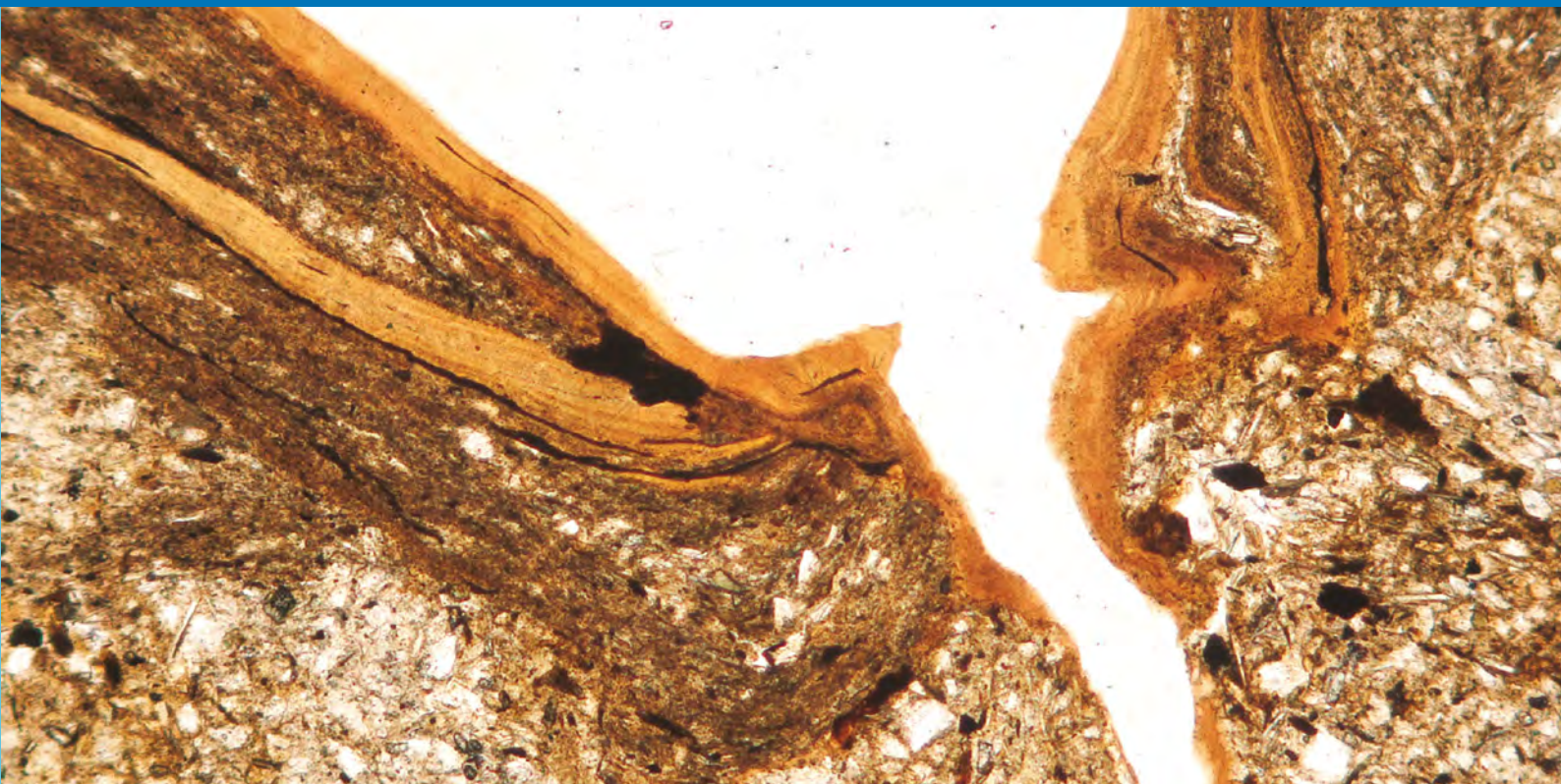


# E&G

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Quaternary Science Journal



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## MIDDLE TO UPPER PLEISTOCENE PALEOSOLS IN AUSTRIA

**GUEST EDITOR** Birgit Terhorst

# E&G

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## Middle to Upper Pleistocene paleosols in Austria

*Preface to the special issue*

Birgit Terhorst

Since the year 2000 we carry out research projects in the area of the Pleistocene Salzach glacier, the Inn terraces of the 'Innviertel', as well as in the area of the Traun-Enns Plate (Traun-Enns-Platte) in Upper Austria. Five papers of the special issue are concerned with Quaternary sediments, paleosols and recent soils in Upper Austria. The focus is placed on the Oberlaab Loess-paleosols sequence, which was studied in great detail. National as well as international working groups were successively involved and there are numerous studies, which comprise classical field and laboratory methods, microscopic, magnetic, and paleobotanic analyses. The research works started on the base of the project 'Geomorphological and Pedological Investigations in the Former Foreland Glacier of the River Salzach/Austria', which was funded by the German Research Foundation (DFG) under the guidance of Prof. Dr. Erhard Bibus, Tübingen. Moreover, subsequent studies were supported by the University of Tübingen and an international project, which was part of the ICSU Grant Program under the title 'Polygenetic Models

for the Pleistocene Paleosols: A new Approach to decoding the Paleosol-Sedimentary Records'. The research team was formed by a German-Austrian-Mexican-Russian group, of which the major part participated in the present special issue (OTTNER & SEDOV, SEDOV et al., SOLLEIRO-REBOLLEDO et al., TERHORST; this volume). A further project in the area was organised by Prof. Dr. van Husen (ÖAW; Austrian Academy of Science). In the frame of this project concerned with the stratigraphy of the Middle Pleistocene (Untersuchungen zur zeitlichen Gliederung des mittleren Abschnitts des Quartärs), detailed studies were carried out in the former brickyard of Wels-Aschet (see SCHOLGER & TERHORST, this volume).

Since 2007 the research area was extended to the east by the loess region of Lower Austria. The research team experienced an enlargement by the cooperation with Austrian archeologists in the Krems region. The extension of the research area is represented by the paper of Sprafke et al. (this volume) and deals with the well-known outcrop of Paudorf.

# A stratigraphic concept for Middle Pleistocene Quaternary sequences in Upper Austria

Birgit Terhorst

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**Abstract:** Three profiles of loess-palaeosol sequences on top of Middle Pleistocene fluvio-glacial terraces of the Traun-Enns-Plate are investigated in the region of Wels (Upper Austria), each of them representing characteristic Middle Pleistocene sequences for the northeastern Alpine Foreland. The sequences comprise thick pedocomplexes, providing the opportunity to distinguish and to classify specific interglacial palaeosols. The loess-palaeosol sequence of Oberlaab developed on top of the fluvio-glacial terrace of the classical Mindel (Younger Deckenschotter) shows four interglacial palaeosols. This fact suggests that the age of the terrace is at least the fifth to last glacial period, correlative to MIS 12. The cover layers on top of classical Günz terrace (Older Deckenschotter) in Neuhofer and Wels-Aschet include five palaeosols. Both sites are characterised by intense pedogenesis in the basal pedocomplex, which is considerably more pronounced than in the overlying palaeosols. Pedostratigraphic results point out that the genesis of the studied Günz Deckenschotter can be correlated to MIS 16 (minimum age).

## Ein stratigraphisches Konzept für mittelpleistozäne Quartärabfolgen in Oberösterreich

**Kurzfassung:** Auf den mittelpleistozänen fluvio-glazialen Terrassen der Traun-Enns-Platte in der Region um Wels (Oberösterreich) wurden drei Löss-/Paläobodensequenzen untersucht. Jedes dieser Profile ist für mittelpleistozäne Abfolgen im nordöstlichen Alpenvorland charakteristisch. Die Profile umfassen mächtige Pedokomplexe, welche eine Differenzierung und Einstufung von interglazialen Paläoböden erlauben. Die Löss-/Paläobodensequenz von Oberlaab ist auf der fluvio-glazialen Terrasse des Mindel-Glazials im klassischen Sinne entwickelt (Jüngere Deckenschotter) und weist vier interglaziale Paläoböden auf. Diese Tatsache macht eine Einstufung der Jüngeren Deckenschotter mindestens in die fünftletzte Kaltzeit wahrscheinlich (MIS 12). Die Deckschichten auf den Günz-Deckenschottern im klassischen Sinn (Ältere Deckenschotter) beinhalten fünf Paläoböden. Beide Lokalitäten weisen eine sehr intensive Pedogenese in ihrem basalen Pedokomplex auf, die wesentlich ausgeprägter ist, als in den überlagernden Paläoböden. Die pedostratigraphischen Ergebnisse lassen eine Einstufung der Älteren Deckenschotter mindestens ins MIS 16 zu.

**Keywords:** *Quaternary stratigraphy, Wels-Aschet, Oberlaab, landscape formation, palaeosols, loess*

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## 1 Introduction

By now, numerous palaeoclimatic curves based on marine drillings give a distinct idea of the number of glacial-interglacial cycles, as well as of the palaeoclimatic evolution of the Middle Pleistocene. Terrestrial studies in Europe are unable to accomplish these targets as they lack dating methods. This is the case especially for the Middle Pleistocene time span in the northern Alpine Foreland. In this context, HABBE (2003) mentions a time gap in numerical datings between the classical Günz and the Upper Pleistocene. The palaeomagnetic boundary at the transition from Lower to Middle Pleistocene ( $\approx 780.000$  yrs) is therefore of crucial stratigraphical importance as it is one of the rare opportunities to date in the middle and lower Middle Pleistocene. The stratigraphical position as well as the ages of the four classical glacial deposits (sensu PENCK & BRÜCKNER 1901–1909) are still under discussion. The Quaternary sediments of the western and the

eastern Alpine Foreland are hardly correlative also because of the use of different stratigraphic systems and because of the fact that terrace bodies can comprise more than one formation period (DOPPLER et al. 2011, SCHELLMANN et al. 2010).

For example, on the one hand, the Matuyama/Brunhes boundary (MBB) could be proved in Günz deposits at the location Heiligenberg (ELLWANGER et al. 1995) for the western Alpine Foreland in Baden-Württemberg. On the other hand, however, the palaeomagnetic reversal occurred within the Younger Deckenschotter (classical Mindel) near Basel (ZOLLINGER 1991), which coincides with pedostratigraphical results (BIBUS 1990). Actually, the stratigraphical system of the Pleistocene Rhineglacier area classifies the formation of the Mindel Deckenschotter as a terrace complex prior to and during the MBB. Furthermore, the Günz Deckenschotter are ranked as Early Pleistocene formations there (ELLWANGER et al. 2011). DOPPLER et al. (2011) developed a stratigraphical scheme for the Bavarian Alpine Foreland, which integrates

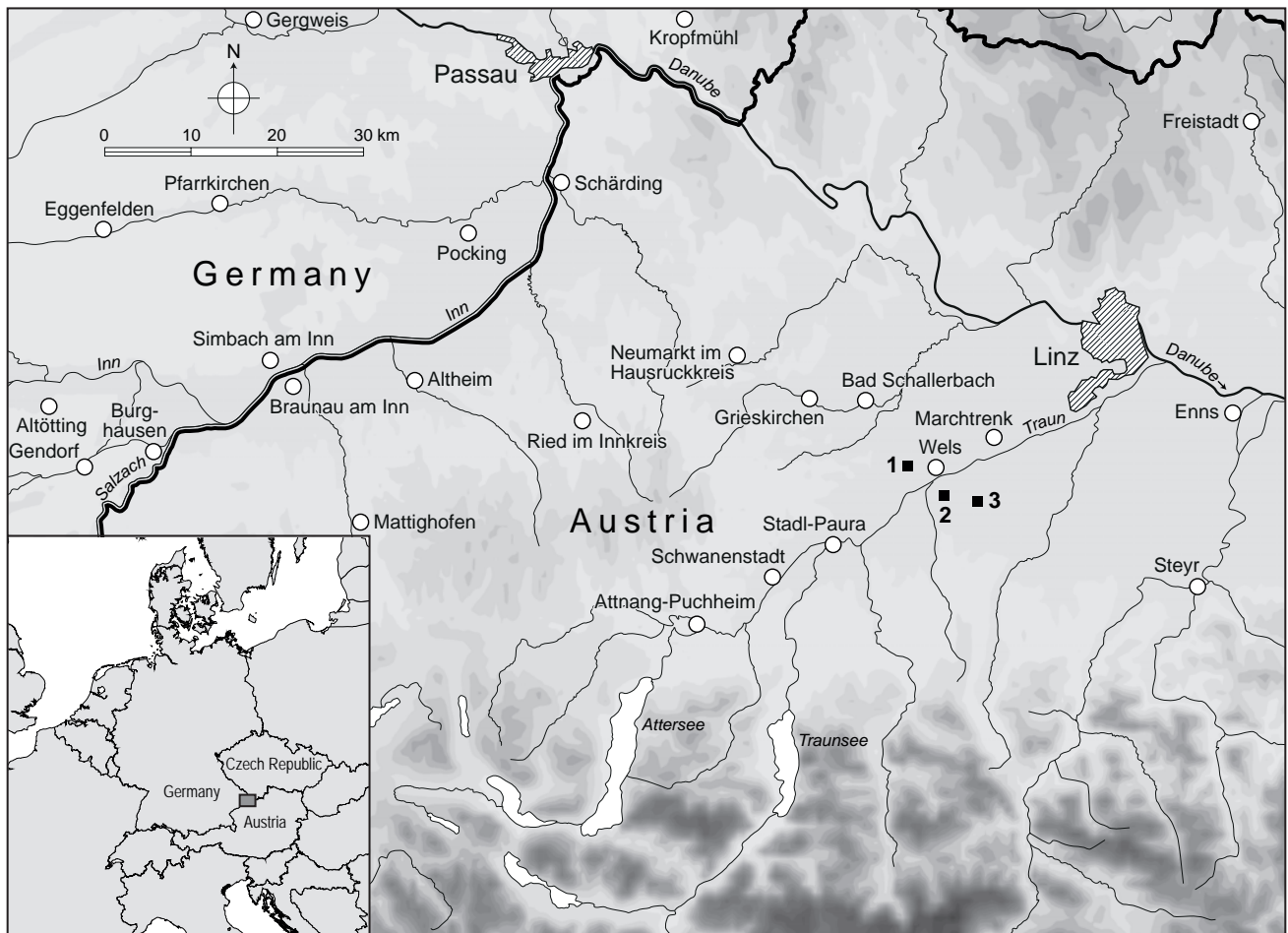


Fig. 1: Study area in Upper Austria. x1 = profile Oberlaab, x2 = profile Wels-Aschet, x3 = profile Neuhofen.

Abb. 1: Untersuchungsgebiet in Oberösterreich. x1 = Profil Oberlaab, x2 = Profil Wels-Aschet, x3 = Profile Neuhofen.

Mindel into the Brunhes-Epoch and Günz sediments mainly to the Matuyama-Epoch. The authors do not exclude that Günz sediments in parts belong to the Brunhes-Epoch. Single datings in the Bavarian foreland obtained with cosmogenic nuclides resulted in  $0.68 \pm 0.23 / -0.24$  Ma for a Mindel deposit and  $2.35 \pm 1.08 / -0.88$  Ma for a Günzian site (HÄUSELMANN et al. 2007). The error rate for the datings are remarkable high. Furthermore, it has to be mentioned that the formation period of Günz, respectively Mindel sediments in Bavaria do not correspond to those in Baden-Württemberg, in general (DOPPLER et al. 2011).

In the northeastern Alpine Foreland of Austria the “Günz complex” is assumed slightly above the MBB (VAN HUSEN 2000, VAN HUSEN & REITNER 2011b), as the transition from reverse to normal magnetisation was not yet verified in any sequence, and the studied sediments show positive magnetisation throughout (cf. KOHL 2000).

