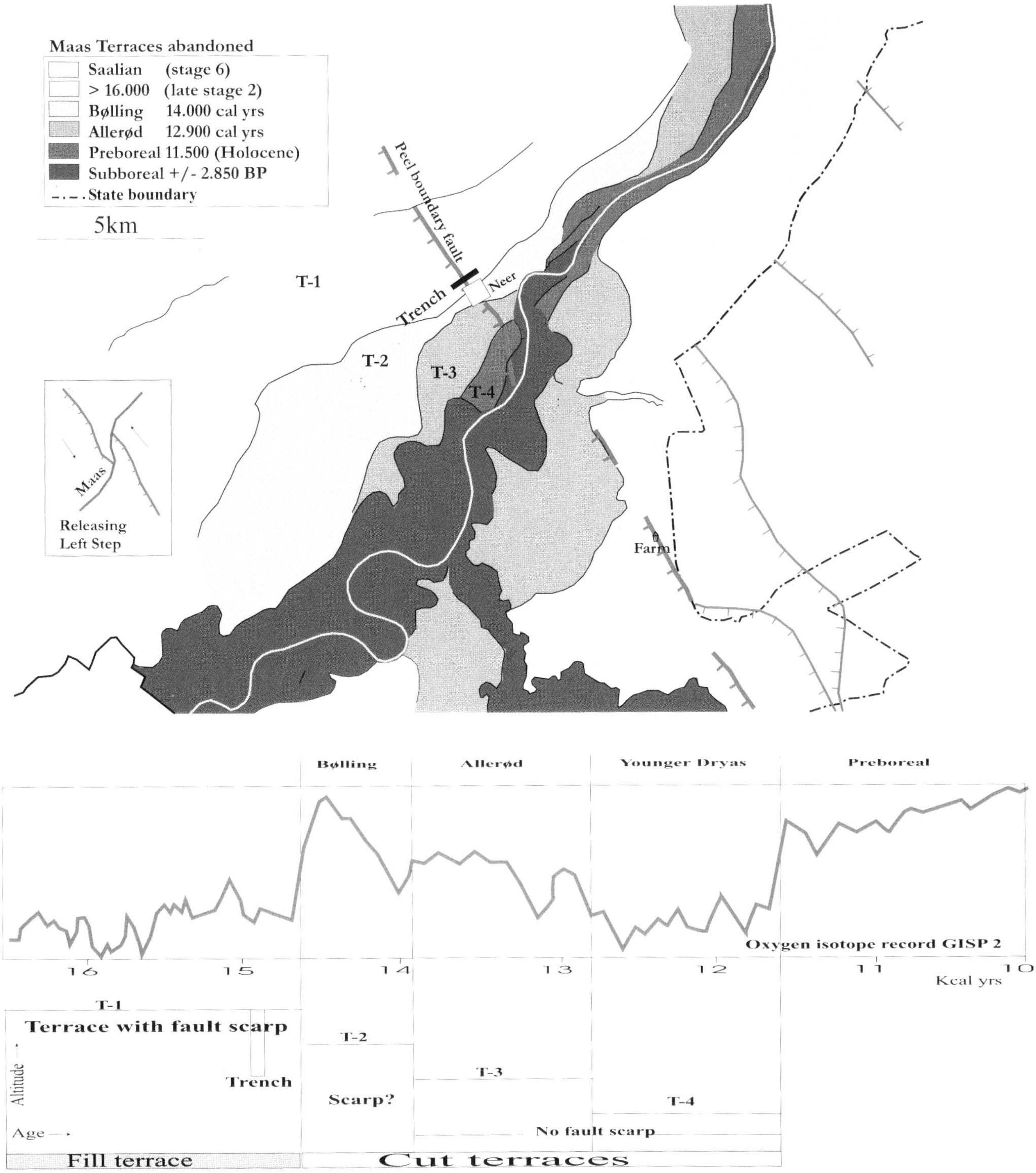

Fig. 3. The location of the trench at the Peel Boundary Fault crosscutting Maas river terraces near the village of Neer. Terraces are shown in different shades of yellow and green. They have been numbered in descending altitudinal order from T-1, T-2, T-3, T-4. The red lines mark the fault trace with the ticks towards the hanging wall. The lower panel shows the chronology of Late Glacial terraces, the oxygen isotope record serves as a proxy for the climate evolution in that period. Position of this figure is marked in Fig. 2.

is raised due to overloading of the system with bedload. Sediment aggradation exceeded subsidence even in the deepest part of the graben. In addition to this, locally floodplain raising was enhanced by the presence of a break in the rivergradient from 0.75 m/km to 0.25 m/km downstream of the Beegden fault. This implies that the regional surface around the Neer site, belongs to a small alluvial fan which formed at the foot of the rising South Limburg.

During interglacial conditions the sediment supply to the river dramatically diminished, initiating a channel downcutting in concert with a steady narrowing of the floodplain width. This apparently was a stepwise process which led to the formation of a series of distinct terraces.

Morphostratigraphy

Within the lower Maas valley we distinguish six topographic levels (Fig. 3). Each topographic level corresponds to a surface of either a terrace proper or a deposit which has buried the original terrace. These levels are not continuous along stream (over a length of over 100 km). We assume that all terrace fragments, that belong to the same topographic level, are of the same age. From this follows that ages of fragments with unknown age can be found by extrapolation of the long profile to a terrace of known age. These terraces are of two types: fill terraces and cut terraces (cf. Merrits et al., 1994). Fill terraces have a surface formed by fluvial aggradation and this original upper level of alluviation may be with or without a cover of aeolian sand. These terraces occupy the highest position in the terraced valley and their planform is characterised by long and gently curved terrace scarps (Fig.3: T1).

The cut terraces have a top surface resulting from downcutting and planation by the river. Cut terraces occur in a flight carved into previously aggraded, mainly sandy sediment. In contrast to the first category, the cut terraces are bound by successions of short and rather strongly curved terrace scarps. In general the age of terraces is taken at the moment of the beginning of floodplain abandonment, this point in time is marked by the age of the earliest infill of fossil channels left behind on the terrace surface. The local chronology is based on pollenzones and 14C ages compiled from various studies in the lower Maas valley. The vegetation development in the Netherlands for this period of time is well known. This allowed Hoek (1997) to correlate the pollenzone boundaries with climate transitions observed in the GISP2 ice core (Grootes et al., 1993). This is important while such a correlation calibrates the relative ages of the pollenzones to an absolute level of calendar years. We have used the calendar years in our age model (see below).

Near the trench the PBF shows a clear scarp at the surface of terrace 1 (T-1). At the level of T-2, the PBF runs through the urban area of the village Neer so we have not recognised any scarp on this terrace. No scarp has been preserved on the level of T-3, although on aerial photos the trace of the fault is clearly visible. Apparently at the levels T-3 and T-4 the scarp has been eroded during the formation of these cut terraces. This observation is important to constrain the ante quem age of scarp formation: older than the onset of the Younger Dryas

Lithostratigraphy exposed in the trench

The trench was dug out in the (fill) terrace 1 (Fig.4). It had a length of 60m and depth of 3-4m. The exposed sediments have been logged in great detail making use of: (1) laqueer peels (for example Fig. 4), to clearify sedimentary – and deformational struc-

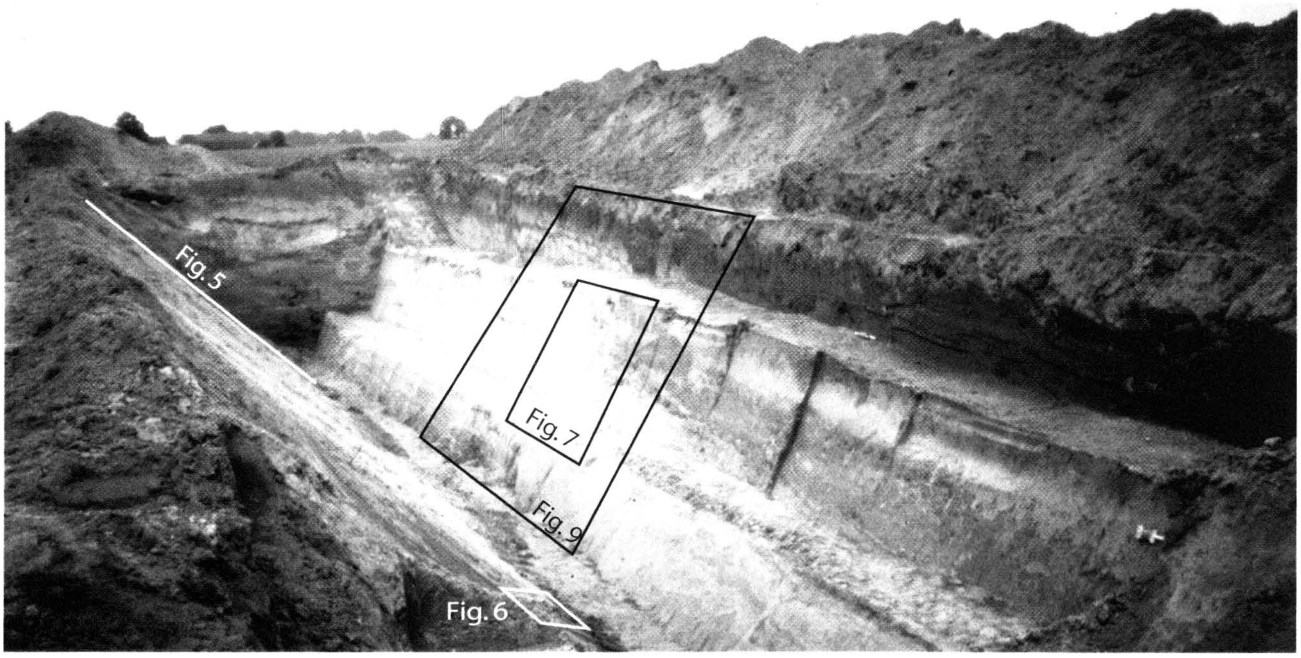

Fig. 4. The trench just after opening sections, showing the NW facing wall. One looks from the hanging wall towards the footwall. Indicated are the positions of in more detail described. (Photo: Th Camelbeeck).

tures, (2) line drawings supported by a 1x1m mesh streched over the exposed surface, (3) selected positions of thin sections (see contribution by Miedema and Jongmans, this issue) and photos.