A stratigraphic classification of Quaternary forms and sediments in the northern Alpine Foreland above the MBB remains problematic. Absolute ages of Middle Pleistocene deposits are ambiguous (cf. PREUSSER & FIEBIG 2009), and detailed subdivisions are still impossible. This eventually results in the difficulty that even comparably complete Quaternary sequences are not available for a correlation on a supra-regional scale. As a consequence, this is also shown in the stratigraphic chart of the German Stratigraphic Commission (DEUTSCHE STRATIGRAPHISCHE KOMMISSION 2002), which

displays large uncertainties below the Riß complex within the chronostratigraphic correlation of glacial complexes between the northwestern and northeastern Alpine Foreland. The problem further becomes evident in the Austrian stratigraphic chart (ÖSTERREICHISCHE STRATIGRAPHISCHE KOMMISSION 2004).

For these reasons, palaeopedological/pedostratigraphical studies represent an important factor for the study area in assessing the stratigraphic position of the Younger and Older Deckenschotter. SCHOLGER & TERHORST (this volume) proposed a stratigraphic scheme for the loess-palaeosol sequence of Wels-Aschet, which is situated on top of the Günz Deckenschotter. The authors proved Middle Pleistocene palaeomagnetic excursions and classified them on the base of pedostratigraphy. According to the mentioned study, the uppermost section of the Günz Deckenschotter is older than 570–560 ka (Emperor - Big Lost - Calabrian Ridge 3 excursion) and thus is correlated with MIS 16 (SCHOLGER & TERHORST, this volume). The palaeomagnetic approach was applied for the first time in loess sequences in order to establish a chronological framework for Middle Pleistocene in the study area. In this context, it has to be mentioned that MARKOVIC et al. (2009, 2011) applied rock magnetic parameters to build up a Middle Pleistocene stratigraphy for palaeosols and loess sediments of the Middle Danube Basin. The authors were able to correlate their results to marine isotopic curves (see also BRONGER 1970).

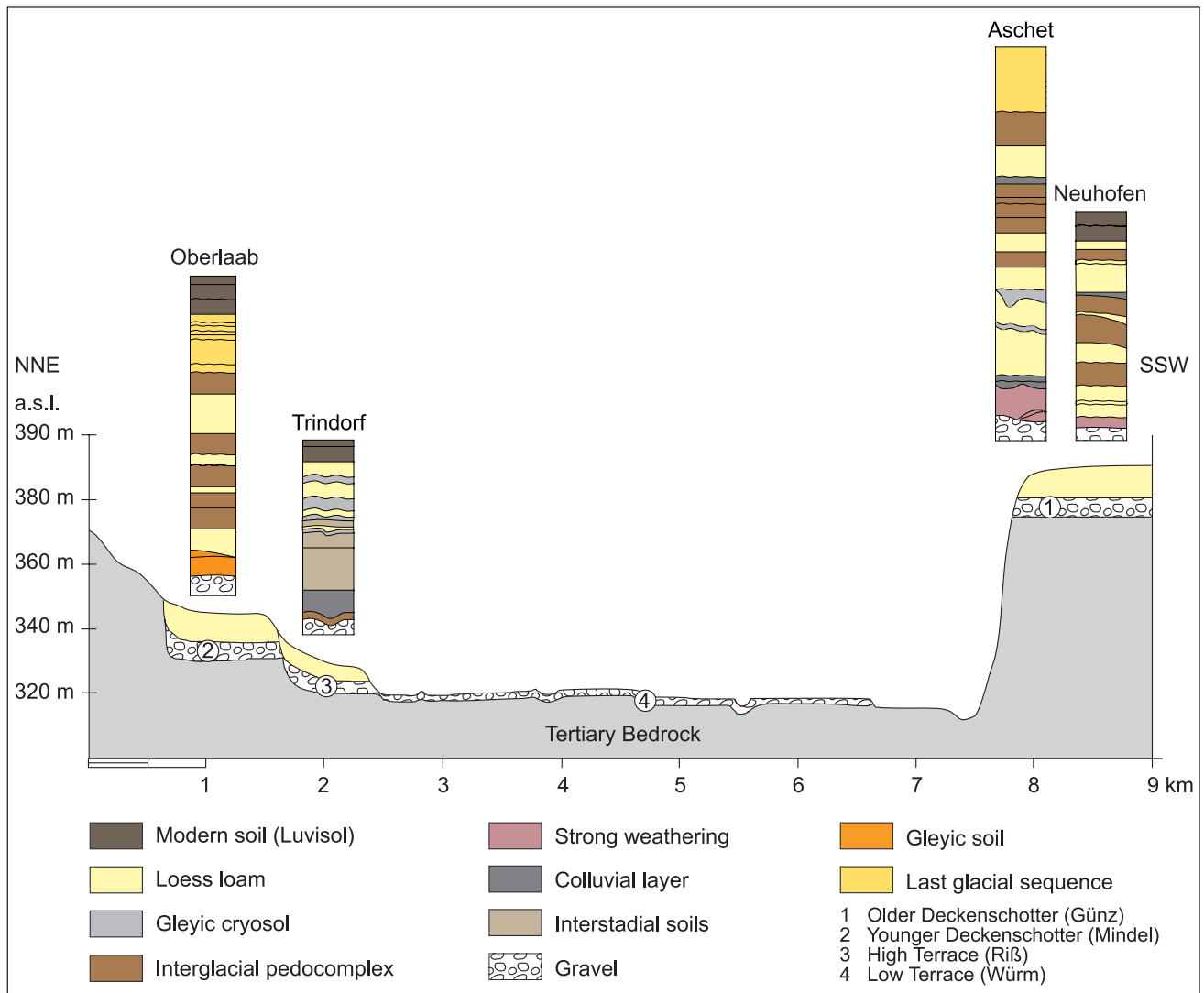


Fig. 2: Geomorphological situation of the presented profiles close to the river Traun/Wels. The uppermost interglacial pedocomplex of all profiles corresponds most probable to the Eemian interglacial soil (see also SCHOLGER & TERHORST, this volume).

Abb. 2: Geomorphologische Situation der präsentierten Profile in der Umgebung der Traun bei Wels. Der oberste interglaziale Pedokomplex entspricht mit hoher Wahrscheinlichkeit jeweils dem Eem-Interglazialboden (vgl. SCHOLGER & TERHORST, dieser Band).

In general, Middle Pleistocene terrestrial stratigraphy is still at its starting point and detailed research with palaeopedological and magnetostratigraphical methods as well as with new luminescence approaches is crucial for the development of a regional terrestrial stratigraphic framework as a first step. Moreover, for future research there is a need i) to correlate terrestrial regional results with each other and ii) to link to results of global palaeoclimatic archives.

This study is intended to propose a Middle Pleistocene stratigraphic framework for Quaternary sediments in the region of Wels.

### 1.1 Study area

The study area is situated on the Traun-Enns plate between Wels and Linz (Fig. 1). The mean annual air temperature is at 9.1°C, precipitation accounts for 821 mm per year. Holocene soils as developed in terraces, loess and loess-like sediments are Luvisols (Parabraunerde) with transitional stages to Stagnosols (Pseudogley).

The sequence of fluvio-glacial terraces is presented in Fig-

ure 2. It shows the spatial situation of the classical morphostratigraphic elements (cf. PENCK & BRÜCKNER 1901–1909). In the lowermost position, two Würmian terraces are present close to the present day floodplain of the river Traun. The Riß terrace in Figure 2 occurs at the left hand side, separated only by a small step to the next higher terrace level of the Younger Deckenschotter (~340m a.s.l.). The position of the Older Deckenschotter in Wels is at ~380 a.s.l. covered by a loess palaeosol sequence of up to 10 m thickness. Profile Neuhofen on top of the Older Deckenschotter is situated 15 km to the east.

For the maximum advances of the younger main glacial phases of Riß and Würm, marine isotopic stages MIS 6 and 2 are correlative (TERHORST et al. 2002). Most probably, Günz and Mindel occur simultaneously with phases of massive global climate decrease during MIS 16 and MIS 12 for the study area (VAN HUSEN & REITNER 2011b).

The studied Younger and Older Deckenschotter are covered with loess and loess loams, reaching a thickness up to 12 m and comprising extensive palaeosols and pedocomplexes.

Studies on Quaternary deposits in the study area are avail-

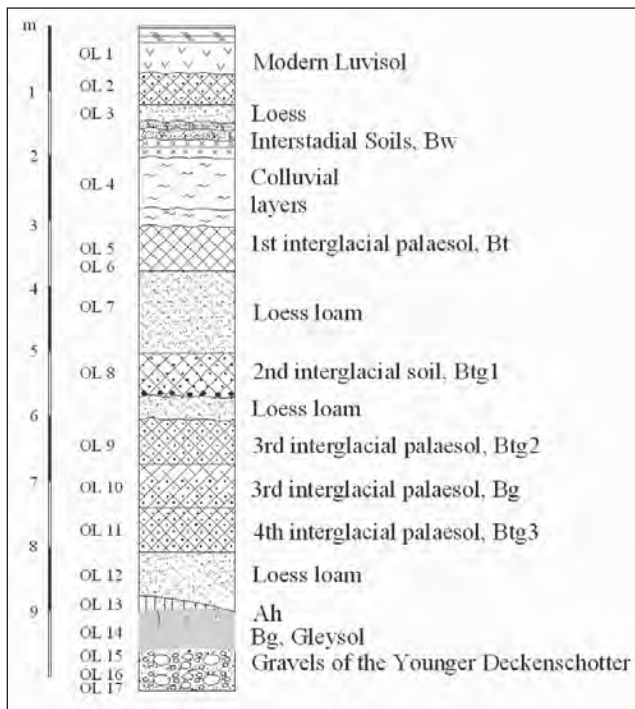


Fig. 3: Detailed palaeopedological profile of the Quaternary sequence Oberlaab on top of the Younger Deckenschotter (Mindel).

Abb. 3: Detailliertes paläopedologisches Profil der quartären Abfolge in Oberlaab auf den jüngeren Deckenschottern (Mindel).

able by KOHL & KRENMAYR (1997), STREMMER et al. (1991) and FINK et al. (1978). Analyses on the sedimentological and mineralogical composition of the cover layers were performed by TERHORST et al. (2003) and TERHORST et al. (2011a, b).

In total, three Middle Pleistocene loess-palaeosol sequences of the Traun-Enns plate are presented, of which one profile is situated on top of the Younger Deckenschotter, while two evolved in the Older Deckenschotter. In general, the studied loess sediments are decalcified loess loams, undergone weathering processes and later redeposition. All described palaeosols have been affected by polygenetic pedogenesis and thus, correspond to pedocomplexes, although the expression partly palaeosol is used.

In order to give the complete view of Quaternary sequences in the study area, the Upper Pleistocene sequence is involved in the stratigraphic scheme, however, not described in detail here (see TERHORST et al. 2002).

## 2 Palaeopedological Results

### 2.1 The cover layers of the Younger Deckenschotter in profile Oberlaab

The loam pit Oberlaab, which is still being quarried, is located about 1.5 km north of Wels and belongs to Pichler brickyard. It represents the only pit currently accessible in the study area, in which the Younger Deckenschotter with superimposed cover layers is exposed.

The walls have been investigated several times and the situation in 2001 is presented by TERHORST (2007). Detailed field analyses were conducted in form of an international cooperation in the year 2003, of which the most complete sequence is presented in SOLLEIRO-REBOLLEDO et al. (this volume).

In 2003, excavated material showed relatively unweathered gravels and sands of the Younger Deckenschotter (Fig. 3, OL 15–17). The terrace gravels are carbonate-free in this location (Fig. 3), and debris from sedimentary rocks is weakly weathered, while crystalline debris shows almost no signs of weathering. However, KOHL (2000) also mentions the occurrence of highly weathered gravels at a former profile section. The terrace deposits are interspersed with distinct bands of iron and manganese.

In the eastern part of the pit on top of the Younger Deckenschotter an intensely greyish Gleysol (OL 14) is encountered, which shows an extraordinarily high number of root traces, and even remains of relic roots occur in places. Immediately on top of it an Ah horizon of up to 60 cm thickness appears locally (OL 13).

The gleyic horizons are covered by a silty, redoximorphic, and also weathered loess loam, which can be traced all over the exposure (OL 12).

The studied wall exhibited before 2003 showed a pedocomplex on top of the above described loess loam with a thickness of up to 2 m (OL 11–OL 9). The latter consists of the basal horizons of two single Stagnic Luvisol pedocomplexes, which were in parts multiply transformed by pedogenesis and correspond to two interglacial phases. Conspicuously, the horizon boundaries are very sharp and linear in some places and could therefore represent erosional surfaces. Interestingly, the single soil horizons (OL 11–OL 9) can be traced by their characteristic features over the entire exposure, which emphasises the independence of each palaeosol. In general, the third Btg horizon is characterised by its intense manganese precipitations and coatings. Additionally, a zone of reduced weathering intensity occurs between the third and fourth fossil soil, which is noticeable by lower clay content with simultaneously higher silt content.

On top of the third Btg horizon a loess loam (OL 8a) follows in some parts of the exposure, which is separated by a stone line from horizon OL 8. OL 8 corresponds to the second well developed Btg horizon and therefore indicates a further interglacial.

The pedocomplex is covered by a loess loam of up to 2 m thickness, which partly was affected by stagnic conditions (OL 7). Based on its stratigraphic position in the profile, the sediment most probably represents a Rißian loess loam.

On top of it the Eemian interglacial soil in form of a Bt horizon is developed (cf. SEDOV et al., this volume).

The western wall of the exposure shows a characteristic sequence of Würmian loess of up to 2 m thickness above the Eemian interglacial soil, which can be compared, even though strongly shortened in this place, to the loess-palaeosol sequences of the Last Glacial, which are present on top of the 'Hochterrasse' (classical Riß) in the study area. At the base colluvial layers are encountered (OL 4). On top of these, weakly redeposited residuals of Middle Würmian interstadials are preserved, covered by thin loess (OL 3).

The modern soil is a Stagnic Luvisol with a thickness of more than 1 m (OL 2, 1).

In conclusion, at least four interglacial pedocomplexes (OL 5/6 and OL 8 to OL 11) can be identified in the cover layers of the presented profile.

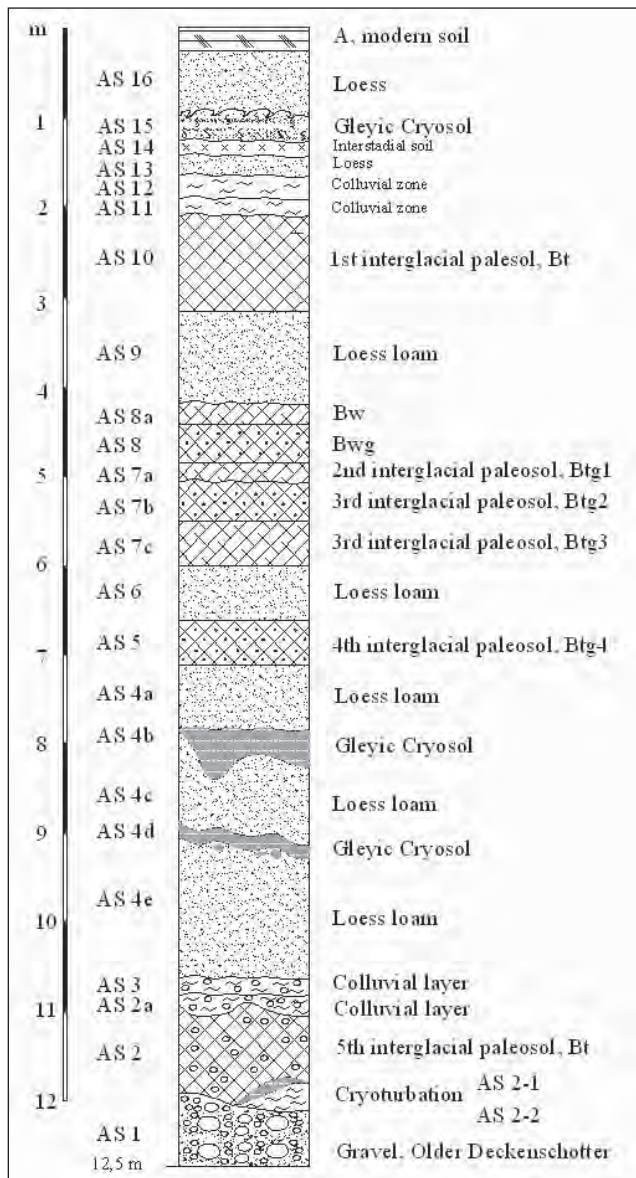


Fig. 4: Detailed palaeopedological profile of the sequence Wels-Aschet on top of the Older Deckenschotter (Günz).

Abb. 4: Detailliertes paläopedologisches Profil der quartären Abfolge in Wels-Aschet auf den Älteren Deckenschottern (Günz).

## 2.2 The cover layers of the Older Deckenschotter in profile Wels-Aschet

Profile Wels-Aschet was dug up and documented anew in 2003 by a working group headed by Prof. Dr. Dirk van Husen. Older studies on the loam pit of Pichler brickyard in Wels-Aschet are available by KOHL & KRENMAYR (1997), KOHL (2000) and STREMMER et al. (1991).

Recently, detailed descriptions are available in the study of TERHORST (2007), TERHORST et al. (2011a, b). Stratigraphical concepts have been presented by PREUSSER & FIEBIG (2009) and SCHOLGER & TERHORST (this volume).

The loess-palaeosol sequence of the studied profile reaches a thickness of 12.5 m. It is mostly decalcified while gravel deposits of the Older Deckenschotter show high lime contents.

The basal section of the profile displays an intensely red-

dish, strong weathering (Fig. 4, AS 2) with dark red clay cutans in the gravels of the Older Deckenschotter (AS 1). Fine earth shows the highest clay content of the profile. In places, the palaeosol is slightly disturbed by cryoturbations.

Above the palaeosol, which represents at least one interglacial, redeposited, gravel-bearing layers are present, which were covered by a 3.5 m thick loess loam (AS 4a–4e), which is disturbed twice by cryoturbation structures. These structures correspond to relic Cryosols and show an intense greyish brown colouring. This double subdivision of the lower loess loam of the Older Deckenschotter was already described by KOHL (2000) for Wels-Aschet, as well as for the former exposure Linz/Grabnerstraße (FINK et al. 1978). Within the loess loam AS 4a a palaeosol developed, which is proved by an intensely stagnic, dark yellowish brown Btg horizon of interglacial intensity (AS 5). The soil texture is characterised by silty clay, and clay cutans are widely distributed across the entire horizon. The former topsoil is eroded.

A further, thin and unstructured layer of loess loam (AS 6) overlies palaeosol AS 5. Above this loam a multiply layered pedocomplex is developed (AS 8a–7c). The basal Btg horizons (AS 7b, 7c) are silty clays and contain distinct reddish brown clay coatings on aggregate surfaces. These two lower horizons of the pedocomplex can be clearly distinguished from the overlying second interglacial palaeosol (Btg horizon, AS 7a) by an erosional unconformity (wavy horizon boundary) and a change in grain sizes. Above all, the difference appears in the enhanced clay content. Additionally, the hydromorphic soil features are significantly weaker. Horizons 8 and 8a above the pedocomplex cannot be interpreted clearly. They show intensive stagnic conditions along former root channels, in which greyish reduced areas are visible.

On top of the pedocomplex loess loam with 1 m thickness is deposited, which is not subdivided any further (AS 9).

Within the loess loam an interglacial soil is developed, which is represented by an intense, 1.10 m thick, dark brown Bt horizon of a former Luvisol (AS 10), in which only weak hydromorphic features are observed. The interglacial soil corresponds largely with the first Bt horizon of the profile Oberlaab described above (TERHORST et al. 2003, SEDOV et al., this volume). This palaeosol horizon is correlative with the MIS 5e (PREUSSER & FIEBIG 2009, TERHORST et al. 2011b).

A Würmian sequence, which in comparison to other profiles is highly shortened, covers the last-mentioned palaeosol. The basal Würmian sediments of profile Wels-Aschet (AS 11, AS 12) correspond to colluvial layers. They are equally interspersed with charcoal as well as rests of the Bt-material of the underlying palaeosol. In the middle section a shortened equivalent of the “Lohne Soil” (acc. to SEMMEL 1968) (AS 14) can be documented, which is preserved as the youngest Middle Würmian interstadial in almost every Upper Austrian loess sequence (TERHORST et al. 2002). In its upper section the “Lohne Soil” is transformed by an intense Cryosol (Reductaquic). The overlying loess (AS 16) is represented by a decalcified, shallow deposit. The recent soil is missing in this profile section of the former loam pit.

Altogether, profile Wels-Aschet contains five palaeosols of interglacial intensity, while strong weathering of the lowermost palaeosol would support also the idea of six interglacial phases recorded in Wels-Aschet.



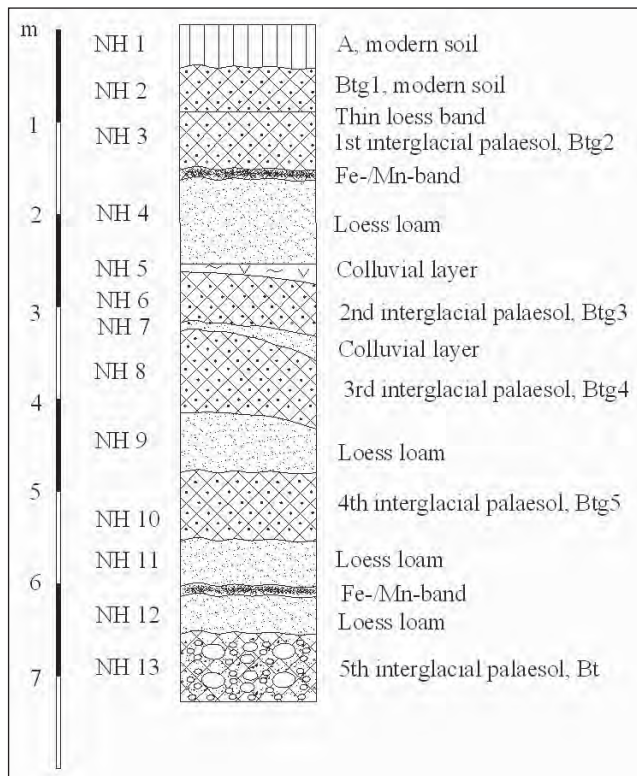


Fig. 5: Detailed palaeopedological profile of the sequence Neuhofen on top of the Older Deckenschotter (Günz).

Abb. 5: Detailliertes paläopedologisches Profil der quartären Abfolge in Neuhofen auf den Älteren Deckenschottern (Günz).

### 2.3 The cover layers of the Older Deckenschotter in profile Neuhofen

The cover layers of the Obermair loam pit in Neuhofen are on average 6.5 m thick, while in places they can reach a thickness of up to 10 m. They lead into a palaeochannel position with a stratified channel filling in northwestern direction. On the southeastern part the lower, extremely weathered Older Deckenschotter was exposed during the past years (Fig. 5).

The lowermost, intensely weathered pedocomplex has developed in the gravels and corresponds to the fifth palaeosol (Fig. 5, NH 13). Judging from its weathering status the soil horizon has to be classified as interglacial, while a formation during more than one interglacial is also probable (cf. TERHORST et al. 2003). Its surface is eroded and covered by a stagnic loess loam of about 1 m thickness (NH 12, NH 11). A distinct iron/manganese band separates this horizon.

Above the loess loam a stagnic Btg horizon has developed (NH 10). Interestingly, this horizon contains locally dark brown to black humic clay cutans, which indicate the former presence of a distinct A horizon in a superimposed position, which is not preserved to date. This Btg horizon within the presented profile corresponds to the fourth interglacial palaeosol.

On top, another loess loam was deposited (NH 9), inside of which a further interglacial palaeosol (NH 8) is developed. This soil horizon is also eroded in its uppermost section. However, the overlying redepositional zone (NH 7) can also be interpreted as marginally redeposited residue of a former E horizon.

Above, NH 6 represents another interglacial palaeosol horizon, which is in places covered by a thin redepositional zone containing soil sediment (NH 5).

A loess loam horizon of up to 1.5 m thickness follows, which is limited to the top by an iron/manganese band in some places of the profile. This band dips into the direction of the channel filling.

The uppermost section of profile Neuhofen is represented by a pedocomplex, which consists of two different superimposed Btg horizons (NH 3, NH 2). Both horizons are at the maximum 1.5 m thick. The pedocomplex is locally separated by a layer of loess loam; therefore two individual soils are recognisable.

### 3 Discussion of stratigraphic results

A highly differentiated structure of the cover layers is present in all Middle Pleistocene sequences presented here. Below a mostly shallow, characteristic Würmian sequence, which is only absent in Neuhofen, the Eemian soil is preserved as uppermost interglacial soil (Fig. 2). This soil usually is substantially less dense and stagnic than the older palaeosols of the warm stages. Underneath, at least three further, in parts highly stagnic, Btg horizons follow in all profiles. Moreover, on top of the Older Deckenschotter, a strongly weathered reddish palaeosol is present.

In general, the palaeosols are developed as pedocomplexes and represent different phases of pedogenesis and geomorphodynamics. The polypedogenetic transformation is in most cases clearly visible macromorphologically.

#### 3.1 Younger Deckenschotter

The loam pit in Oberlaab exposes one of the best structured profiles of cover layers of the Younger Deckenschotter in the entire northern Alpine Foreland. Overall four distinctly developed interglacial soils indicate a minimum age for the gravel surface of the fifth to the last glacial (MIS 12).

On the Younger Deckenschotter in Oberlaab a basal, intensely weathered Gleysol occurs in addition to the existent interglacial soils (cf. OTTNER & SEDOV, this volume). However, the palaeoclimatic meaning of this soil remains unclear. An interglacial formation cannot be fully excluded in this case, however a formation in pre-weathered flood deposits must also be discussed.

Classical ideas according to PENCK & BRÜCKNER (1909) of a formation of the Younger Deckenschotter (Mindel) in the third glacial do not apply for every case in the presented profile; however, a categorisation into the fourth to the last glacial also underestimates their age. KOHL already described three fossil interglacial soils (KOHL & KRENMAYR 1997). Therefore, the classical view on the age of the Younger Deckenschotter being set in the third to the last glacial period was already in question at that time.

Our results support the classification of Mindel Deckenschotter in the Austrian Alpine foreland as belonging to MIS 12, which is in agreement with the palaeogeographical studies of VAN HUSEN & REITNER (2011a) and palaeomagnetic results of SCHOLGER & TERHORST (this volume).

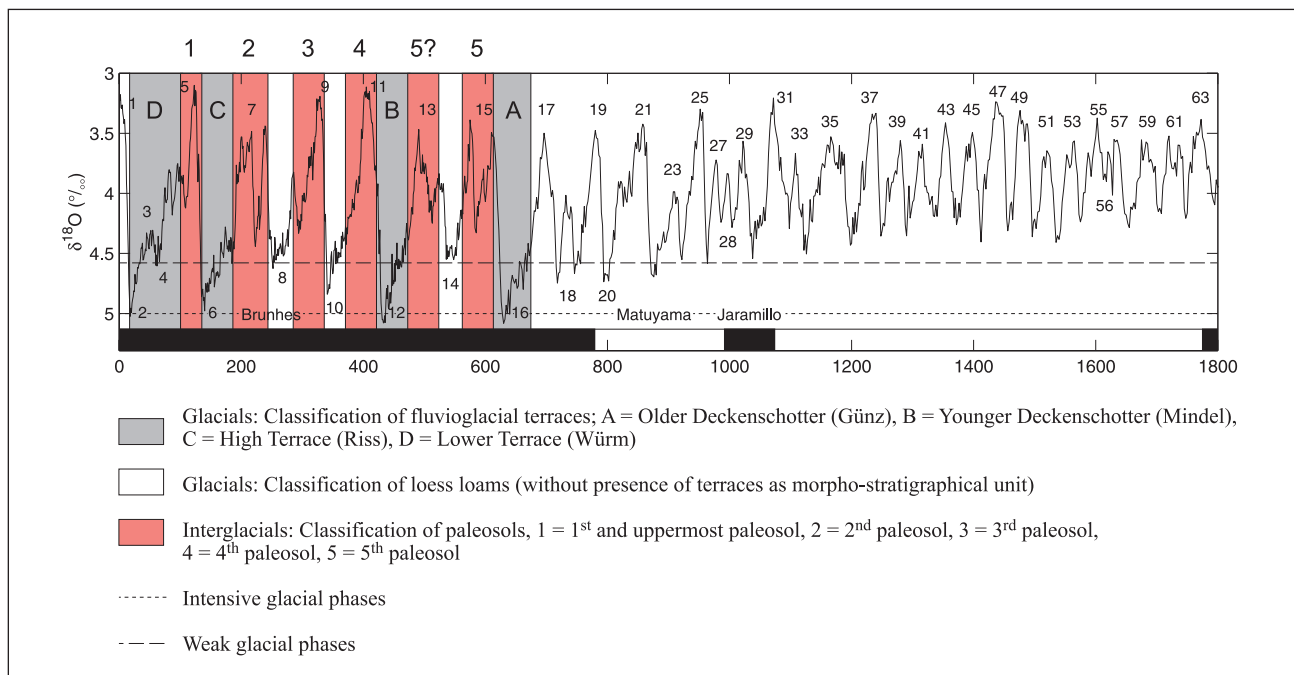


Fig. 6: Classification and correlation of palaeosols, fluvioglacial terraces and loess loams on the base of the marine oxygen isotope curve according to LISIECKI & RAYMO (2005), modified (source: TERHORST et al. 2011b).

Abb. 6: Einstufung und Korrelation von Paläoböden, fluvioglazialen Terrassen und Lösslehmen im Vergleich mit der marinen Sauerstoffisotopenkurve nach LISIECKI & RAYMO (2005), modified (source: TERHORST et al. 2011b).

### 3.2 Older Deckenschotter

The cover layers of the Older Deckenschotter in Wels-Aschet and Neuhofen contain one more interglacial palaeosol than in the Younger Deckenschotter profile of Oberlaab and consequently have to be classified as at least one glacial age older. It can be assumed that the intensely weathered basal palaeosol of both profiles contains not just one interglacial phase. The high degree of weathering in the Bt horizon of the oldest palaeosol (AS 2) is reflected in the intense red clay cutans as well as in the high clay content of about 60.0%. Furthermore, only stable minerals, such as quartz, are present. The clay fraction is dominated by mixed layer minerals, and in this palaeosol vermiculites and illites are completely weathered (TERHORST et al. 2011b).

Thus, the minimum age for the Older Deckenschotter is assumed to belong to the sixth to seventh to the last glacial phase. In this context the fact that the MBB could not yet be proved within the cover layers, leads consequently to a classification in the Brunhes epoch (SCHOLGER & TERHORST 2011). The proof of distinct palaeomagnetic excursions designated to the pedogenic phase of the basal intensely weathered palaeosol of AS 2 might be assigned to MIS15, towards whose end the geomagnetic excursion E-BL-CR3 (Emperor - Big Lost - Calabrian Ridge 3; 570–560 ka) occurs (SCHOLGER & TERHORST, this volume). Therefore, the classification according to VAN HUSEN (2000) can be confirmed by our pedostratigraphical results.