The great detail of the features used in the final interpratation does not allow for a representation by means of complete drawing. To give an impression of the type and age of the sediments exposed, we will describe and show below a representative section of the undistubed footwall (Fig. 5). A linedrawing of the faultzone will be presented in the section on the faulting history.

The characteristic facies succession of this terrace, shows from bottom to top: a facies sequence grading from fluvial via fluvio-aeolian into aeolian bedding (c.f. Fig.5). All these sands are all relatively fine-grained, well-sorted and devoid of organic layers. The top soil only is grey-black stained by humus of anthropogenic origin (a fimic A horizon). Tabular sedimentary units and horizontal beddingplanes were most prominent in the exposure. We observed, from bottom to top, four marker horizons which could be followed across the faultzone and these were used as a measure for tectonic displacement. The first marker, intercallated within a wet aeolian unit, is a bleached horizontal wave-ripple horizon (marker 1 in Fig. 5). The formation of this bed cannot be correlated with an event in a high resolution climate record, so we have to rely in this case on the OSL dating. These data suggest an age for this bed between 20 and 35 ka (see contribution by Frechen and Van den Berg, this issue)

About 2m higher in the section we recognised a second marker: a deflation horizon and associated frost cracks of up to 1m length (this unit was destroyed by the topsoil, an anthropogenic plough horizon in Fig. 5). The associated cracks have deformed the original horizontal bedding into a wavy bedding. These cold-climate-soil features are commonly interpreted to represent, a characteristic feature of the Late Pleniglacial. The formation of the pebble bed (including the so called Beuningen pebble layer) probably occurred during deep cold spells during this period, that me be part of the Dansgaardt – Oeschger cycles (millennial scale cyclical climate changes). The associated deflation horizons have been recognised in other exposures as well within this fill-terrace (Van den Berg, 1998). The one we have observed in the trench was found close to the land surface and therefore has been interpreted to reflect the last deep cold spell. We correlate this with the last maximum in the 18O values between 16000 and 15500 cal a BP, (ignoring the short amelioration in between). This interpretation is sustained by the OSL outcome (Frechen and Van den Berg , this issue).

On the footwall side of the trench this fine pebble layer was the uppermost lithostratigraphic unit which could be traced across the fault zone. Above this layer all sedimentary structures had been obliterated by soils of Late Glacial, Holocene and anthropogenic origin. This in contrast to the hanging wall side where we observed directly above the Beuningen unit with a 0.20 m thickness of relatively coarse, parallel to cross-bedded (water-lain) sands overlain by a unit of similar thickness which consisted of an alternation of loess and fine-sand laminae (silty sands), interpreted as an aeolian deposit (Fig. 6).

These aeolian deposits are taken as a third marker because they are also well known from another (former) exposure (sandpit near Kappert, about 1 km north in the same terrace but closer to the river (Van den Berg and Schwan in: Van den Berg, 1996). The settling of the loess component of this deposit requires a closed cover of grasses. We tentatively put these conditions during the short-lived optimum between the climatic minimum of the Beuningen bed and the optimum of the Bølling interstadial.

The stratigraphic significance of the cross-bedded fluvial sands is unsolved. They did not occur in Kappert which one would expect if these were Maas river deposits; we interpret from this the local nature of this coarser unit. This interpretation is sustained by an observed southward oriented depositional flow (the Maas river flows towards the north). The tectonic position together with the local occurrence of the coarse sands may suggest its nature as a colluvial wedge associated with a local terrain step. Such a step cannot be related to one of the three main faults observed in the trench (see below). The fourth marker was the B-horizon of the soil discussed hereafter.

Soil stratigraphy

The B–horizon of the soil profile exposed in the trench is markedly vertically displaced (Bts in Fig. 7). Therefore it forms another (the fourth) important marker horizon for the reconstruction of the tectonic events.

The sequence of soil horizons starts with a fimic A-horizon classified as a fimic anthrosol. Its thickness increases on the downthrown side. Below this A-horizon follows a transition into an eluvial horizon, in turn followed by a B-horizon which is expressed by several discontinuous, rubified, argillic horizons, up to several decimetres thick, that are formed by the downward migration (illuviation) of fine clay particles with bonded iron within the host sediment. For the age of this process one generally takes the Late Glacial (Miedema, 1983).

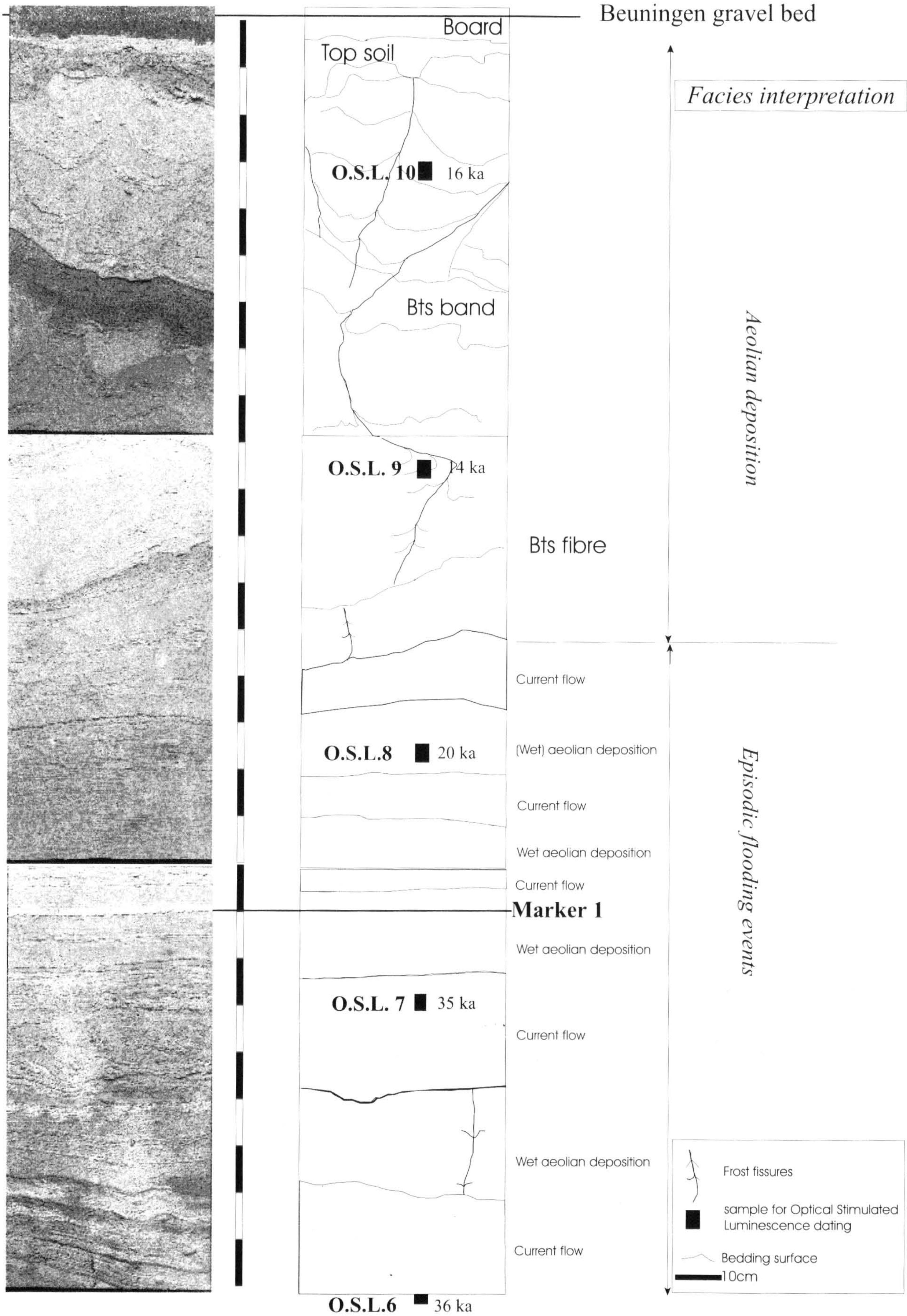

Fig. 5. Undisturbed footwall sediments displayed on lacquer peels with the sedimentary interpretation and position of OSL samples; for discussion on the OSL-dates see Frechen and Van den Berg (this issue). Position of this figure is marked in Fig. 4.