Findings of *Carya* and *Pterocarya* pollen in Wels-Aschet (KOHL 2000) yield a further stratigraphic indicator below the loess loam, which was by the author classified as Rißian. These pollen occur until the Holsteinian Interglacial, which can be parallelised with MIS 11 according to SARNTHEIN et

al. (1986, among others). The results imply a comparably old age of the basal profile sections. This fact as well as palaeopedological results were not considered in the study of PREUSSER & FIEBIG (2009). The above-discussed stratigraphic findings are in contrast to the stratigraphic approach of PREUSSER & FIEBIG (2009), who discuss much younger ages on the base of luminescence dating (IRSL). In this study, ages >200 ka do not significantly increase in the basal parts of the sequence, reflecting that the applied luminescence method can be problematic above 200 ka (cf. PREUSSER & FIEBIG 2009). Recently, methodological studies were able to extend the range of the luminescence method to 300 ka using a new post-IR IRSL protocol (290 °C) (THIEL et al. 2011), an approach not yet applied to the sequence of Wels-Aschet.

At present the Older Deckenschotter are classified as seventh to the last glacial in the Austrian Stratigraphic Chart (ÖSTERREICHISCHE STRATIGRAPHISCHE KOMMISSION 2004).

KOHL (2000) describes already four interglacial soils in Neuhofen and therefore, also in this case, assumes one interglacial more than in classical stratigraphy (acc. to PENCK & BRÜCKNER 1909).

Alltogether, the gravel surface of the Günz gravel body is younger than the MBB. This study classifies it at least to belong to MIS 16 (c.f. Van HUSEN & REITNER 2011b, SCHOLGER & TERHORST 2011).

### 3.3 Comparison of cover layers in the northern Alpine Foreland

Overall, at the present state-of-the-art it is impossible to compare stratigraphic results of the eastern Alpine Foreland with the western area in Baden-Württemberg. The most recent stratigraphic approach and nomenclature of Baden-

Württemberg deviates strongly from the Bavarian and Austrian stratigraphic system (DOPPLER et al. 2011). This fact complicates all attempts to correlate terraces of the eastern with those of the western Alpine Foreland. Furthermore, it has to be taken into account that the described single terrace bodies have to be regarded as terrace complexes, which include groups of terraces (SCHELLMANN et al. 2010). Thus, it is highly probable to obtain different ages in one terrace body.

BIBUS (1995) describes cover layers on the Iller-Lech plate and discusses the fourth or fifth to the last glacial as minimum age for the Younger Deckenschotter. Based on palaeosols in the cover layers of the Younger Deckenschotter in Allschwil brickyard near Basel he even assumes a minimum age of sixth to seventh to the last glacial (BIBUS 1990), which was later confirmed by ZOLLINGER (1991) in the verification of the MBB within the Younger Deckenschotter (cf. also discussion in BIBUS 1995). For the Upper Austrian study area no indication for a classification of the Younger Deckenschotter into seventh to the last glacial could be found so far. Thus, they are presumably younger than in the Rhine glacier area.

A comparison of the Older Deckenschotter in Upper Austria and studies on the Iller-Lech plate (BIBUS 1995) appears difficult because of the differentiated morphological structure of the terraces in the western Alpine Foreland. The Younger Deckenschotter cover layers in Roßhaupten brickyard are best compared to the sequences of the Older Deckenschotter presented here. BIBUS (1995) classifies the Younger Deckenschotter based on at least five to six the palaeosols into the seventh to the last glacial. Also the eighth to the last glacial period could be possible in this place, as the MBB could be proved above the gravels (cf. JERZ 1993). Consequently, a younger age of the Upper Austrian terraces seems to show here, too.

The classification according to pedostratigraphic investigations is consistent with glacial morphological studies in the northern Alpine Foreland by DOPPLER & JERZ (1995) and DOPPLER et al. (2011). However, it has to be taken into account critically that the sequence of the presented profiles is not preserved completely. According to DOPPLER & JERZ (1995) the lower boundary of the Günz complex has to be classified in the MIS 24 (Jaramillo), while the upper boundary is set at the transition of MIS 16 to MIS 15 based on results from the former glacial areas. Regarding the pedostratigraphic results presented here, the Older Deckenschotter of the Traun-Enns plate appear to belong to the younger most section of the Günz complex.

### 3.4 Correlation with marine curves

Regarding the marine oxygen isotope curve according to LISIECKI & RAYMO (2005), around eight interglacial soil formations respectively eight glacials are possible after the MBB, starting with MIS 19 (Fig. 5). However, based on the number of palaeosols in the studied profiles the Older Deckenschotter can be classified at the maximum into MIS 16. The generally intensive, ferretto-like weathering of the Older Deckenschotter in Upper Austria (fifth/basal palaeosol) should therefore correspond to the following interglacial (MIS 15), which is recorded in the palaeomagnetic studies as well (Emperor - Big Lost - Calabrian Ridge 3; 570–560 ka) (SCHOLGER & TERHORST, this volume). According to the results of the

palaeomagnetic analyses, loess loams deposited on top of the MIS 15 palaeosol can be correlated to MIS 14. Subsequently, five interglacials are recorded in marine curves, which should result in a sequence of five palaeosols, respectively, only four of which are clearly present in the study area on the Older Deckenschotter. If we assume that the fifth palaeosol of Wels-Aschet and Neuhofen includes the two interglacials of MIS 15 and 13, the superimposed loess sediment could also belong to MIS 12. The present stage of investigation does not allow a clear classification (SCHOLGER & TERHORST, this volume).

Pursuing these stratigraphical considerations for the Younger Deckenschotter, which also occur consistently in Oberlaab, and additionally regarding the number of palaeosols on the Younger Deckenschotter, MIS 12 or 14 can consequently be considered as probable times of origin for the Younger Deckenschotter. However, it has to be taken into account that MIS 14 is not as pronounced as the other glacials (cf. Fig. 6) and therefore may not be suitable for extensive deposition of gravel in form of an individual morphostratigraphic unit (acc. to PENCK & BRÜCKNER 1909). Further, the age classification of the younger Mindel complex can be brought into accordance with the stratigraphy by DOPPLER & JERZ (1995), who rank classical Mindel to MIS 12. In the study area four interglacial soils are still recorded on the Younger Deckenschotter, whose age of formation has to be assigned to MIS 11, 9, 7 and 5, according to the previous considerations. The intermediary glacial phases apparently did not induce the formation of independent, morphologically visible fluvioglacial terraces in the study area. Only MIS 6 is distinctly present in Rißian upper terrace deposits. The latter has to be classified as Younger Rißian stage according to datings and pedostratigraphic studies (TERHORST et al. 2002). The two older glacials MIS 8 and 10 are not represented by individual morphostratigraphic terraces here. However, during both stages sedimentation of loess loams was active. These are mostly preserved as deposits of more than 1 m thickness. According to the glacial stratigraphy for the northern Alpine Foreland (DOPPLER 2003, DOPPLER & JERZ 1995) the Riß complex can be divided into three glacial phases: Older, Middle and Younger Riß, corresponding to two terrace bodies in the western area of the Pleistocene Salzach glacier.

In this context it is remarkable, that datings by STREMMER et al. (1991) in Wels-Aschet below the Eemian soil yielded ages between 136 ka ( $\pm 13$  ka) and 128 ka ( $\pm 15$  ka). The age of the loess loam below the second palaeosol is stated in the same work as 233 ka ( $\pm 35$  ka) and 245 ka ( $\pm 51$  ka).

## 4 Conclusions

The cover layers of the studied profiles allow for the conclusion that Younger and Older Deckenschotter have to be assessed differently than it is assumed in the scope of the classical morphostratigraphic classification. The pedostratigraphic documentation over many years further proves an older formation age of the Deckenschotter than it has been reported in literature so far. The gravel surface of the Older Deckenschotter in the study area seems to be younger than the MBB and corresponds at least to MIS 16.

The Younger Deckenschotter in Oberlaab are one to two

glacial/interglacial cycles younger than the Older Deckenschotter and can, after detailed discussion, be classified at least as corresponding to MIS 12. The studies further confirm that a largely complete record of palaeosol sequences on the Traun-Enns plate can be expected.

At a supraregional level both conformities and discrepancies become obvious in the stratigraphic classifications. For basic correlations existent dating methods, such as optically stimulated luminescence (OSL), have to be further expanded, and magnetostratigraphic methods have to be applied systematically in further loess sequences.

In future it is desirable and absolutely necessary to establish a chronostratigraphy for the terrestrial Middle Pleistocene all over European loess sequences. This is the only way to enable supraregional or even global correlations. Furthermore, a reliable stratigraphy enables to decode regional landscape formation as well as the response of geosystems to climate change.

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# Magnetic excursions recorded in the Middle to Upper Pleistocene loess/palaeosol sequence Wels-Aschet (Austria)

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**Abstract:** We present the palaeomagnetic examination of the Middle to Upper Pleistocene loess/palaeosol sequence in the former pit 'Würzburger' located in Aschet near Wels. Five strongly developed palaeosols, respectively pedocomplexes alternate with interposing loess loam. By means of excavation access was gained to a profile with a thickness of 12 m. Altogether, 587 orientated samples were taken for the magnetostratigraphic analysis at the Paleomagnetic laboratory of the Montanuniversität Leoben, ensuring a near-complete sampling. The samples were demagnetised by means of alternating field as well as thermal demagnetisation. The magnetic parameters show a sequence of sections with strong magnetite formation in the soil, which can be assigned to the relatively warm climate of interglacial periods. The majority of samples showed characteristic remanence directions within the range of the normal global magnetic field of the Pleistocene. Some sections showed strongly deviating remanence directions, indicating excursions of the global magnetic field. The excursions observed in the Wels-Aschet profile are assigned to the time interval of 570 ka (Emperor-Big Lost-Calabrian Ridge) until 110 ka (Blake) due to palaeopedological-stratigraphical considerations. The Brunhes/Matuyama-Boundary (776 ka) was not reached.

## Magnetische Exkursionen in der Mittel- bis Oberpleistozänen Löss-/Paläoboden-Sequenz Wels-Aschet

**Kurzfassung:** Die vorliegende Arbeit stellt die paläomagnetische Bearbeitung der Mittel- bis Oberpleistozänen Löss-/Paläoboden-Sequenz im Areal der ehemaligen Ziegelei Würzburger in Aschet bei Wels vor. Fünf intensiv entwickelte Paläoböden, bzw. Pedokomplexe wechseln mit dazwischen geschalteten Lößlehmlagen ab. Im Rahmen einer Aufbaggerung konnte ein Profil mit einer Mächtigkeit von über 12 m erschlossen werden. Für die magnetostratigraphischen Laboruntersuchungen im Paläomagnetiklabor der Montanuniversität Leoben wurden insgesamt 587 orientierte Proben entnommen, so dass eine beinahe lückenlose Beprobung vorliegt. Die Proben wurden mit magnetischen Wechselfeldern sowie thermisch abmagnetisiert. Zur Bestimmung der magnetischen Trägerminerale in den Sedimenten wurden Curiepunkt-Bestimmungen durchgeführt, die eine Hauptträgerphase mit einem Curie-Punkt bei ca. 580°C (Magnetit), sowie untergeordnete Anteile von Hämatit mit 670°C Curie-Punkt ergaben. Die magnetischen Parameter zeigen eine Folge von Bereichen mit intensiver Magnetitbildung in den Paläoböden, die dem relativ wärmeren Klima von Interglazialen zugeordnet werden können. Die Mehrzahl der Proben zeigen charakteristische Remanenzrichtungen im Bereich des normalen pleistozänen Erdmagnetfeldes. In einigen Profilabschnitten traten stark abweichende Remanenzrichtungen auf, die auf Exkursionen des Erdmagnetfeldes hinweisen. Die beobachteten Exkursionen im Profil Wels-Aschet werden aufgrund paläopedologischer-pedostratigraphischer Ergebnisse in das Zeitintervall von 570 ka (Emperor - Big Lost - Calabrian Ridge) bis 110 ka (Blake) gestellt. Die Brunhes/Matuyama-Grenze (776 ka) wurde nicht erreicht.

**Keywords:** *Pleistocene, loess, palaeosol, magnetic excursion, chronostratigraphy, rock magnetic properties, Upper Austria*

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## 1 Introduction

### 1.1 Stratigraphic framework of the study area

The Older Deckenschotter/Günz terraces of the Traun-Enns-Platte are classified according to stratigraphic considerations as belonging to the seventh-last glacial period. In this context VAN HUSEN (2000) and VAN HUSEN & REITNER (2011) proposed a chronostratigraphic framework for the glacial deposits and processes during the Middle Pleistocene of the Eastern Alps. In this study we assume that the four classic glacial periods are to be assigned to the Brunhes-Chron.

VAN HUSEN (2000) assumes that Günz is to be correlated with MIS16 and Mindel with MIS12. Furthermore, the author places the Riß-Glacial in MIS 6 and the Würmian Glacial in MIS 2. On the basis of cover layer analyses in the Linz/Wels area, TERHORST (2007) and TERHORST et al. (2011, 2012) arrive at similar results. The authors also correlated Günz with MIS16 and Mindel with MIS12. Earlier investigations in the cover layers of fluvio-glacial terraces in the Linz/Wels area clearly show that too many interglacial soils occur to support the classic conceptions according to PENCK & BRÜCKNER (1909) (KOHL & KRENMAYR 1997, STREMMER et al. 1991).

## 1.2 Palaeomagnetic investigations of loess/palaeosol sequences

Climate change has an impact on weathering and sedimentation processes which affect the magnetic minerals within rocks (DEKKERS 1997). Therefore magnetic parameters such as magnetic susceptibility are sensitive indicators for climate dependant variations of the type, grain size and concentration of magnetic minerals in sediments (THOUVENY et al. 1994, VEROSUB & ROBERTS 1995). The magnetic susceptibility of rocks depends on the concentration and grain size of the magnetic mineral phases, which are part of the rock composition. In general, magnetite has the highest importance due to its high mineral specific susceptibility, and magnetic susceptibility can be used as a proxy for the magnetite content of rocks (THOMPSON & OLDFIELD 1986).

Several extensive investigations of European and Asian loess profiles document that palaeosols mainly contain finely grained magnetite as carrier mineral of the magnetic properties, while the magnetic components of loess samples are characterised by larger single-domain and multidomain grain sizes (EVANS & HELLER 2003). Since the magnetic susceptibility of SP-magnetite is significantly higher, it is possible to use the temporal variation of these parameters as an indicator for the intensity of soil formation (HELLER & LIU 1984). Model hypotheses assume that changing redox conditions in the soil during the weathering of ferrous minerals lead to the release of iron and the regeneration of ferrihydrite. At the end of a reaction chain, SP-magnetite or maghaemite regenerate in the soil under oxidising conditions (MAHER 1998). CHEN et al. (2005) refer to the importance of the organic content in connection with the activity of iron-reducing bacteria.

To date only few palaeomagnetic results of loess/palaeosol sequences in Austria are available. Earlier investigations of the clay pit of the brickyard Würzburger in Aschet/Wels are described in FINK et al. (1976). For the first attempt at a palaeomagnetic dating of these sediments, samples were taken from the cover layer above the older sheet gravels every 20 to 40 cm, which all showed a positive inclination and a declination of approximately zero. Sedimentary processes at the base of the cover layers are given as a possible reason for the absence of the expected polarity reversal (Brunhes-Matuyama Boundary) as well as the strong variation in declination values in the lowest part of the profile (FINK et al. 1976). The profile Wels-Aschet was dug up and documented anew in 2003 by a working group headed by Prof. Dr. Dirk Van Husen (SCHOLGER & TERHORST 2011).