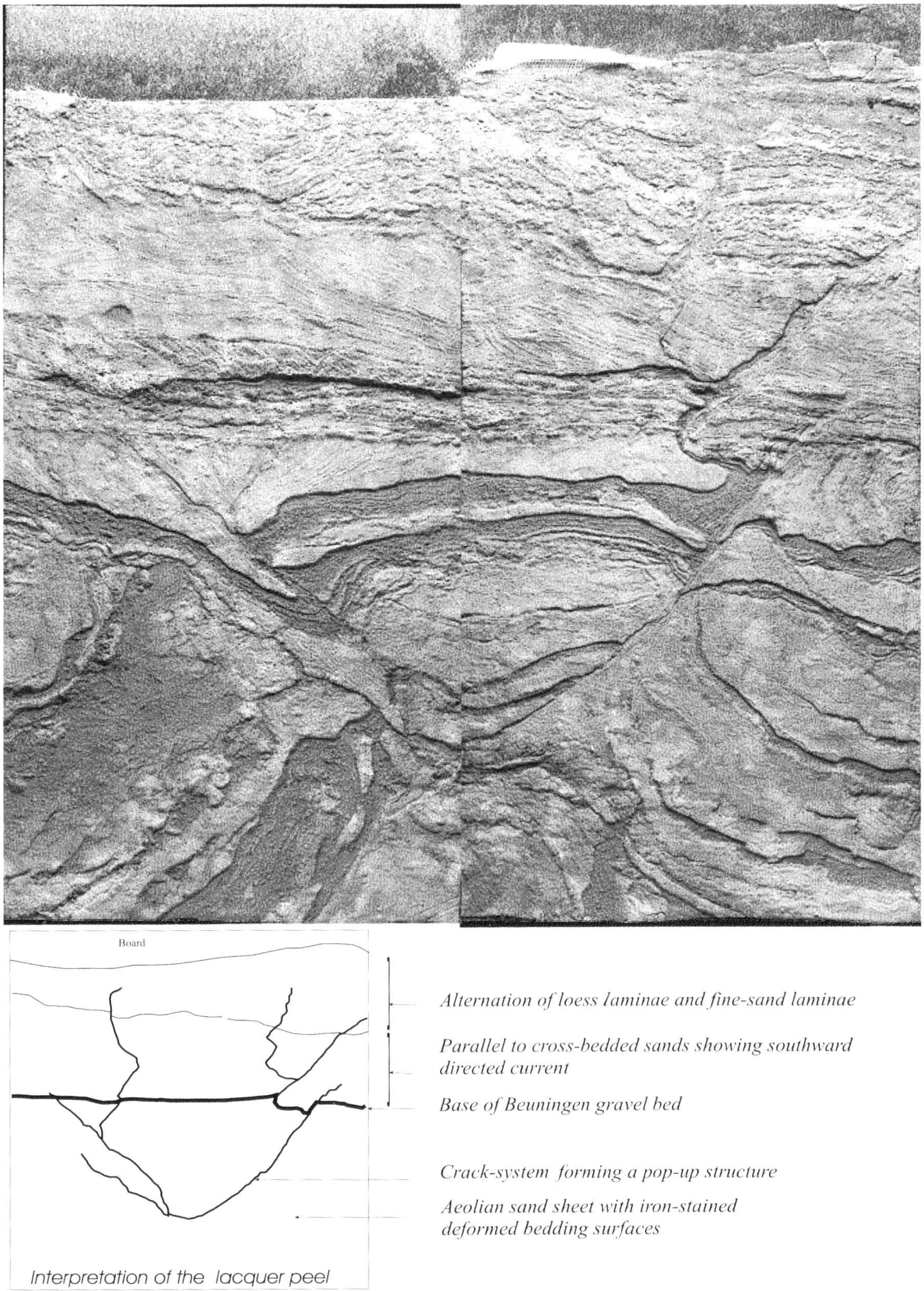

Fig. 6. Fluvio-aeolian sands on a lacquer peel displaying the two sedimentary sets directly overlying the Beuningen gravel bed. A system of cracks, crosscutting the Beuningen bed, dies out in the set with alternating fine sand and loess laminae. This system is interpreted as a pop-up structure related to the F-1 faulting event and marks the lower age boundary of this event. Pannel scale: appr. 1 sqare m. Position of this figure is marked in Fig. 4.

Three aspects are important for Bt formation: pH of the soil, soil-wetness and temperature, their development through time is important to get insight into the chronostratigraphic significance of the soil. This

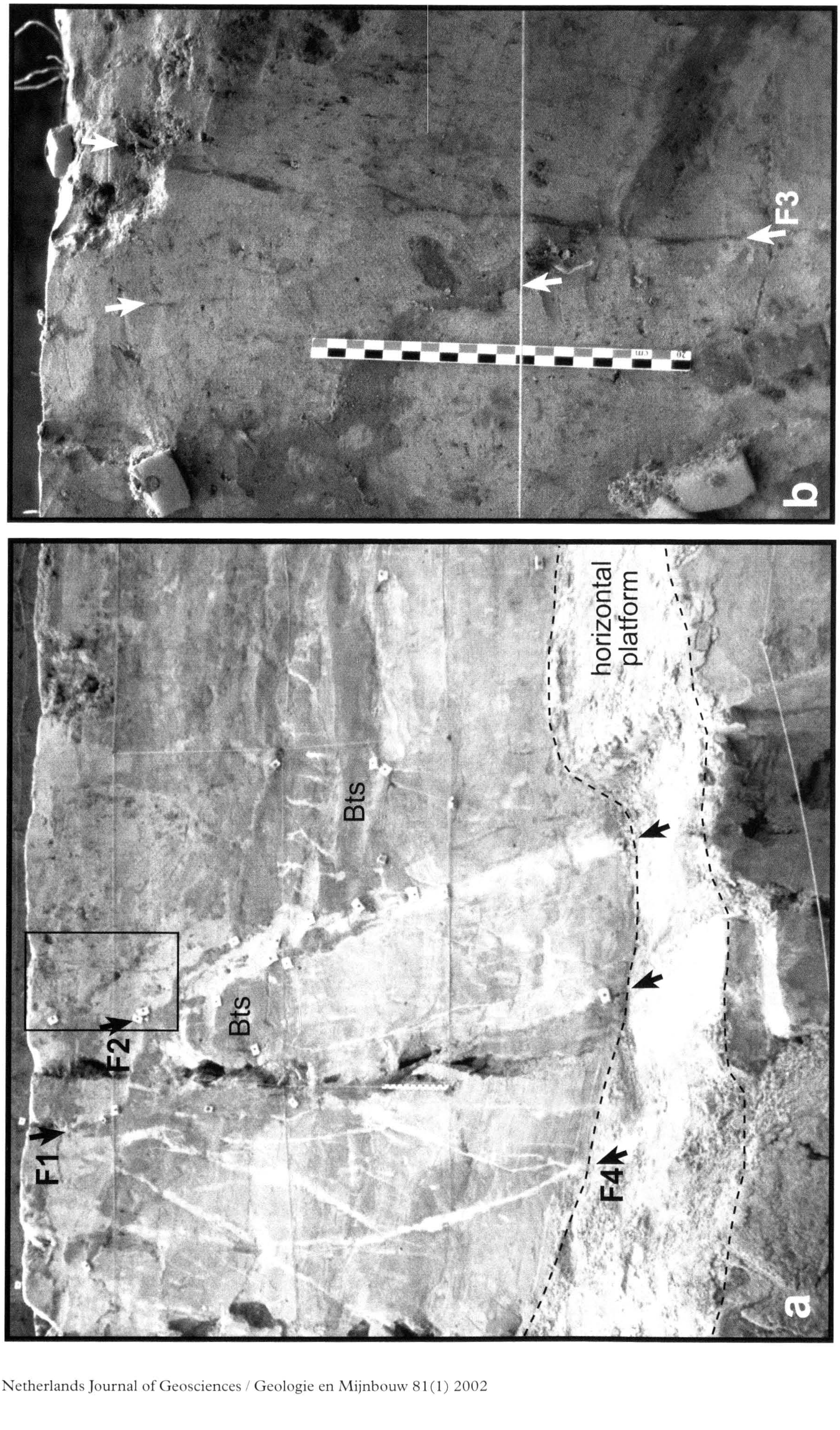

Fig. 7. Detail of the faultzone, for line-drawing see Fig. 9., for the position see Fig. 4.

will be discussed below and is summarised in figure 8.