## 2 Palaeopedological setting of the Wels-Aschet sequence

Older studies on the loam pit of Pichler brickyard in Wels-Aschet are available by KOHL & KRENMAYR (1997), KOHL (2000) and STREMMER et al. (1991). A detailed description and sedimentological analyses can be found in TERHORST et al. (2011, 2012).

Altogether, the loess/palaeosol sequence of the studied profile reaches a thickness of 12.5 m. It is mostly decalcified while gravel deposits of the Older Deckenschotter show high CaCO<sub>3</sub> contents. The sequence contains five palaeosols of

interglacial intensity, while strong weathering of the lowermost palaeosol is especially conspicuous. The basal sections of the profile display an intensely reddish-brown weathered horizon (AS 2) with dark red clay cutans in the gravels of the Older Deckenschotter (AS 1). Above the palaeosol, which represents at least one interglacial period, redeposited, gravel-bearing layers are deposited, which were later covered by a 3.5 m thick loess loam (AS 4a–4e), which intercalates two Cryosols. On top of the loess loam (AS 4a) a palaeosol developed in form of a dark yellowish brown Btg horizon of interglacial intensity (AS 5).

A further, thin and unstructured layer of loess loam (AS 6) overlies palaeosol AS 5. Above this loam a multiply layered pedocomplex is developed (AS 8a to 7c). The basal, 3<sup>rd</sup> palaeosol (AS 7b, 7c) contains distinct reddish brown clay cutans. This pedocomplex can be clearly distinguished from the overlying 2<sup>nd</sup> Btg horizon (AS 7a) by an erosional unconformity and a change in grain sizes (TERHORST et al. 2011). Two stagnic horizons (AS 8 and 8a) are superimposed on the described pedocomplex. On top of this pedocomplex, a loess loam with 1 m thickness is present, which is not subdivided any further (AS 9). Within the loess loam an interglacial soil is developed, which is represented by an intense, 1.10 m thick, dark brown Bt horizon (AS 10). This palaeosol horizon is correlative with the MIS 5e (PREUSSER & FIEBIG 2009, TERHORST et al. 2011). A Würmian sequence, which in comparison to other profiles is reduced and poorly weathered, covers the Eemian palaeosol (AS 11 to AS 16).

## 3 Methods

By means of two excavations (2001, 2003) access was gained to a profile of a thickness of more than 12 m. Samples for the palaeomagnetic laboratory analysis were collected using plastic cubes (edge length 2cm) which were pushed into the sediment either by hand or hammer along a guide rail. The orientation of the samples was measured according to routine palaeomagnetic proceedings. 162 (2001) and 425 (2003) samples were taken for the magnetostratigraphic laboratory analyses. The undetermined vertical distance between individual samples lies between 0 and 2 cm, ensuring a near-complete sampling. The continuous in situ measuring of the magnetic volume susceptibility in the profile was performed with an Exploranium KT-9 Kappameter.

Magnetic volume susceptibility as well as natural remanent magnetisation were measured at the Paleomagnetic Laboratory of the Montanuniversität Leoben as reference values for further investigations. The remanence measurements were performed with a 2G-Cryogenic-magnetometer with integrated alternating field demagnetisation, the susceptibility measurements on a Kappabridge Geofyzika KLY-2.

We first demagnetised the orientated samples with increasingly strong alternating magnetic fields (from 2 to 140 mT), and subsequently thermally demagnetised selected samples at temperatures ranging from 200° to 550°, the remaining NRM being measured after every step of the cleaning process. The demagnetisation behaviour indicates the magnetic carrier minerals and allows the separation of primary fossil magnetisation directions from secondary magnetisations created by magnetic overprints and weathering.

To determine the magnetic carrier minerals in the sediments, we performed Curie point determinations and examined the magnetic saturation behaviour. Depending on the deposit and formation conditions of the sediments and soils, different significance to the various magnetic parameters connected to certain sedimentologic or climatic processes is assigned. For the present study we chose the parameter AF@IRM according to LARRASOANA et al. (2003), which is regarded as an indicator for the haematite content within rocks. We first exposed the samples to a magnetic saturation at 0.9T and subsequently demagnetised them in an alternating field at 100mT. The remaining magnetisation is most likely due to haematite. At the same time we measured the anisotropy of the magnetic susceptibility (AMS), which determines the fabric of the magnetic minerals within the rock.

## 4 Results

### 4.1 Carrier minerals of the magnetic properties

The results of the Curie point determination yielded a magnetic carrier phase with a Curie point of approximately 580°C (magnetite) as well as a subordinate haematite content with a Curie point of 670°C. An increase in magnetic susceptibility between 280°C and 350°C most likely results from a transformation of iron sulphides or iron hydroxides into secondary magnetite or maghaemite during the heating process (GENDLER et al. 2006). The magnetic saturation indicates various magnetic minerals in the loess sediments and palaeosols of the Aschet profile. Samples from the palaeosols are characterised by a relatively low coercivity and reach magnetic saturation faster than the samples from loess horizons. The saturation behaviour of the samples indicates variations of low- and highly-coercive magnetic minerals in different concentrations. A highly coercive mineral (haematite) is present in low concentrations in all samples; the palaeo contain at least one additional low-coercive phase (magnetite) which developed during soil formation.

The measurements of the anisotropy of the magnetic susceptibility showed weakly anisotropic and in many cases isotropic magnetic fabric for all sections of the Aschet profile. The values for the magnetic fabric parameters, lineation (L) and foliation (F), lie below 2% for most samples (median value and standard deviation from 165 samples:  $L = 1,014 \pm 0,018$ ;  $F = 1,018 \pm 0,017$ ). We could not observe a preferred orientation of the Kmax direction, which would allow inferences as to palaeowind direction.

### 4.2 Characteristic magnetisation directions

The stepwise demagnetisation in an alternating field was performed in 6 to 15 stages ranging from 2 to 140mT (Figure 1). Considering the mineral magnetic results, the lower field range was examined in more detail to isolate those remanence vectors which reside in different phases of magnetite. In general we only observed a low impact of the highly coercive phases on demagnetisation behaviour. For the majority of samples the remanence could be demagnetised completely using alternating field demagnetisation. After demagnetisation in the alternating field, some samples showed remaining remanences which are due to highly coercive phases (e.g. haematite). We subsequently thermally demagnetised these

samples at temperatures ranging from 200°C to 550°C. We generally observed two magnetic phases which in most cases however yielded the same remanence direction (Figure 1 a,c).

We observed similar demagnetisation vectors in every loess and soil sample. The majority of samples showed characteristic remanence directions within the range of the normal recent or Pleistocene global magnetic field. In some sections of the profile strongly deviating remanence directions occurred (Figure 1 b,d), which indicate excursions of the global magnetic field which also occurred repeatedly during the Pleistocene (LUND et al. 1998). However, we did not observe a complete polarity reversal of the magnetic field indicating the Brunhes-Matuyama Boundary.

Up to 90% of the natural remanent magnetisation (NRM) in the Bt horizon of the 1<sup>st</sup> palaeosol (AS 10) as well as the subjacent loess loam (AS 9) were already demagnetised at low alternating fields. The subsequent thermal cleaning up to 550°C yielded further demagnetisation of the same directional component (Figure 1a). In the upper section of the 1<sup>st</sup> palaeosol (AS 10) at the transition to the colluvial layer (AS 11), we observed strongly deviating remanence directions within an interval of approximately 20 cm in the profile (1.90 to 2.10 m). Similar deviating magnetisation vectors occurred in the lower part of the loess loam AS 9 at a depth of 3.60 to 3.95 m as well as in the 2<sup>nd</sup> and 3<sup>rd</sup> palaeosols (Btg horizons, AS 7a, b) at a depth of 4.55 to 5.34 m (Figure 1b). The samples taken from AS 7 are characterised by a higher coercivity. The subsequent thermal cleaning resulted in a further loss of intensity at the temperature interval between 500°C and 550°C, which indicates magnetite as the sole carrier mineral of the remanence. In the layers AS 4 the demagnetisations indicated an additional magnetic carrier mineral both in the loess loam as well as in the Cryosols. The alternating field demagnetisation showed that only approx. 50% of the NRM were linked to lower coercive magnetic phases, most part of the remaining intensity could be removed during the first stages of thermal cleaning (Figure 1c). The samples taken from the 5<sup>th</sup> palaeosol (Bt horizon, AS2) were similar to the 2<sup>nd</sup> palaeosol in their demagnetisation behaviour. We were able to demagnetise the majority of NRM at alternating field strengths of up to 40 mT; Figure 1e). We mainly observed normal characteristic remanence directions for the whole lower part of the profile, between AS 1 and AS6, while deviating magnetisation vectors only occurred in a few samples taken from two depth sections (at approximately 9.5 m and 11.6 m; Figure 1d).

### 4.3 Variation of magnetic parameters in the profile

Magnetic susceptibility, NRM intensity, saturation intensity and the parameter AF@IRM (LARRASOANA et al. 2003), which is used as an indicator for haematite, all showed significant variations in the profile (Figure 2). The continuous susceptibility profile recorded in situ during the sampling and the measurements of the samples show a strong link between magnetic susceptibility and lithologic conditions, with the palaeosols generally emerging as areas of increased susceptibility with concurrent large variation in susceptibility values. The susceptibility values for loess loam (approx.  $150 \cdot 10^{-6}$  SI) represent the magnetic background value of the sediment, while the values in the palaeosols are as much as



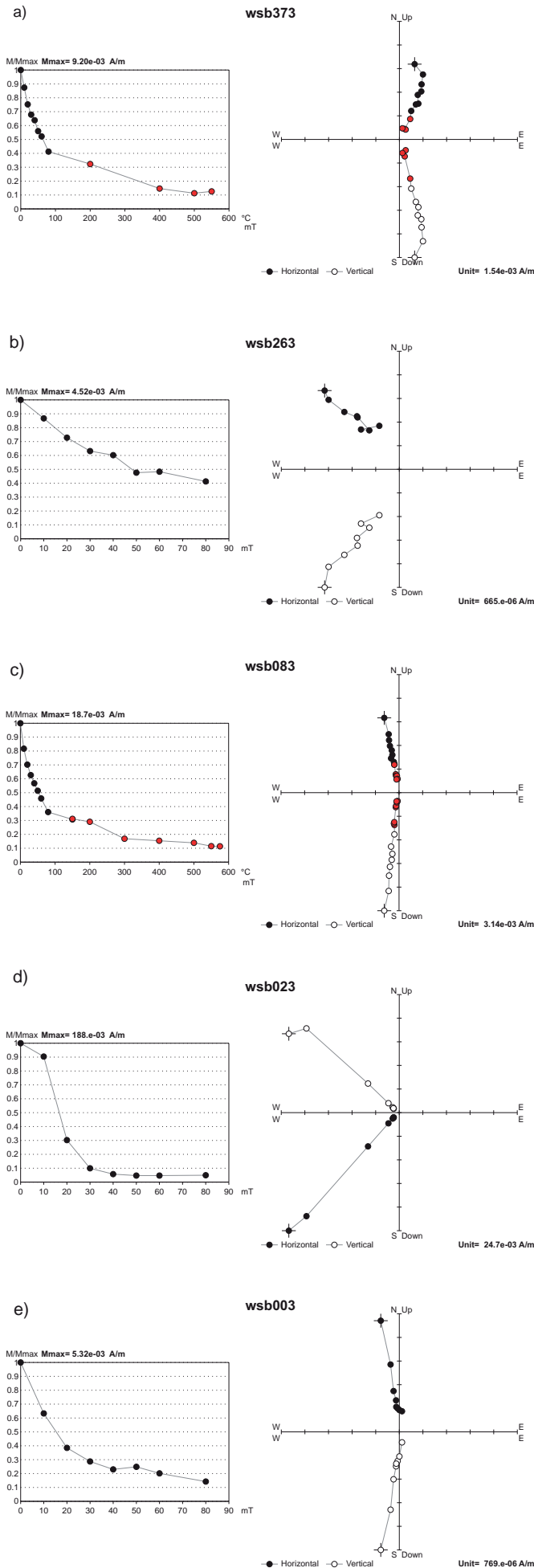


Fig. 1: Vectorplots (horizontal und vertical projection) and intensity decay during alternating field-(black) and thermal (red) cleaning. (a) WSB373 of AS 10, (b) WSB 263 of AS 7, (c) WSB83 of AS 4e, (d) WSB23 of AS 2, (e) WSB 3 of AS 2. Alternating field demagnetisation in 6 to 15 steps in the range between 2 and 140 mT, successive thermal demagnetisation at temperatures between 200°C and 550°C (a, c). In some sections of the profile strongly deviating remanence directions occurred (b,d), which indicate excursions of the global magnetic field.

Abb. 1: Vektordiagramme und Intensitätsverhalten ausgewählter Proben bei der Abmagnetisierung. Horizontale und vertikale Projektion der Remanenzvektoren, sowie Intensitätsverlauf bei der Wechselfeld- (schwarz) und thermischen (rot) Reinigung. (a) WSB373 aus AS 10, (b) WSB 263 aus AS 7, (c) WSB83 aus AS 4e, (d) WSB23 aus AS 2, (e) WSB 3 aus AS 2. Abmagnetisierung im Wechselfeld in 6 bis 15 Schritten im Bereich zwischen 2 und 140 mT, nachfolgende thermische Abmagnetisierung bei Temperaturen von 200°C bis 550°C (a, c). In einigen Bereichen des Profils traten stark abweichende Remanenzrichtungen auf die auf Exkursionen des Erdmagnetfeldes hinweisen (b, d).

ten times higher due to magnetite-formation during pedogenesis. The intensity values of NRM and SIRM largely follow the same trend. Within the loess content of the profile, the haematite indicator AF@IRM shows similar behaviour to the other magnetic parameters. In contrast, we did not observe an increase of the AF@IRM values in the palaeosols. This confirms the hypothesis that the increased values of susceptibility, NRM and SIRM in the palaeosols are due to higher magnetite content.

The intensely weathered reddish palaeosol in the basal section of the profile (12 m–11 m) with the highest loam content in the profile represents at least one interglacial period and is slightly remoulded by cryoturbation (TERHORST 2007). In this section (AS 2) the susceptibility values are generally increased. A susceptibility and SIRM-maximum at 11.5 m and diverging remanence directions may indicate a short-term excursion of the global magnetic field during soil formation or during a subsequent intense weathering period. In the superjacent loess loam complex with a thickness of 4m (AS 4a–4e), the saturation magnetisation in particular indicates an increased magnetite content in the two Cryosols. At 9.5 m we observed an indication for a magnetic excursion. In the Btg horizon (AS 5) between 7.0 m and 6.6 m, only individual samples show increased magnetic susceptibility, while the superjacent unsegmented loess loam layer AS 6 (6.6 m–6.0 m) is characterised by significantly increased susceptibility values. Within the multi-segmented pedocomplex AS 8a–7c (6 m to approx. 4.2 m), with greyish iron-reduced areas along visibly developed root traces (TERHORST 2007), increased susceptibility and SIRM values also indicate partially intense magnetite formation, which can strongly vary locally depending on the availability of organic substances. In this pedocomplex we observed a magnetic excursion over a longer profile section, the magnetisation directions in the lower part (AS 7b) indicating a possible additional subdivision.