- The pH of the soil should be slightly acid to allow for clay migration. As long as fresh calcareous coversand is deposited the pH of the site will not fall below 7; as soon as this deposition stops, rapid decalcification sets in; it will take only a few hundred years to lower the pH sufficiently. Coversand depsition lasted during till the Early Bølling and resumed after the Late Glacial interstadial conditions during the second half of the Younger Dryas (Kasse, 1999).
- The soil-wetness is affected by the permeability and the rainfall excess minus the evapotranspiration. As long as permafrost conditions ruled the soil permeability, downward migration of soil water was prohibited; such conditions lasted till the Beuningen period. After this, only deep seasonal frost affected the soil until the Bølling climatic optimum. Similar conditions lasted during the Younger Dryas.
- Temperature peaks during the Bølling and shows a general decline afterwards to the Younger Dryas. The pollen records show for the Bølling/ Allerød sequence a reverse development in the canopy cover (Hoek, 1997). During the Bølling and Early Dryas pollen diagrams reflect less than 50% in tree pollen, suggesting a savannah-type landscape. During the Allerød and Holocene the forest vegetation was probably already much denser. A savannah-type habitat favours population growth of large herbivores on which dung scarab beetles thrive (Brussaard and Runnia, 1984). We suggest an optimum in their burrowing activity during these intervals. Trace fossils of these beetles have been recognised in the trench, overprinting the horizon with Bt-features.

The combination of all these factors may have provided optimal conditions for Bt-formation during the Late Bølling and the following Older Dryas as well as during the transition from the Younger Dryas to the Holocene. The observation that scarab beetle burrowing disturbs the Bt-band confirms an early timing of the main phase of Bt-formation within the Late Glacial. A tentative approximation of soil forming intensity and dung beetle activity is also presented in Fig. 8.

The B-horizon shows not only characteristics of illuviation of clay (Bt) (observed in thin sections) but also of sesqui-oxides (Bs), proven by chemical extraction (pers comm. J.van Mourik, 2000). These observations point to two phases in soil development: first a phase of clay transport followed by a phase of pod-

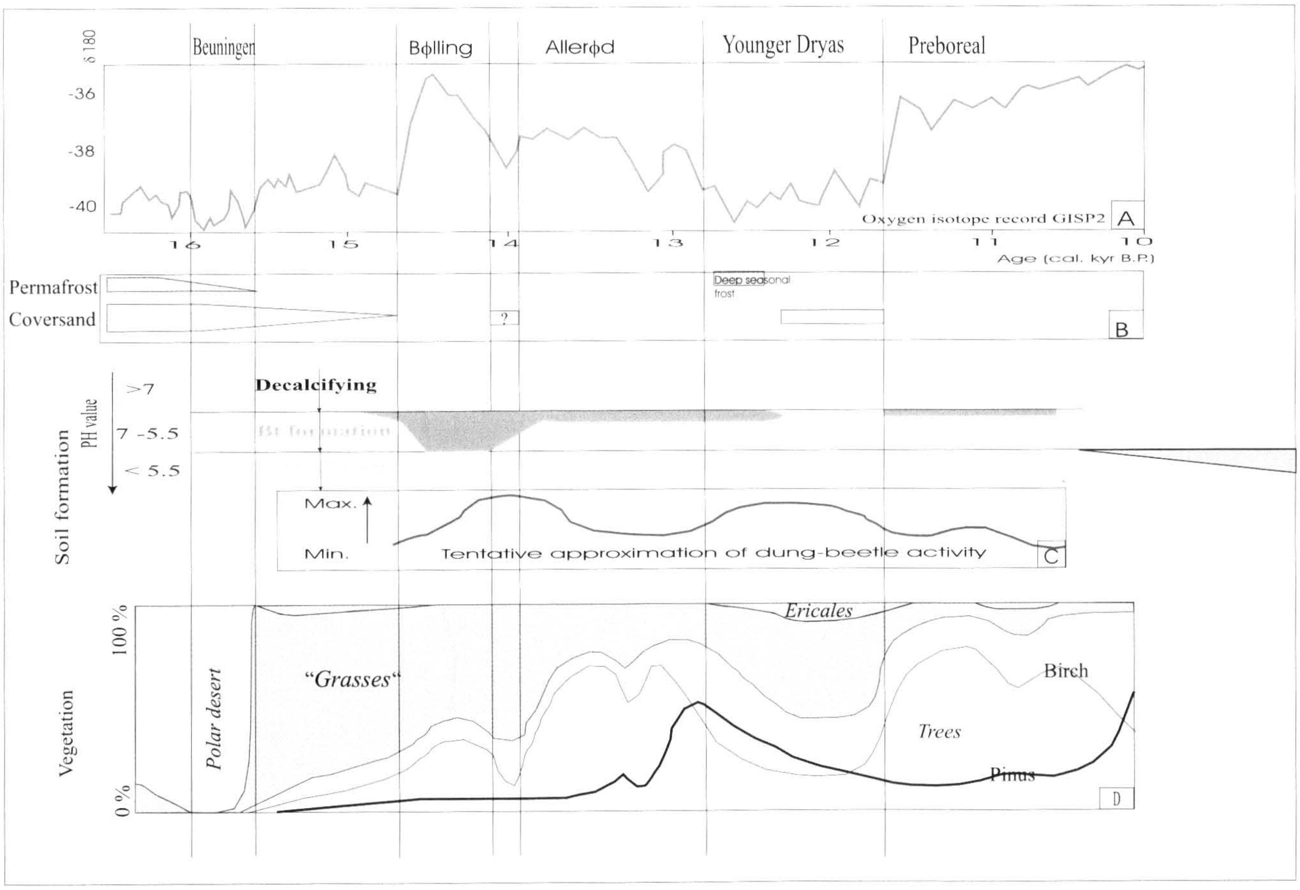

Fig. 8. Panels showing: (A) the climatic evolution (Stuiver et al., 1995), (B) periods with coversand and permafrost conditions (data according to Hoek, 1997 and Kasse, 1997), (C) main soil forming processes (modified after Miedema, 1983) and (D) ecosystem evolution.

zolisation. The main soil forming process for the formation of the Bs is cheluviation. This is the transport of soluble humus-metal complexes (chelates) out of the surface layer (A and E horizons) into the B horizon. This process can proceed only after the phase of clay illuviation was finished, providing the pH drops below 5.5. 14 C dating of these chelates suggests a mid-Holocene age (Van Mourik, pers comm 2000) for this process.

Faulting geometry and history

Fig. 9 shows a detailed map of the zone of faulting on the NW-facing wall in the Neer. The relatively horizontally stratified fluvio-eolian sands are displaced by a set of normal faults concentrated in a 2-m-narrow fault zone which is situated more or less at the foot of the (artificially reduced in height) scarplet. The main faults, F1 and F2, are dipping 75° SW and are oriented N 120° E and N 140° E, respectively. F3 is a more or less vertical fault strand branching upward from F2, and F4 is a minor synthetic fault which seems to become more important laterally. The summed fault displacement of the main stratigraphic marker, the Beuningen gravel bed, is about 90 cm (F1: 55 cm, F2 + F3: 30 cm, F4: 5-10 cm). Considering that the Beuningen gravel bed originated as an essentially flat surface in this area, an additional amount of flexure must be taken into account, however, so that the total offset across the fault zone is in the order of 150-175 cm. Another distinct horizon, which can be assumed to deposited horizontally was the 'bleached wave-ripple horizon' (marker 1 in Fig.5), here about 1.5 m below the Beuningen gravel bed. This marker shows the same total displacement, implying that no fault motion occurred during the time interval between deposition of both horizons. We roughly estimate the length of this timeslice at about 15 kyr inferred from OSL-datings. All deformation observed in the trench thus post-dates the Beuningen gravel bed (i.c. 15800 cal yrs BP).

Detailed analysis of the faulting history in the uppermost part of the section is hampered by strong human activity (ploughing), reaching locally down to 1 m below the ground surface (e.g. 11-12 m and 16-17 m in Fig.9), and by soil development processes obscuring the upper part of the stratigraphic section. Certain aspects of the pedology can help to constrain the timing of slip on the different fault branches, however. The most prominent soil features are Bt-bands, which have been described above. The base of these Bt-bands is usually associated with certain bedding planes. The main phase of Bt-development in the sands bordering the Maas valley in the RVG has been discussed above, (Fig. 8). Fault strands F1 and F2 show a different relationship with the Bt-bands. While the main Bt-band is sharply displaced by the same amount as the Beuningen gravel bed on F2, there does not appear to be a true offset (discontinuity) on F1; base and top of the main Bt-band are more or less continuous from one side of the fault to the other, except that the base of the Bt-band changes from a more or less horizontal attitude in the hanging wall to a clearly inclined attitude in the footwall, cross-cutting the original stratification. We interpret this to indicate that development of this Bt-band occurred after faulting on F1, but was mostly completed before movement of F2. The inclination of the base of the Bt-band in the footwall adjacent to F1 would be caused by interruption of the horizon normally associated with the base of the Bt-band and by illuviation along the fault plane itself. The latter is evidenced by macroscopic rubifaction in and parallel to the F1 fault plane and was confirmed by thin sections showing in-situ clay coatings on sediment particles oriented parallel to F1, which thus post-date F1 activity (Miedema and Jongmans, this volume). In contrast, the zone adjacent to F2 has a more whitish appearance, which is interpreted, (on the bases of sedimentary structures evident from the laqueer peels: a clear flow fabric and the presence of fragments of host sediment embedded within the white sand.) as a clastic dyke, formed by upward injection of liquefied sand. This is suggested by Several syn- and antithetic fissures and faults with small offset (including F4) in the footwall are associated with white-coloured clastic dykes as well, and some are cross-cutting the inclined Bt-band adjacent to F1 (Fig.7 12-13 m). These relations are consistent with a creation of these faults and fissures contemporaneous with F2, after development of the Bt-band.

In the hanging wall, the yellow fluvio-aeolian sands are overlain by a unit of pale, silty fine sands (E horizon of the Holocene soil). The unit is completely structureless, and obscured by a more than 50 cm thick B-horizon. Judging from the lighter colour of this horizon it is a different type of B-formation, less intense (maybe a younger Bt) joining the Bt-bands in the unit below. Analysis of thin sections from the contact between stratified fluvio-aeolian sands and overlying, structureless, silty sand (Miedema and Jongmans, this volume), shows that the latter unit contains fragments of earlier formed and later broken banded Bt-horizons, thus demonstrating that it is a colluvium. In addition, small fissures cutting through the Bt-bands in the unit below, do not extend into the colluvium, but on the contrary are infilled by colluvial material. These observations suggest that the base of the

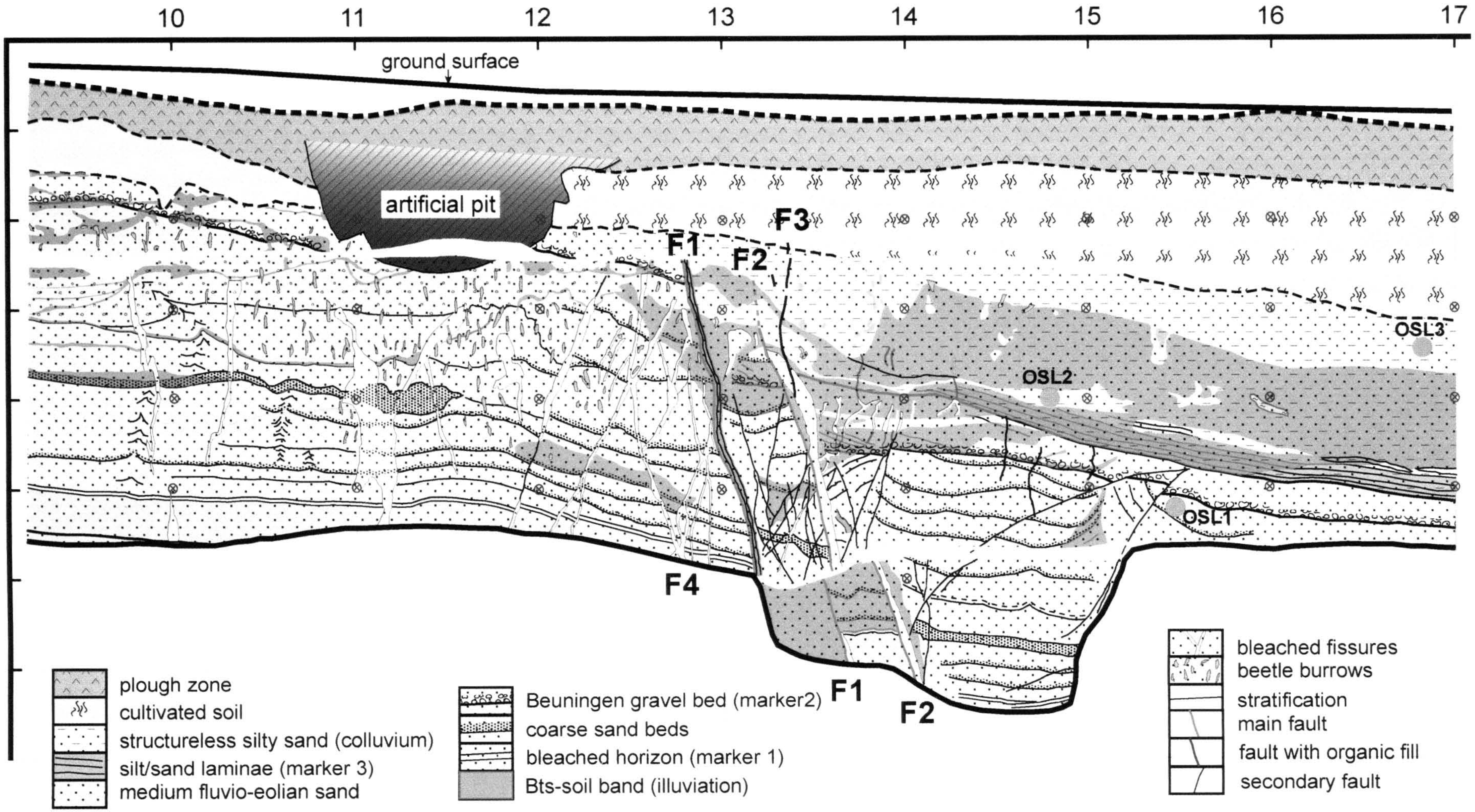

Fig. 9. 4 m x8 m section of the NE facing trench wall showing that fault planes are concentrated within a 1m-wide zone. Colours: yellow= sand; brown = clay and iron infiltration (Bts soil horizon); white bands = bleached fissures; grey-tones = antropogenic plough horizon; sub-horizontal black lines = bedding planes; oblique lines = fault planes; red dots indicate sample sites for O.S.L. dating.

silty sand unit may represent an event horizon, related to faulting on F2. The silty sand unit itself could thus have been deposited as resulting from degradation of a fresh fault scarp. Extensive farming activities have obliterated the top of the silty sand unit in the hanging wall, and do not allow to determine if this unit is present in the footwall or not, thus hampering further analysis. The colluvial silty sand unit is transected by another small, near-vertical fault, F3, branching upward from F2, and extending to the base of the cultivated soil. This fault is filled with dark grey organic matter (moder humus formed by decay of Holocene roots, (Miedema and Jongmans, this volume) and also sharply displaces the B-band that developed within this colluvium by 5-10 cm (Fig. 7). Unfortunately, the fissure has not been sampled for radiocarbon dating, leaving the minimum age of this fault motion undetermined.

In summary, stratigraphic and pedologic relations provide evidence for three separate faulting episodes (or events) on the Peel Boundary Fault at Neer since the Last Glacial Maximum. The most plausible reconstruction of faulting history at the trench site is presented as a cartoon in Fig. 10. Fig. 11 shows the integration of the various stratigraphic tools into a timeframe.

No faulting occurred between the start of the trench record (± 25 kyr BP, i.e. OSL-age of bleached wave-ripple horizon) and the Last Glacial Maximum (represented by the Beuningen gravel bed). The largest amount of displacement occurred on F1, shortly after deposition of the Beuningen gravel bed (± 15.8 kyr BP), but before the main development of the Bt-banded soil, which started around 14.5 kyr BP, a fairly narrow age bracket. The fluvio-aeolian sequence was thrown down at least 55 cm on F1, probably accompanied by a comparable amount of flexure, though the exact share of the total amount of > 60 cm of observed flexure cannot be resolved for each of the faulting episodes. We have not recognised a colluvial wedge related to the F1 event. During the Bølling Interstadial, the main Bt-bands started forming within the fluvio-eolian sandsheet. When development of the Bt-bands was mostly completed, the sandsheet, and the Bt-bands contained within it, were displaced by a new set of faults (F2 and associated syn- and antithetic faults). The total amount of vertical offset was about half as large as for the previous faulting episode, and was at least 30 cm (25 cm on F2 + 5 cm on F4), possibly twice that value when taking flexure into account. Following this faulting episode, a thick unit of structureless silty sand was deposited in the hanging wall, as a colluvial wedge derived from degradation of the surface scarp, possibly in combination with continued eolian input (we don't know if the silty sand is present in the footwall or not). The base of this colluvium truncates some minor fissures that are probably related to faulting on F2, and thus resembles a so-called event horizon or earthquake unconformity commonly described in literature as the result of coseismic surface rupturing (McCalpin, 1996, and references therein). The lower age boundary for this second faulting episode is situated after the main Bt-development took place, thus somewhere within the Bølling/Allerød Interstadial. However, the terrace morphology suggests that the Peel fault scarp looses its topographic expression in the terraces younger than terrace T-2. (described above, Fig.3) This suggests that the major fault activity occured before terrace T-3 was abandoned by the river Maas (before 13 Kyrs calibrated age). In agreement with the OSL-dating of the base of the associated colluvial wedge which yielded an upper age boundary of 12.2 ± 1.4 kyr BP. A timing which is confirmed by independent trench evidence that most of the faulting should pre-date beetle burrowing activity, which is supposed to both have peaked during the late Bølling and the Younger Dryas. The time-window for the second fault activity is thus narrowed down to a period of about one thousand years.