## 5 Discussion

Saturation intensity and magnetic susceptibility strongly vary within the profile, depending on the different lithologic horizons. Several horizons with strongly increased susceptibility values occur, which are due to higher concentrations

## Profile Wels-Aschet

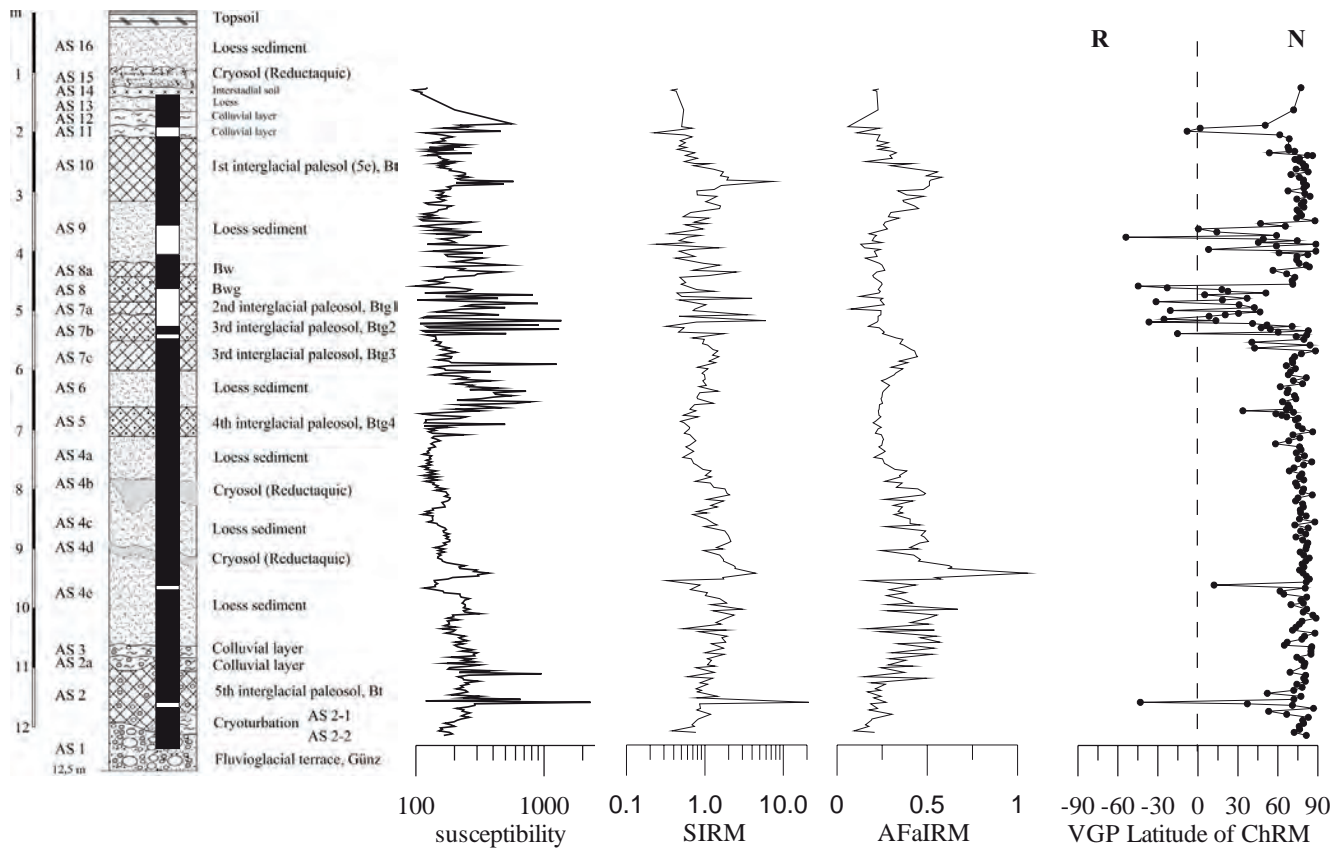


Fig. 2: Variation of magnetic parameters in the profil Wels-Aschet. Magnetic susceptibility (in  $10^6$  SI), intensity of saturation magnetisation (SIRM in A/m) and remaining remanence after successive alternating field demagnetisation (AFaIRM), VGP-Latitude calculated for the characteristic remanence component (ChRM). R, N: Inverse und Normal polarity. Geomagnetic Instability column for the profile Wels-Aschet (black: stable North direction, white: magnetic excursion).

Abb. 2: Magnetische Parameter der Proben aus dem Profil Wels-Aschet. Magnetische Suszeptibilität (in  $10^6$  SI), Sättigungsmagnetisierung (SIRM in A/m) und Rest-Remanenz nach anschließender Abmagnetisierung im Wechselfeld (AF@IRM), VGP-Breitengrad der charakteristischen Remanenzkomponente (ChRM). R, N: Inverse und Normale Pollage. Geomagnetische Zuordnung im Profil (schwarz: stabile Nordrichtung, weiß: magnetische Exkursion).

of finely grained magnetite. Such horizons indicate climate-regulated in situ soil formation during relatively warm climatic conditions. Investigations of loess profiles in China showed that palaeosol horizons are always characterised by high magnetic susceptibility and high saturation intensity (HELLER & LIU 1984, HUNT et al. 1995). EVANS & HELLER (1994) proved that the magnetic characteristics are due to magnetite and pyrrhotite, which were created by pedogenesis. For attempting a chronostratigraphic classification, the magnetic susceptibility of profile sections is used as a climate indicator and compared to the oxygen isotope curve according to LISIECKI & RAYMO (2005). For the chronostratigraphic classifications in the Aschet profile near Wels we used a current survey of internationally recognised observed excursions within the Brunhes-Chron (LAJ & CHANNELL 2007, LANGEREIS et al. 1997, SINGER et al. 2008). Geomagnetic excursions represent short-term deviations from the normal global magnetic field, which can be observed as trans-regional phenomena (SINGER et al. 2008) and are therefore suitable for chronostratigraphic correlation. Magnetic excursions or polarity reversal within the assumed stable Brunhes-Chron were first discovered at Lavas located in the French

Massif Central (Laschamp, approx. 40 ka) (BONHOMMET & BABKINE 1967). Since then numerous magnetic excursions in sediments and vulcanites, whose existence and chronostratigraphic age determination are not always beyond doubt, have been described worldwide. The various authors primarily disagree concerning the assessment of the relevance of data taken from drill cores as well as the importance of independent geochronologic datings of the examined rocks (LAJ & CHANNELL 2007). According to LUND et al. (1998) a total of 15 magnetic excursions can be observed in ODP cores within the Brunhes-Chron. According to LANGEREIS et al. (1997) 12 excursions are to be considered, 6 of which are referred to as well dated, globally observed excursions. According to SINGER et al. (2008), all excursions which have been dated radiometrically via the Ar-dating of vulcanites or astrochronological via O-isotopes in the sediments need to be considered. In contrast, LAJ & CHANNELL (2007) consider only those 7 excursions for which worldwide proof and precise dating are available: Mono Lake (33 ka), Laschamp (40 ka), Blake (120 ka), Iceland Basin (188 ka), Pringle Falls (211 ka), Big Lost (560–580 ka) and Stage 17 (670 ka).

Above the Older Deckenschotter between 11.0 and 12.0 m

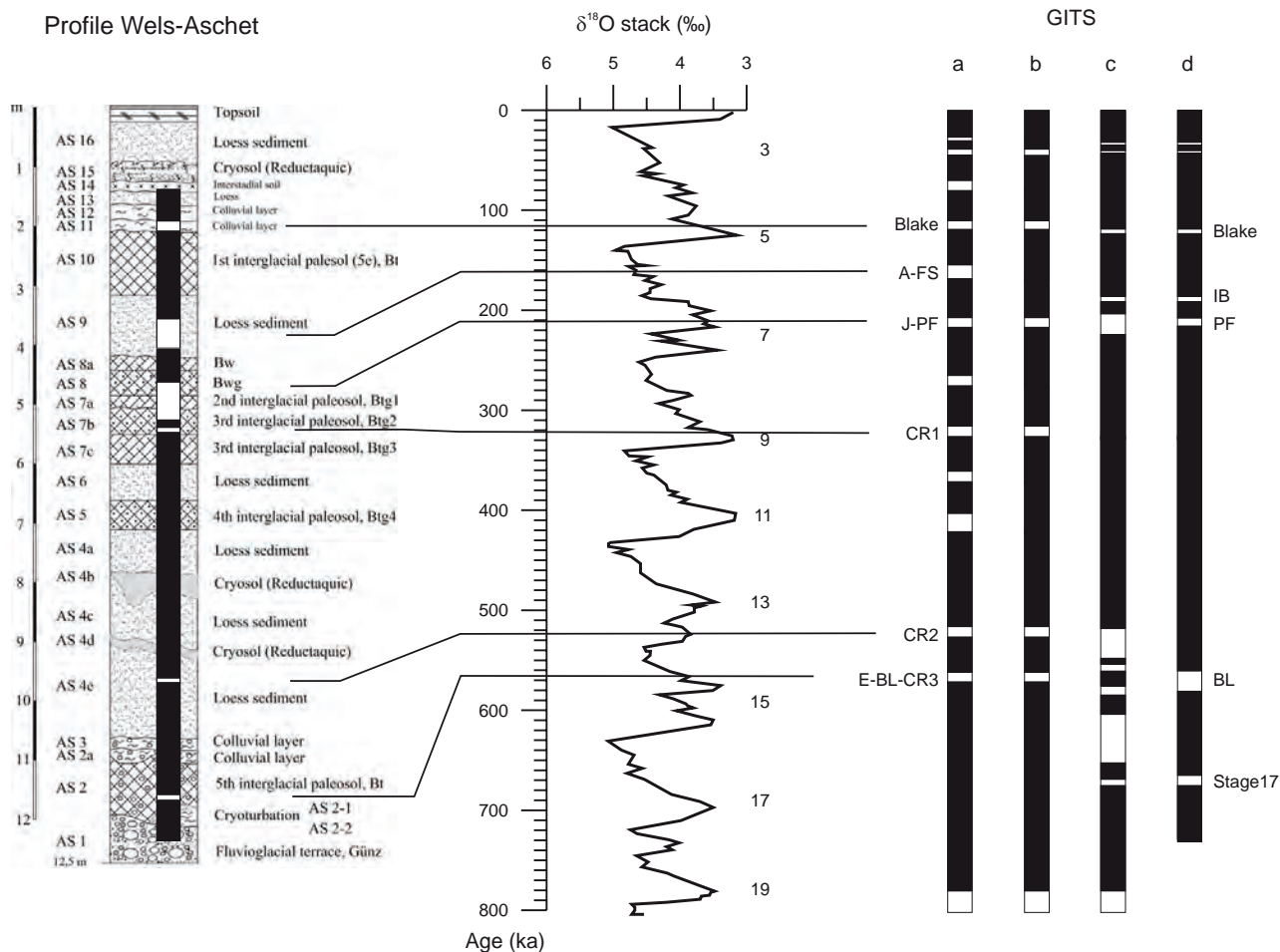


Fig. 3: Tentative correlation of magnetic parameters in the profile Wels-Aschet to chronostratigraphic reference scales. Oxygen isotopes after LISIECKI & RAYMO (2005) and Geomagnetic Instability Time Scale (GITS; (a) and (b) after LANGEREIS et al. (1997), (c) after SINGER et al. (2008), (d) after LAJ & CHAN-NEL (2007). Excursions (reversals) named after localities: Albuquerque (A), Fram Strait (FS), Iceland Basin (IB), Pringle Falls (PF), Calabrian Ridge (CR), Emperor (E), Big Lost (BL). Palaeopedologic profile after TERHORST et al. (2012).

Abb. 3: Vergleich der magnetischen Parameter im Profil Wels-Aschet mit chronostratigraphischen Referenzdaten: Sauerstoffisotopen nach LISIECKI & RAYMO (2005) und Geomagnetic Instability Time Scale (GITS; (a) und (b) nach LANGEREIS et al. (1997), (c) nach SINGER et al. (2008), (d) nach LAJ & CHAN-NEL (2007). Bezeichnung der Excursionen bzw. Umpolungen nach Fundorten: Albuquerque (A), Fram Strait (FS), Iceland Basin (IB), Pringle Falls (PF), Calabrian Ridge (CR), Emperor (E), Big Lost (BL). Paläopedologisches Profil aus TERHORST et al. (2012).

of the profile a palaeosol (AS 2) has developed, which represents at least one interglacial period and, at its lowest section at 11.6 m in the profile, contains a magnetic excursion. The pedogenic phase might be assigned to MIS15, towards whose end the geomagnetic excursion E-BL-CR3 (Emperor - Big Lost - Calabrian Ridge 3; 570–560 ka) occurs. Above the palaeosol, colluvial, gravel bearing layers were deposited (AS 3, AS 2a). The superjacent thick loess loam (AS 4) contains a further magnetic excursion in the section under a the Cryosol AS 4d at 9.5 m in the profile, for whose geological age two models need to be discussed. If the excursion is correlated with Calabrian Ridge 2 (CR2; 525–515 ka according to LANGEREIS et al. (1997)) or West Eifel 5 (528+/-16ka according to SINGER et al. (2008)) during MIS14, it ensures that the loess sedimentation (AS 4 and AS 5) encompasses a total of two glacial periods, MIS14 and MIS12, and that an erosional discordance in AS 4 removed the soil corresponding to MIS13. The alternative model for this profile section would be that the more strongly weathered palaeosol AS 2 represents two soil formation phases (MIS13 and MIS15) and that scarcely any sediments from the glacial period MIS14 have been pre-

served or they have been completely overprinted by the subsequent pedogenesis. This would mean that the whole loess loam of the layers AS 4a–e would have to be assigned to just one glacial period (MIS12).

In the loess loam AS 4a a palaeosol (AS 5) developed, which could be assigned to MIS11. The subsequent sedimentation phase in MIS10 encompasses the unsegmented loess loam (AS 6) at 6.6 to 6.0 m in the profile and the palaeosol (AS 7b–AS 7c), which developed from it in MIS9, in whose uppermost section at 5.4 m in the profile a magnetic excursion occurs which correlates with CR1 (Calabrian Ridge 1; 325–315 ka). These horizons can easily be distinguished from the superjacent soil horizon AS 7a due to a significant erosional discordance and a change in grain size (TERHORST 2007, TERHORST et al. 2011).

The loess sediments from MIS8 were completely overprinted during the subsequent soil formation phase in MIS7, so that only one segmented pedocomplex (AS 7a–AS 8a) has been preserved at 5.0 to 4.2 m in the profile. In the lower part of this profile section, in addition to the palaeomagnetic information on the soil formation phase during MIS7 with the

excursion J-PF (Jamaica - Pringle Falls; 215–205 ka), some remains of the remanent magnetisation of the original loess sediment have been preserved in a highly coercive magnetic phase, which also show vector directions deviating from the North direction.

This is followed above by a further loess loam (AS 9), which was deposited during MIS6 and contains a magnetic excursion at 4.0 m to 3.5 m in the profile which can be correlated with the A-FS excursion (Albuquerque - Fram Strait; 165–155 ka). The uppermost interglacial soil in the Aschet profile (AS 10) can be assigned to Eem (MIS5e) due to its palaeopedologic characteristics as well as by superimposed sediments of Würmian age. Due to datings at the Wels-Aschet site (PREUSSER & FIEBIG 2011) and in comparable profiles (TERHORST et al. 2002), an Early Würmian age can be presumed for the magnetic excursion in the colluvial layer AS 11 at 2.0 m in the profile. The formation of the colluvial layer AS 11 is linked to the beginning decline in climate during the close of Eem, which is why at this point, the colluvial layer correlates chronostratigraphically with the occurrence of the Blake excursion (120–110 ka).

The above-discussed stratigraphic findings are in contrast to the assumption of PREUSSER & FIEBIG (2009), who discuss much younger ages on the base of luminescence dating (IRSL). However, in this study, ages >200 ka do not significantly increase in the older sediments (c.f. PREUSSER & FIEBIG, 2009). Recently, methodological studies were able to extend the range of the luminescence method to 300 ka using a new post-IR IRSL protocol (290°C) (THIEL et al. 2010), an approach not yet applied to the sequence of Wels-Aschet.