Finally, the peculiar fracture F3, branching upward from F2, shows evidence of a smaller movement (only 5-10 cm) in more recent times. Its fault plane contains dark grey humic matter that results from the decay of a plant root (Miedema et al., this volume). This organic material is most likely late Holocene in age, possibly even quite recent.However, the fissure cannot possibly be attributed to the M_W=5.42 Roermond earthquake in 1992, which originated at a depth of ± 17 km (Camelbeeck et al., 1994), and is therefore not expected to have caused visible surface rupture (Camelbeeck, pers. comm.).

We thus conclude that two relatively large faulting events occurred in the relatively short (around less than 3 kyrs) span of time between the Last Glacial Maximum (± 15.8 ka. BP) and the onset of the YoungerDryas (± 13 ka. BP), whereas only one very small displacement occurred throughout the Younger Dryas and Holocene.

The trench record also contains some indications that support a coseismic nature of the identified fault displacements. First, the combination of absolute (OSL) and relative (stratigraphic, pedologic and geomorphologic) dating techniques show that the two largest fault movements each occurred in a fairly short span of time (about 1300 years for the oldest 'event', and even less for the penultimate 'event'). In contrast, virtually no fault displacement occurred

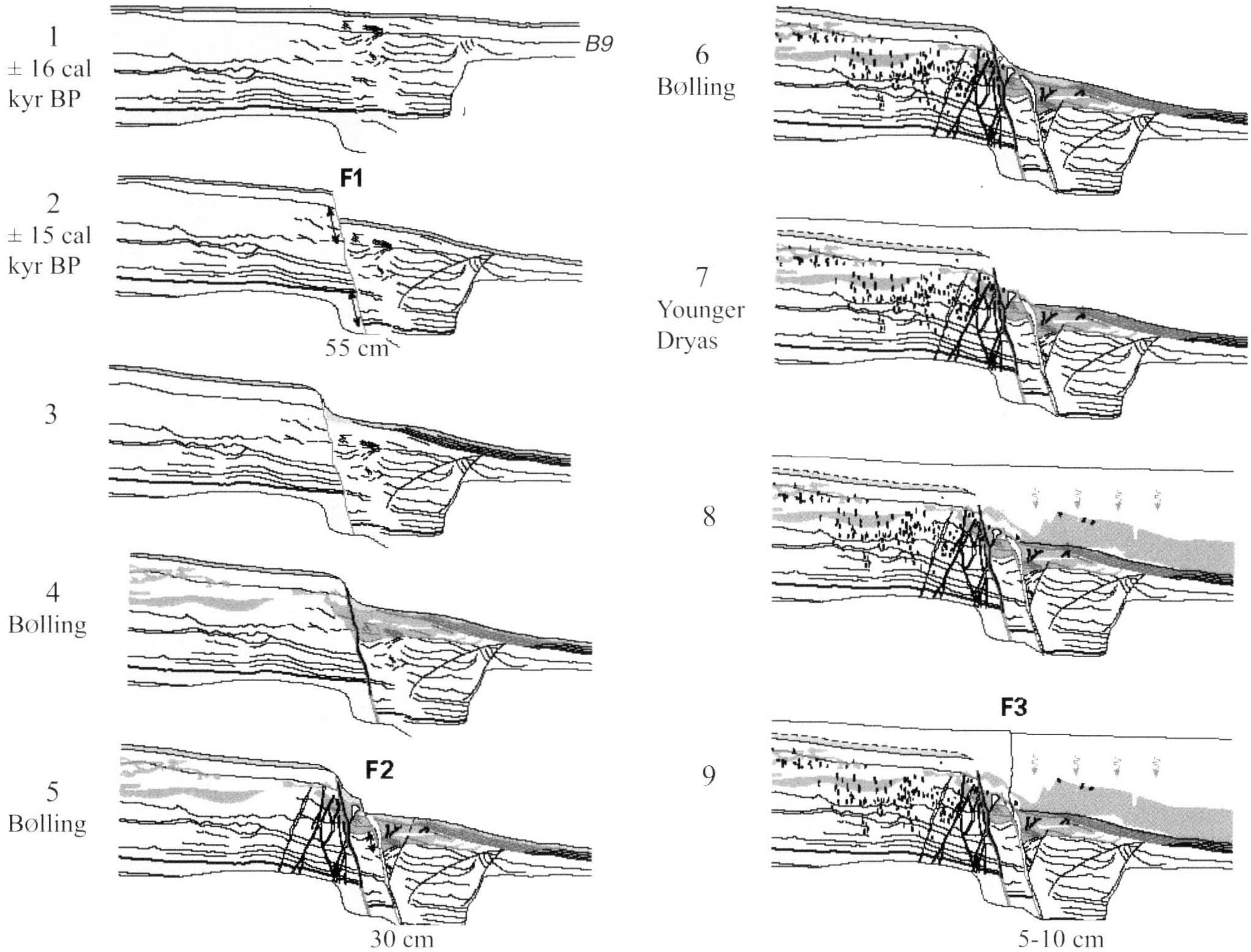

Fig. 10. Tentative reconstruction of faulting history of the Peel Boundary Fault at Neer: 1) situation just after deposition of the silt/sand alternation (Bg = Beuningen gravel bed, bh = bleached horizon = reference level 1); 2) displacement on F1; 3) continued deposition of alternating sand/silt laminae; 4) development of the main Bt horizon during the Bølling; 5) displacement on F2; 6) main activity of beetle burrowing; 7) deposition of aeolian sands, burying of the faultscarp during the Younger Dryas; 8) development of the Bs, cheluviation in the aeolian sand; 9) small displacement on F3; modification of the profile by human activity resulted in the present-day situation presented in Fig. 9.

during the remainder of the trench record. This can be proven for the interval between the deposition of the 'bleached horizon' and the Beuningen gravel bed, and, if our dating of the penultimate event is correct, is also inferred for the time interval end of the Allerød till the present, with the exception of the faulting event that produced F3. It should be noted, however, that for the latter period aseismic flexuring cannot be excluded, as it is impossible to unambiguously attribute any observed amount of flexure to individual faulting events.

In addition, we have shown that a unit of structureless silty sand in the hanging wall may be interpreted as a colluvial wedge, and its base, which truncates several small fissures, as an earthquake unconformity or event horizon. These stratigraphic relations are often observed in exposures of normal fault zones known to produce surface rupture (McCalpin, 1996). A final argument in favour of coseismic faulting concerns the dykes of white sand that appear to be associated with F2 and a set of syn- and antithetic faults. Detailed inspection of these dykes show that they contain liquefied sand (see above). The most likely cause of liquefaction in this environment is seismic shaking.

Finally, it is important to note that the faults exposed in the trench at Neer show no evidence of fault slip related to the rapid relative block movements, in the order of ± 0.5 mm/yr at the trench site, that are inferred from repeated levelling (Van den Berg et al., 1994). As shown above, fault F1 is sealed by Bt-development, while F2 becomes less distinct macroscopically towards the top of the trench wall due to soil processes. F3 is the only fault visibly reaching up to the base of the cultivated soil, but it is filled with organic matter that is probably very young. The organic material shows no shear deformation, indicating that no fault creep has occurred on this fault dur-

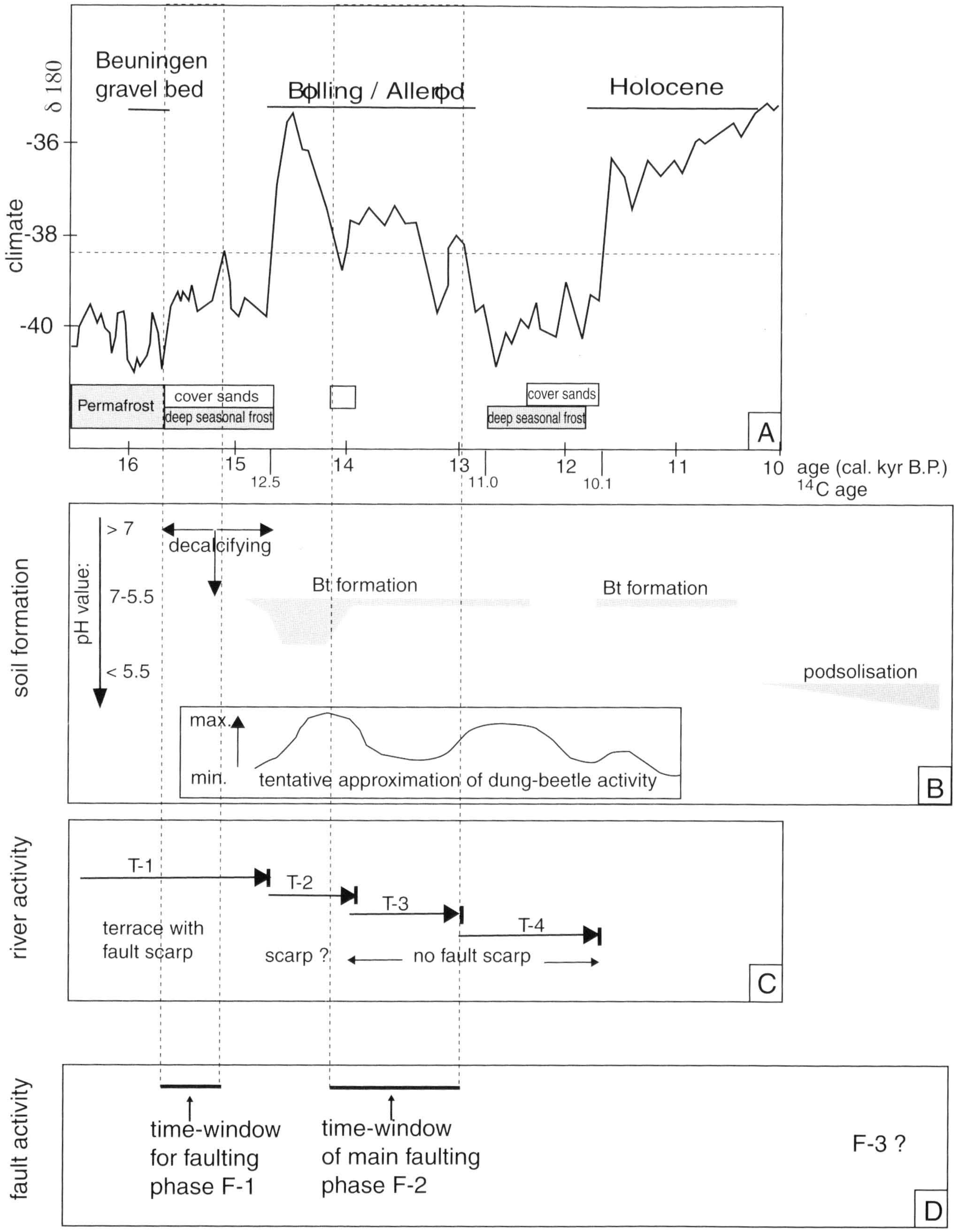

Fig. 11. Three phases in fault activity along the Peel Boundary Fault near the village of Neer (D). Their position is related to the linear-time-window in (A) according to their crosscutting relationship with soil forming processes (B) and river activity (C). B= modified after: Miedema, 1983, and C= after Van den Berg, 1996.

ing the period of levelling observation. It can not be excluded, however, that these modern block movements are responsible for part of the flexuring observed in the trench.

Regional significance

We have established a relatively high-resolution record of significant, high magnitude (see below), crustal activity along the PBF during the close of the last glacial period. The nearby Eifel-Rhenish Shield Palaeozoic massif (which is seated above a mantle plume) is well known for its volcanic activity during the same period and we want to explore the likelyness of a structural relationship between the two areas.

The Palaeozoic massif is dissected by a SSW-NNE striking zone with various discrete eruption cores of different ages. Two of them are located in the Eifel; respectively the West Eifel and the East Eifel, they are about 40 km apart and positioned in the SE prolongation of PDZ's of the Roer Valley graben (Fig. 12). This suggests that these fractures continue into the massif but without a stepped surface expression. The volcanic activity shows that crustal-scale fracturing is involved in the massif, a characteristic which was shown by deep seismic images for other sections of the PDZ's as well.

A closer look at the cluster of the various eruption centres from both the West- and East Eifel volcanoes shows prefered orientations in their distribution. Meyer and Stets (1979) have shown that several of these eruption centra are linked up by structural elements. Their observation suggests that the above mentioned preferred orientation may be also related to lineaments. A NW-SE strike of the cluster is most prominent both for the West Eifel as well for the East Eifel (not shown here). Less prominent are SW-NE to W-E orientated patterns.

This angular relationship resembles an arrangement according to a conjugate set of structures (Fig. 12). We interpret these lineaments to be related to a fracture set related to a, deep-seated, basement shear zone, whose postulated position and orientation perfectly lines up with the overall strike of the southern PDZ of the Roer Valley Graben (the Feldbiss Fault zone) (Fig 12). It is remarkable that the same orientation can be recognised in a lineament formed by several volcanoes including the Laacher See, 40 km to the east (the same distance as the distance between the southern – and the northern PDZ of the Roer Valley Graben). Even the position of this Laacher See lineament is in perfect alignment with the PBF. All this suggests a spatial relationship between the PDZ's and the volcano fields.

A recent study of Maar lake infill revealed that at least two eruptions took place here between 13 Kyrs calibrated age and 11 Kyrs calibrated age (Zolitschka, 1998), respectively the Laacher See and the Ulmener Maar explosions. Their ashes have been identified in the varve record of the lake infill. We have combined these observations into one graph with our results (Fig. 13), this reveals a remarkable similar event-recurrence time in the order of 1700 years for both types of crustal activity. This may hint at a temporal relationship in addition to the above concluded spatial relationship.

Maybe both the Neer record and the Maar record, although different in expression, reveal dynamics of the same crustal fractures. This points to a hypothesis about the existence of a crustal dynamical rhythm leading to high magnitude events along the fracture zone as a whole. But both records also show that we cannot expect to find the complete evidence of this high-magnitude stress release at one single site because such events apparently wander along the whole length of the crustal fracture. We need more trenches to testify this as well as to establish the length of the fracture-sections involved.

Evaluation of seismic hazard

The main results from trenching are the identification of three stages of events with a displacement of 10, 30 and 55 cm and an additional flexure of > 60 cm. In this section we evaluate the interpretation of the trench data in terms of seismic hazard.

Magnitude

Although empirical relations exist between surface displacement, rupture length, rupture area and magnitude (Wells and Coppersmith, 1994), a careful evaluation of assumptions and uncertainties is required. First of all a variety of magnitudes is used and the conversion between them is not always clear. There is a general acceptance that moment magnitude (M_W) is the reference magnitude, since this is based on measurements of the seismic moment M_o which is a direct measure of the energy at the source. Other definitions as local magnitude, M_L and surface wave magnitude, M_S, are good approximations in a limited magnitude range. In general M_S underestimates M_W at small magnitudes (<6), while M_L underestimates M_W at large magnitudes (>6, saturation, see McCalpin, 1996).

A test of the relations between magnitudes for the 1992 Roermond event, using empirical relations between M_o and M_L (Ahorner, 1983, 1994), shows a

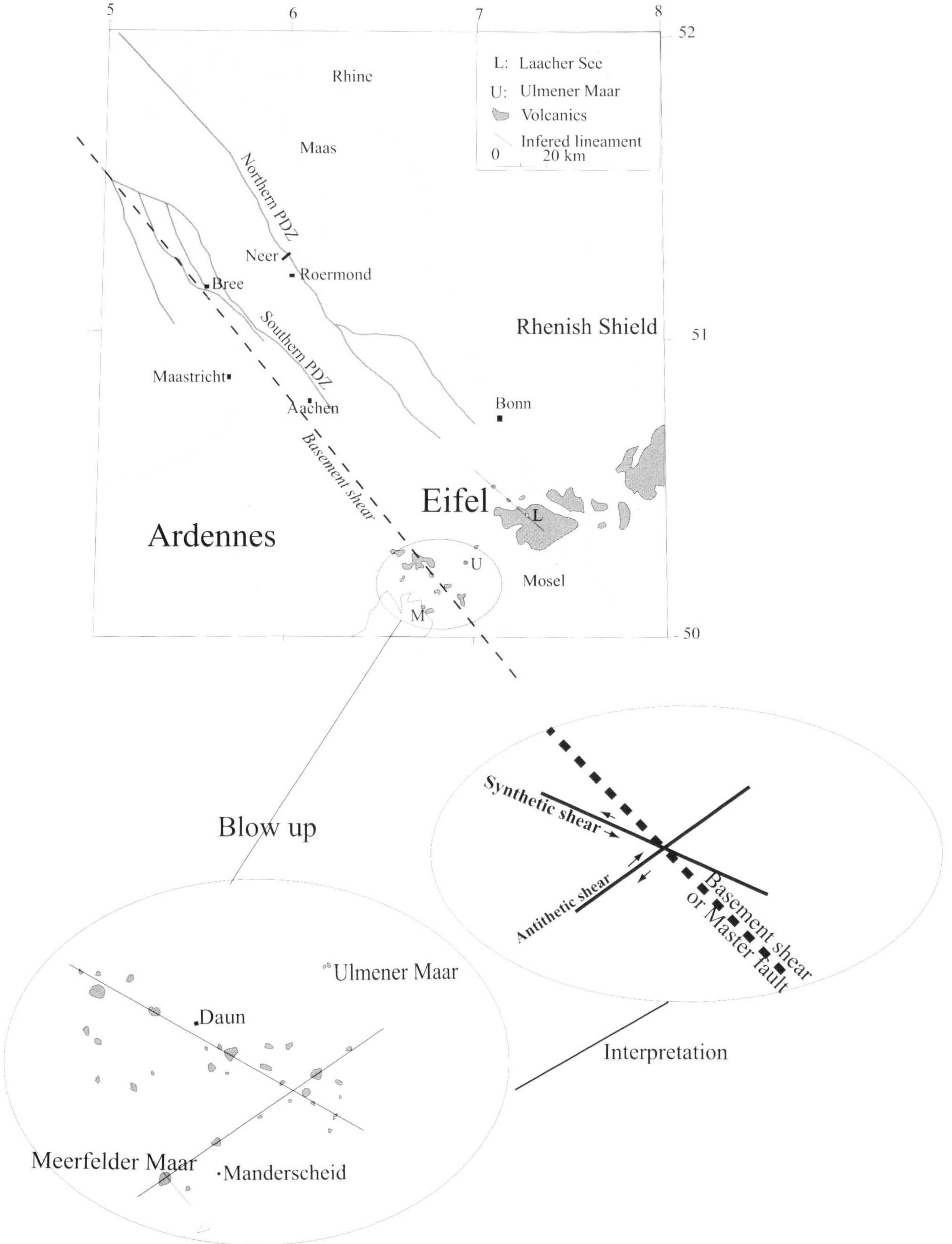

Fig. 12. The position of the Eifel volcanics with respect to the studied fault system together with the fractures inferred from lineaments and interpreted first and second order stress pattern. (PDZ: principal displacement zone).