## 6 Conclusion

The confirmation of Middle Pleistocene magnetic excursions accompanied by palaeopedologic-pedostratigraphic analyses are important implements for the development of a chronostratigraphic framework for the studied area as well as for the stratigraphical framework of Europe. For this reason, we executed a detailed magnetostratigraphic study of the plot of the former brickyard Würzburger in Aschet.

In general, for loess/palaeosol sequences, there is a methodological dependant wide gap in absolute age datings between the Matuyama/Brunhes boundary and MIS 6 in Pleistocene periglacial and glacial sediments (see discussion in TERHORST et al. 2011). Chinese loess sequences record five distinct palaeosols (S1 to S5) as well as three weak soils (S6 to S8) on top of the MBB and according to GENDLER et al. (2006), five to six pedocomplexes superimpose the MBB in the western Black Sea loess area. MARKOVIC et al. (2011) suppose seven interglacial soils for Northern Serbia and for the Russian plane, five to six interglacial soils are present in the Brunhes-Subchron including the transition close to the Matuyama/Brunhes boundary (VELICHKO 1990).

The loess deposits of the middle and lower Danube basin in southeastern Europe comprise the most comprehensive terrestrial records of the Pleistocene environmental dynamics (FITZSIMMONS et al. 2012). MARKOVIC et al. (2009) proposed a stratigraphic model for the Middle and Late Pleistocene based on magnetic susceptibility records from Loess sequences in the Vojvodina region (Serbia) that revealed a continuous record of paleoclimatic variations for the last

five glacial/interglacial cycles at least. BUGGLE et al. (2009) compared susceptibility records from pedocomplexes at Serbian, Romanian and Ukrainian loess/palaeosol sections with susceptibility data of the Chinese Loess Plateau and with the oxygen isotope record of ODP 677 and related the oldest pedocomplex in their study area to Marine Isotope Stage (MIS)17.

According to our interpretation, the oldest pedocomplex in the profile Wels-Aschet (AS2) might be assigned to MIS 15 and the thick package of AS4 represents very strong cooling in the Mindel glaziation with the most extensive glaciers in this area during the MIS 12. The Older Deckenschotter terrace apparently formed during the cold phase of MIS 16 (VAN HUSEN & REITNER 2011).

The correlation of magnetic parameters presented here (Figure 3) is based on the interpretation of sedimentation and soil formation phases on the basis of the palaeopedologic profile (TERHORST 2007, TERHORST et al. 2012). The magnetic parameters show a series of sections with intense magnetite formation in the soil, which can be assigned to the relatively warm climate of interglacial periods. In some sections of the profile the trend of the magnetic susceptibility shows a notable consistence with the oxygen isotope curve. Indications of short-term magnetic excursions can be observed in the section of the palaeosols as well as in the loess sediments. Due to sedimentologic considerations, the excursions observed in the Wels-Aschet profile are equated with the excursions Blake (120–110 ka), A-FS (Albuquerque - Fram Strait; 165–155 ka), J-PF (Jamaica - Pringle Falls; 215–205 ka), CR1 (Calabrian Ridge 1; 325–315 ka) and E-BL-CR3 (Emperor - Big Lost - Calabrian Ridge 3; 570–560 ka) of the geomagnetic reference time scales. For the excursion in the loess complex (AS4e), depending on which reference time scale is used, either CR2 (Calabrian Ridge 2; 525–515 ka), West Eifel 5 (528+-16ka) or a much younger age are up for discussion. The Brunhes-Matuyama Boundary (776 ka) was never reached.

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# Paleopedological record along the loess-paleosol sequence in Oberlaab, Austria

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**Abstract:** A detailed study of a loess-paleosol sequence in Oberlaab, Upper Austria, is presented with emphasis on macro- and micromorphological features, grain size distribution, rock magnetism properties, and weathering degree that allows correlation with other loess-paleosol sequences in neighboring areas, and interpretation of main pedogenic trends. The studied sequence comprises four paleosol complexes, which likely developed during four interglacial stages MIS 11, 9, 7 and 5e, and a modern soil. The oldest paleosol complex (OL5) represents three phases of soil formation, and distinct sedimentary events never reported in the area, with strong reductomorphic properties. The OL4 profile also results from three phases of pedogenesis with increased reductomorphic features in the deepest zone (affected by cryoturbation events). OL3 has abundant features related to gleyic/stagnic processes, but shows signs of clay illuviation. OL2 (Eemian soil) correlates with the MIS 5e. This paleosol shows higher degrees of clay illuviation and weathering, and fewer features related to reductomorphic processes. The modern soil is also polygenetic and constitutes a pedocomplex. Its lowermost part is formed by Würmian glacial deposits, where no well-developed soils are found; only reworked materials and pedosediments. Main pedogenic trends in the sequence are clearly differentiated. All of the paleosols were formed in humid environments, but differing in drainage conditions. The base, with OL5 and OL4 paleosols, was more affected by gleyic processes, while in the upper paleosols, especially OL2, clay illuviation is dominant. We interpret such differences to be caused by the topographic position. The basal paleosols were more affected by fluvio-glacial processes due to their position on top of the terrace. The upper paleosols received increased amounts of sediment through fluvial, colluvial and aeolian (loess) input.

## Das paläopedologische Archiv der Löss-Paläoboden-Sequenz in Oberlaab, Österreich

**Kurzfassung:** Eine detaillierte Untersuchung einer Löss-Paläoboden-Sequenz in Oberlaab, Oberösterreich, wird vorgestellt, deren Schwerpunkt auf makro- und mikromorphologischen Merkmalen, der Korngrößenverteilung, magnetischen Gesteinseigenschaften und dem Verwitterungsgrad liegt. Die untersuchten Aspekte ermöglichen einerseits eine Korrelation mit anderen Löss-Paläoboden-Sequenzen in benachbarten Gebieten und andererseits die Interpretation wichtiger pedogenetischer Entwicklungen. Die untersuchte Sequenz besteht aus vier Paläoboden-Komplexen, deren Entwicklung sehr wahrscheinlich während der vier Interglaziale MIS 11, 9, 7 und 5e stattfand, sowie einem rezenten Boden. Der ältere Paläobodenkomplex (OL5) beinhaltet drei Phasen der Bodenbildung mit unterschiedlichen Sedimentationsereignissen, die in diesem Gebiet bislang noch nicht beschrieben wurden. Darüber hinaus finden sich stark redoximorphe Merkmale. Das Profil OL4 zeigt ebenfalls drei Bodenbildungsphasen mit zunehmend redoximorpher Prägung im unteren Bereich, wobei eine Überprägung durch kryptoturbate Prozesse sichtbar ist. Auch an OL3 konnten Merkmale grund- bzw. stauwasserbeeinflusster Prozesse dokumentiert werden, es finden sich aber auch Hinweise auf Tonverlagerung. Der Eemboden OL2 entspricht dem MIS 5e. Hier zeigen sich die intensivste Tonverlagerung und der stärkste Verwitterungsgrad, sowie ein Rückgang der redoximorphen Merkmale. Auch der rezente Boden ist polygenetisch und stellt sich als Pedokomplex dar. Der unterste Bereich besteht aus würmzeitlichen glazialen Ablagerungen. In diesem Profilteil finden sich keine gut entwickelten Böden, sondern ausschließlich umgelagertes Material und Pedosedimente.

Die vorherrschenden pedogenetischen Prozesse in der Sequenz können klar abgegrenzt werden. Alle Paläoböden entstanden unter humiden Bedingungen, wobei aber jeweils eine unterschiedliche Drainage vorhanden war. Die Profilbasis, die durch die Paläoböden OL5 und OL4 gebildet wird, ist stärker von Grundwasser beeinflusst, während in den oberen Paläoböden, v.a. in OL2, die Tonverlagerung dominiert. Es ist anzunehmen, dass diese Unterschiede in der topographischen Position begründet sind. Die unteren Paläoböden sind aufgrund der tieferen Lage der Terrasse stärker von glazifluvialen Prozessen betroffen, während die oberen Paläoböden nach der Sedimentation von großen Materialmengen (fluvial, kolluvial und/oder Löss) eine höhere Position einnehmen.

**Keywords:** *Middle Pleistocene, loess, paleosol, Oberlaab, pedogenesis*

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## 1 Introduction

Loess/paleosol sequences are a valuable source of information about past climatic and environmental conditions during the Quaternary, and are one of the most complete terrestrial records of glacial-interglacial cycles (PÉCSI 1990; PÉCSI & SCHWEITZER 1993; MUHS & BETTIS 2003) successfully correlated to the global climate proxies as marine isotope curves (KUKLA 1978; BRONGER et al. 1998). In temperate zones they have provided reliable information about paleoenvironmental change, particularly for the last glacial/interglacial cycle, as in Eurasia (LIU 1985; ROZYCKI 1991; SUN et al. 1997; MUHS & BETTIS 2003; FENG et al. 2007), United States (MUHS & BETTIS 2000; BETTIS et al. 2003), and South America (MUHS & ZÁRATE 2001; QUATTROCCHIO et al. 2008). In Austria, several works contain detailed descriptions for loess-paleosol sequences (e.g. FINK 1976; KOHL 1976; HAESAERTS et al. 1996; NEUGEBAUER-MARESCH 1996; NIEDERHUBER 1997; TERHORST 2007). However, our understanding of the pedostratigraphy and pedogenesis of loessic paleosols remains incomplete. In particular, in Upper Austria, where over a century of research on Pleistocene landscape history has been conducted since the classical works of PENCK & BRÜCKNER (1909), loess-paleosol records are still yielding new data for regional paleoenvironmental reconstructions.

We studied a loess-paleosol sequence in Upper Austria, situated on the top of a Mindel fluvioglacial terrace. The Austrian stratigraphic table classifies the classical Mindel glacial as belonging to the MIS 12 (ÖSTERREICHISCHE STRATIGRAPHISCHE KOMMISSION 2004). VAN HUSEN (2000) and VAN HUSEN & REITNER (2011) correlate Mindel glacial sediments in the Eastern Alpine Foreland with the MIS 12, while the Rissian deposits are associated to the MIS 6. On the basis of pedostratigraphical analyses in the Linz/Wels area, TERHORST (2007; 2013, this volume) and TERHORST et al. (2011) propose a similar correlation for the loess/paleosol sequence of Oberlaab, which includes four interglacial paleosols located on top of the Mindel terrace. These paleosols are linked to MIS 11, 9, 7 and 5, respectively (TERHORST et al. 2011, 2013), and span parts of the Middle to the Upper Pleistocene. According to TERHORST (2007) and TERHORST et al. (2011, 2013) Middle Pleistocene sequences in Upper Austria are the result of an alternation of geomorphogenetic-pedogenic processes. During Middle Pleistocene times, geomorphodynamics in the region were controlled by unstable glacial periods (causing phases of erosion and sedimentation), which alternated with stable interglacial periods and soil formation. Using the concept of “soil memory” (in the sense of TARGULIAN & GORYACHKIN 2004) the resulting paleosols contain information about environmental conditions of the interglacials.

In Oberlaab, we conducted a paleopedological survey, which involved a multidisciplinary approach to identify types of pedogenesis and degrees of paleosol development. Previously, TERHORST (2007) studied a similar sequence in the loam pit of Oberlaab, which comprises thick pedocomplexes, related to four interglacial periods, covering a time span of >400 ka. In this paper, we contribute a detailed morphological description of pedostratigraphic units accompanied by a set of quantitative pedogenetic characteristics:

morphometric analysis of diagnostic pedofeatures, grain size distribution, magnetic parameters (magnetic susceptibility and frequency dependence of susceptibility), as well as geochemical indicators of weathering. These characteristics provide a base for reliable comparison between buried paleosols and the Holocene surface soil, in order to establish differences in the typology and grade of development of the main pedogenic processes. The resulting paleopedological record contributes to the reconstruction of the environmental conditions of the four last interglacials and provides complementary data for the regional geomorphological and stratigraphic schemes.

## 2 Materials and methods

The study site is situated in a quarry exploited for clay for brick production, 1.5 km to the north of Wels, Austria (for details see TERHORST 2013, this volume). The general stratigraphy was presented by TERHORST (2007), who has recognized four interglacial paleosols. The loess/paleosol sequence of Oberlaab was developed on a Mindel terrace (Younger Deckenschotter). The basement of the sequence is formed by gravels, exposed previously in a cut operation during 2003. The whole sequence has a thickness of 15.80 m, which has been sectioned and labeled as the units OL1, OL2, OL3, OL4 and OL5. Each of these sections contains a soil/paleosol/pedocomplex unit (Figure 1).

Genetic horizons of modern soil and paleosols were defined according to their morphology following the World Reference Base (IUSS WORKING GROUP WRB 2007). We observed ped features under a 10X lens in the field. In all cases, interpretation was based on comparison of paleosols with the modern soil.

### 2.1 Micromorphology

For micromorphological studies we took undisturbed samples from every genetic horizon (Figure 1). Thin sections were prepared and described with the guide of BULLOCK et al. (1985).

To quantify the proportion of pedofeatures in different horizons, selected thin sections were scanned with high resolution (4800 and 9600 dots per inch); the digital image was later analyzed with the Image Pro Plus v.6.0 software. A contrast enhancement of every image was produced to distinguish zones with diffuse contacts to the groundmass, using the blue RGB channel acquisition. The false color was used for the segmentation of areas occupied by each pedofeature with a previous conversion to shades of grey. The most important features analyzed were: 1) iron nodules and impregnations, 2) bleaching areas and 3) clay coatings. However, in most of the thin sections the complex mixture of iron and clay was very difficult to separate by the software algorithms. Thus, we decided to study clay coatings semi-quantitatively using visual schematic petrographic diagrams (CASTRO 1989). Porosity was not estimated in the total average, because in some thin sections voids had the same color as felsic minerals.

Besides micromorphometric analysis of selected pedofeatures we established the assemblage of the primary minerals, their distribution in different size fractions and their weath-

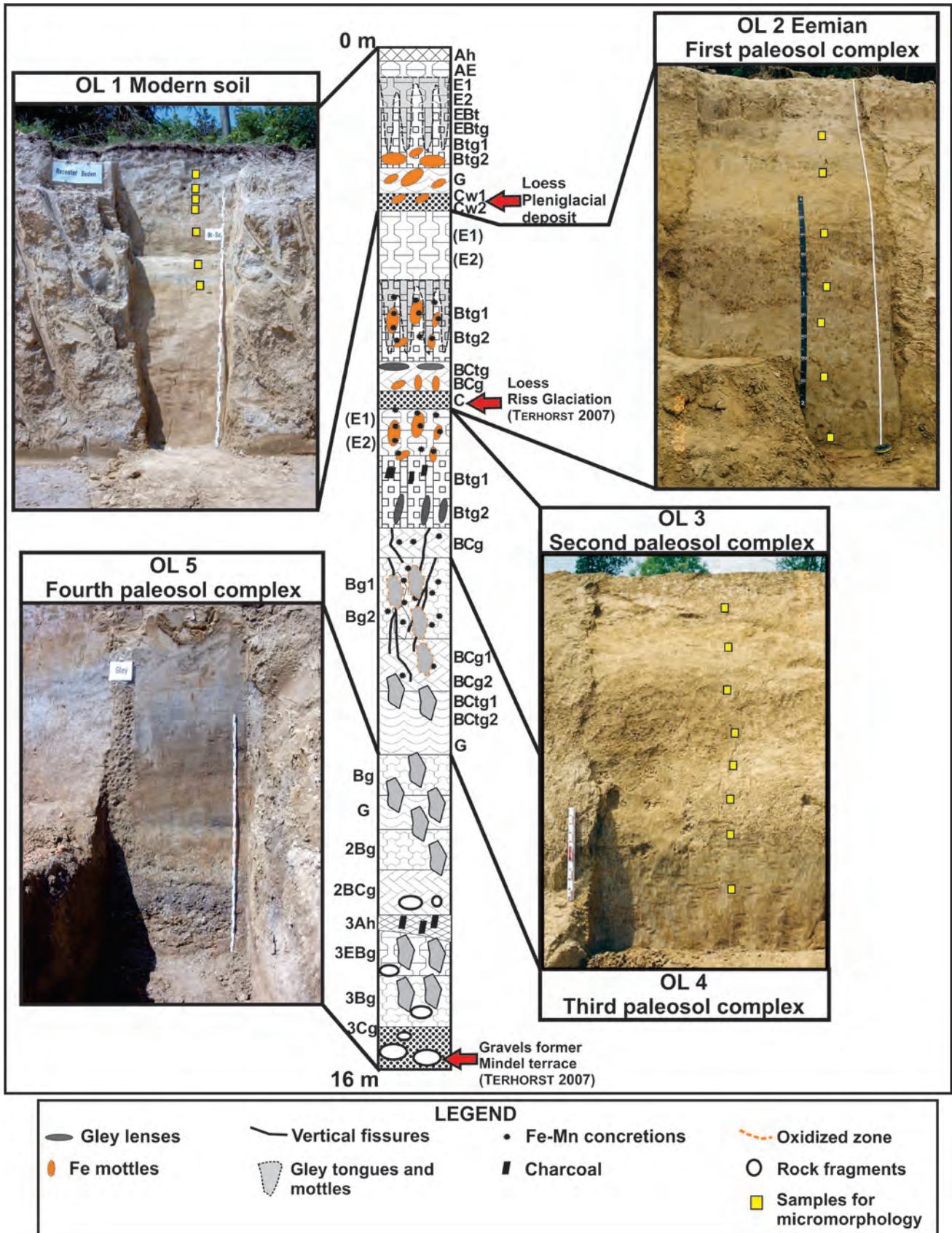


Fig. 1: Oberlaab loess/paleosol sequence on top of a Mindel terrace (Young Deckenschoter), which includes four interglacial paleosols and the modern soil.