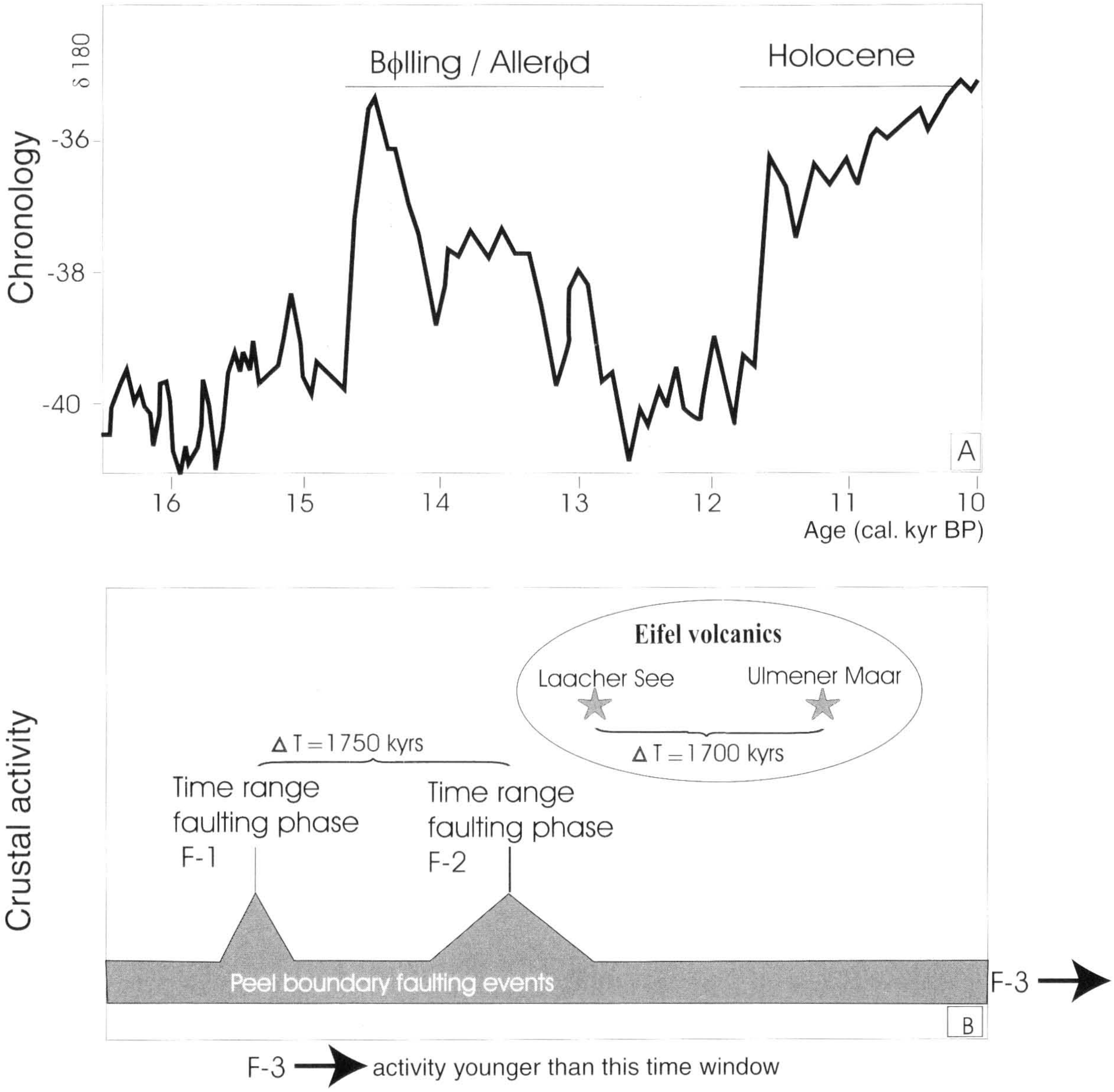

Fig. 13. A composite of panels showing the chronology of the field observations and derived appr.1700 yrs periodicity of crustal activity.

different relation. A measured M_L=5.8 corresponds to a lower value of M_W= 5.2, close to the measured value for M_S. This inconsistency makes some authors (e.g. Camelbeeck et al., 2000) use M_S as approximation for M_W and may result in a underestimate of low magnitudes. In turn this will lead to an overestimate of b-values in a cumulative frequency versus magnitude plot and results in an overestimate of recurrence times. A detailed study of this magnitude inconsistency is therefore a requirement for the interpretation of paleoseismic results in the region in terms of seismic hazard.

An estimate of the error in the magnitude determination is given by comparison of the measured values for M_o , which lies between 5.4E+16 Nm and 1.5E+17 Nm (Camelbeeck et al., 1994). In moment magnitude this means between M_W= 5.1 and 5.4, while M_L varies from 5.8 to 6.2. Even for a well-recorded modern event a considerable difference in magnitude is encountered.

Bearing in mind the afore-mentioned uncertainties and taking into account that only one trench was realised, the magnitudes of the paleo-events can be estimated using two methods (Wells and Coppersmith, 1994): length of seismic rupture and maximum displacement.

In this study, a preliminary estimate of the seismic rupture length is made on the basis of geomorphology, regional geology- the lateral extent of the single faultplane as can be followed in seismic reflection lines- and on the basis of regional subsidence data. The estimated rupture length at Neer is 10-20 km and implies a magnitude of M_W=6.2 (10km) to 6.6 (20km).

The maximum displacement measured for a single event in the trench is 10-50cm (or even more if one considers flexuring as well). These values relate to a magnitude of M_W= 6.0 (10cm) and 6.5 (50cm) assuming that they represent the average displacement along the total rupture length.

Both methods give comparable results. However, accuracy is limited and can only be improved when more trenches are opened along the fault zone. Then average displacement can be used and the continuation of the events can be traced.

Recurrence times

The present status of dating is the indication of two to three surface-rupturing earthquakes

Occurring within the last 16000 years. Possibly, the first two events did occur in a time span of less than 2000 years.

It is tempting to extrapolate cumulative frequency-magnitude curves, based on recent activity, to a distant past (e.g. Camelbeeck et al., 2000). The main problem is that these curves are expected to behave linearly in only a limited magnitude range and show an exponential decrease at large magnitudes. Although the present curves show a linear behaviour until M_L= 5.8, it is unknown where the non-linear behaviour starts. If this is combined with a relation where the magnitude is M_S instead of M_W, extrapolated recurrence times have no real meaning.

Discussion and conclusion

Palaeoseismological investigations strongly benefit from a multi-disciplinary approach both with respect to research fields (knowledge of soils, stratigraphy, geophysics etc), as well as to research methods (for example: absolute- vs. relative dating techniques; thin-section analyses vs. field observations; geophysical tools in combination with more conventional tools). It appeared that important information not only came from the investigations within the trench itself but also from the regional context. A remarkable result is the observation that magnitudes over M_W= 6.0 did occur in the region.

The accuracy of the trench record is very course compared to the registration of the volcanic events. On the time axis the two Allerød events seem to have been very close. This raises the question whether the Peel boundary F-2 faulting event and the Laacher See volcanic event were in reality more closely linked in time and indeed different expressions of the same crustal stress release or that different mechanisms were involved and that their close occurrence is pure coincidence. It is obvious that we need more recordings to solve this temporal problem.

The volcanic activity stopped with the Ulmener Maar explosion, but the faulting went on although the recorded magnitude so far, has been much lower. The reason for this is still unknown. One possibility, suggested by the episode during which the events took place, is that the driving force was a-tectonic in nature. It may have been generated by crustal unloading due to melting of the large continental ice sheets during the last Termination. If this were to be the case, such high magnitudes would indeed be restricted to the remote past only. Volcanic activity however is known to have occurred both during glacial periods as well as during the interglacials suggesting tectonics were the driving force. In that case we only can conclude if we know the representativity of the trench results for the whole fault zone. To this more trenches have to be opened along the bounding faults of the Roer Valley Graben. This stresses our general conclusion that palaeoseismic research has a large potential, but should be carried out on a larger scale to possibly detect younger, but similar, events along other sections of the same structural element. A condition sine qua non, our results can be used in seismic hazard analysis.

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