Abb. 1: Stratigraphie der Löss-Paläoboden-Sequenz Oberlaab, sowie ein allgemeiner Überblick über alle vier Paläoböden und den rezenten Boden.



ering status basing on the crystalloptical properties under magnification with a petrographic microscope.

## 2.2 Laboratory analyses

Size fractions were separated in OL1, OL2, OL4 and OL5. OL3 was not evaluated due to the lack of samples. The sand fraction (2–0.063 mm) was separated by sieving; silt (0.063–0.002 mm) and clay (<0.002 mm) fractions by gravity sedimentation with previous destruction of aggregating agents; H<sub>2</sub>O<sub>2</sub> (15%) was used for organic matter, and dithionite-citrate-bicarbonate (DCB) for iron oxides. As the deposits had no carbonates, no additional pre-treatments were necessary for their destruction.

For magnetic measurements, 200 g samples were collected at roughly 20 cm intervals, homogenized, and placed in 8 cm<sup>3</sup> acrylic boxes. Magnetic susceptibility ( $\chi$ ), which is a measure of the concentration of magnetic minerals, was registered in all samples at low (0.47 kHz) and high frequencies (4.7 kHz) with a Bartington MS2B dual sensor. We calculated frequency dependence of susceptibility  $\chi_{fd}\%$  as  $[(\chi_{lf}-\chi_{hf})/\chi_{lf}]100$ , to approximate possible ultrafine (< 0.05  $\mu\text{m}$ ) superparamagnetic (SP) contribution. DEARING (1994) and DEARING et al. (1997) suggest that values of  $\chi_{fd}\% < 2\%$  point to a content of <10% of SP (superparamagnetic) grains; values around 8% indicate a SP contribution of 75%, and between 10–14%, the material is domain by SP particles.

## 2.3 Geochemical indices of weathering

Bulk chemical composition was obtained by X-ray fluorescence in a Siemens SRS 3000 spectrometer, using powder pellets at the Institute of Geology, UNAM. All results were calculated by weight on oven dry (105° C) soil.

As the parent material of the section is mainly loess rich in silica, we decide to evaluate the degree of weathering using the chemical index of alteration (CIA). CIA measures changes in major cation content in relation to immobile alumina, instead of changes in the silica/alumina composition. This proxy was first proposed by NESBITT & YOUNG (1982) and introduced in loess studies by LIU et al. (1995). CIA is evaluated following the formula:  $\text{CIA} = (\text{Al}^{2}\text{O}_3 / (\text{Na}_2\text{O} + \text{CaO} + \text{MgO} + \text{K}_2\text{O})) * 100$ . The reason to use this weathering index is the high concentration of K-feldspar found in the mineralogical assemblage, according to BUGGLE et al. (2011) and the lack of carbonates (thus the CaO content is only associated to silicates). A value near 100 represents the highest degree of weathering, while values lower than 50 correspond to fresh sediments.

## 3 Results

### 3.1 Morphology of the Oberlaab paleosol sequence

The study sequence in Oberlaab includes a modern soil (OL1) and 4 paleosol units (OL2 to OL5) from the top to the bottom (Figure 1).

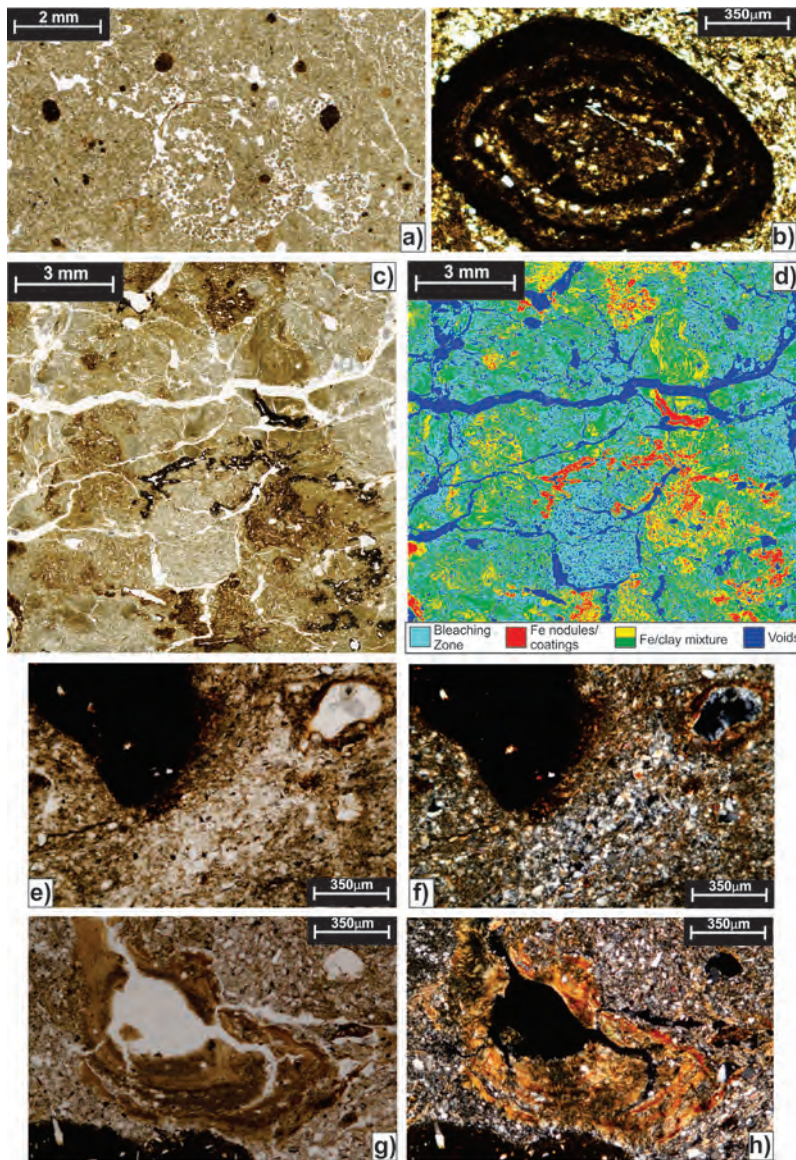
OL1. Modern soil (depth 0–190 cm). The OL1 soil shows a well-developed profile with Ah-AE-E1-E2-EBt-EBtg-Btg1-Btg2-G-Cw horizons. The Ah horizon is thin (5 cm-thick), dark brown, very porous, with a consistent granular

structure and a high density of roots. The transitional AE horizon is also thin (10 cm thick) and paler than Ah. The underlying E horizon is 30 cm thick, divided into E1 and E2. This division was made because E2 is notably darker with less silt. In general, the E horizon is light gray, silty and with subangular blocky structure. The contact with the underlying Btg1 is also transitional, through the EBt and EBtg horizons (45–100 cm). Light gray tongues of compacted, silty material, starting from the E horizon, enter into Btg (100–150 cm). These tongues are separated from each other by 15 cm in some cases, and 25 cm in others. Roots do not penetrate the tongues. The Btg1 horizon is grayish-yellowish brown, with yellowish brown mottles. Clay coatings are thick and frequent. The Btg2 horizon has a similar color, but is distinguished by a network of cracks. The cracks boundaries are gray, contrasting with the brownish color of the matrix. The lowermost G horizon (150–160 cm) is gray in color, more dense and has a silty texture. The subangular blocky peds exhibit a platy structure. The Cw horizon (160–190 cm) corresponds with the uppermost part of a pale-yellow, silty, loess-like deposit, showing horizontal gray bands.

The underlying Cw horizon corresponds to a loess-like deposit (190–390 cm). This material constitutes a pedosediment, which contains a mixture of sediments and reworked soils.

OL2. Eemian soil (depth 390–670 cm). The next stratigraphic level corresponds to the Eemian soil (Figure 1), and contains a sequence of (E1)-(E2)-Btg1-Btg2-BCtg-BCg-C horizons. Vertical cracks cross the first 200 cm (390–590 cm), filled by a gray material, of which the inner part is black. The boundary between the gray and black materials is reddish.

Because of the paler color (brownish yellow) of the first 60 cm (390–450 cm), we initially assumed that the material corresponds to an E horizon, which is divided into E1 and E2 with a silty loam texture. However, a more detailed morphological study revealed that the horizon is constituted by two layers of reworked pedosediments. The loess-like sediment is affected by weak pedogenesis, with rounded sediment grains and soil fragments. The soil fragments occur as small subangular blocks breaking into granular aggregates, with a tendency to form a weak platy structure. Moderate amounts of 2–3 mm-diameter Fe/Mn-concretions were detected, varying from dark brown to red. The transitional limit to the underlying Btg horizon (Btg1 and Btg2) is wavy. The Btg horizon (450–540 cm) is yellowish brown, darker than the reworked pedosediments of the upper part, more compact, and clayey. Btg1 has a better developed platy structure, which breaks to subangular blocks. Clay coatings are abundant. In the upper surface of the platy aggregates, discontinuous dark brown clay coatings are present. In this horizon the vertical cracks are covered by clay and silt coatings (the latter are observable with a 10X-hand lens). Fe/Mn-concretions are abundant. The horizon shows strong mottling, which consist of bright yellow areas with brownish and grayish zones. These spots are mainly oriented along the vertical cracks. In Btg2, clay coatings are fewer and more disperse. The BCtg horizon (540–570 cm) is paler and also compact. Here, fewer clay coatings and Fe-Mn concretions are detectable. Vertical



**Fig. 2: Micromorphology of the study section:**  
 a. Coalescent excremental infillings in chambers in AE horizon of the modern soil. High resolution image (4600 dpi) of scanned thin section.  
 b. Concentric iron nodule in G horizon of the modern soil, photomicrograph, plane polarized light (PPL).  
 c. Bleached zones in the groundmass of BCg1 horizon in OL4. High resolution image (4600 dpi) of a scanned thin section.  
 d. Same as 2c: Image analysis in false color.  
 e. Bleached zone (center) with sharp contact with the groundmass in OL3-BCg horizon. Fe nodules showing abrupt boundaries with the groundmass, photomicrograph (PPL).  
 f. Same as 2e, cross polarized light (XPL).  
 g. Clay coatings around a void associated with a bleached zone in OL3-Btg1 horizon. The dark-yellowish color of the clay is related to the presence of iron and coarse silt fraction, photomicrograph (PPL).  
 h. Same as g, XPL. Note the high interference color related to the presence of 2:1 clay structure.

**Abb. 2: Mikromorphologie des untersuchten Abschnitts:**  
 a. Koagulierte Exkrementfüllungen in Hohlräumen im AE-Horizont des rezenten Bodens. Hochauflösende Darstellung (4600 dpi) von gescanntem Dünnschliff.  
 b. Konzentrische Eisenkonkretion im G-Horizont des rezenten Bodens. Mikroskopaufnahme unter eben polarisiertem Licht (plane polarized light, PPL).  
 c. Gebleichte Zonierung in der Matrix des BCg1-Horizonts von OL4. Hochauflösende Darstellung (4600 dpi) von gescanntem Dünnschliff.  
 d. Wie 2c: Falschfarbenbildanalyse.  
 e. Gebleichter Bereich (Zentrum) in scharfer Abgrenzung zur Matrix im BCg-Horizont von OL3. Mikroskopaufnahme (PPL). f. Wie 2e., gekreuzte Polarisatoren (cross polarized light, XPL).  
 g. Kugelförmige Struktur (roter Rahmen rechts) in OL4. Hochauflösende Darstellung (4600 dpi) von gescanntem Dünnschliff.  
 h. Tonüberzüge in einem Hohlraum in Verbindung mit einer Bleichzone in Btg1 in OL3. Die dunkelgelbe Farbe des Tons ist bedingt durch den Eisengehalt und der Korngrößenfraktion Grobschluff. Mikroskopaufnahme (PPL).

oriented channels are also present. They are filled by laminated clay (dark brown) and silt coatings (yellow and gray). The BCg horizon (570–635 cm) is similar, but here coatings are very rare. The most conspicuous feature is the change of color as well as the intensity of gleyic characteristics throughout the horizon. Grey, horizontally oriented lenses appear in the first 15 cm. The matrix of the next 30 cm is grayish-yellowish brown with some yellow spots. The lower part is light yellowish brown and has diffuse mottles. The C horizon (635–670 cm) has a silty loam texture and a brown to yellowish brown color with rusty mottles. There are no Fe concretions. Structure is weak varying from subangular blocky to massive.

OL3. Second paleosol complex (depth 670–900 cm). Similar to the Eemian soil in OL2 (Figure 1) this unit has two levels of reworked pedosediments (E1 and E2) on the top (670–730 cm), which are superimposed on the following compact horizons: Btg1-Btg2-BCg. The first pedosediment level (30 cm thick) is strongly mottled. Matrix has a light brownish yellow color with gray stagnic mottles and round rusty patches (1 to 1.4 cm in diameter). Structure shows large subangular blocks and inside the block aggregates

there is a tendency to form a thin platy structure. Vertical cracks are frequent and filled by clay pellets. There are bleached areas with light colors that contrast with darker stagnic zones. The second level of reworked pedosediments (30 cm thick) corresponds to a strongly patchy silty loam. In this case the color is paler because of dominant grayish mottles. Rounded mottles are covered by dark brown clay coatings. Btg1 (730–790 cm) is characterized by a dark brown matrix with even darker mottles that alternate with brownish gray spots. Structure is very well developed in strong angular blocks, which break into smaller angular aggregates. Dark brown Fe/Mn-concretions with a size from 1 to 5 mm diameter are frequent. Dark brown clay coatings are present; however, they are very rare and appear in few peds. Small quantities of charcoal are present and associated to thin root channels. Due to an increase of vertically oriented gley mottles, Btg2 (790–860 cm) has a clear change in color as well as in structure (subangular blocks). In this horizon we detect a decrease in the abundance of clay coatings, and in the quantity and size of the Fe-concretions. Some rounded rock fragments are found. The BCg horizon (860–900 cm) is strongly patchy. Matrix is

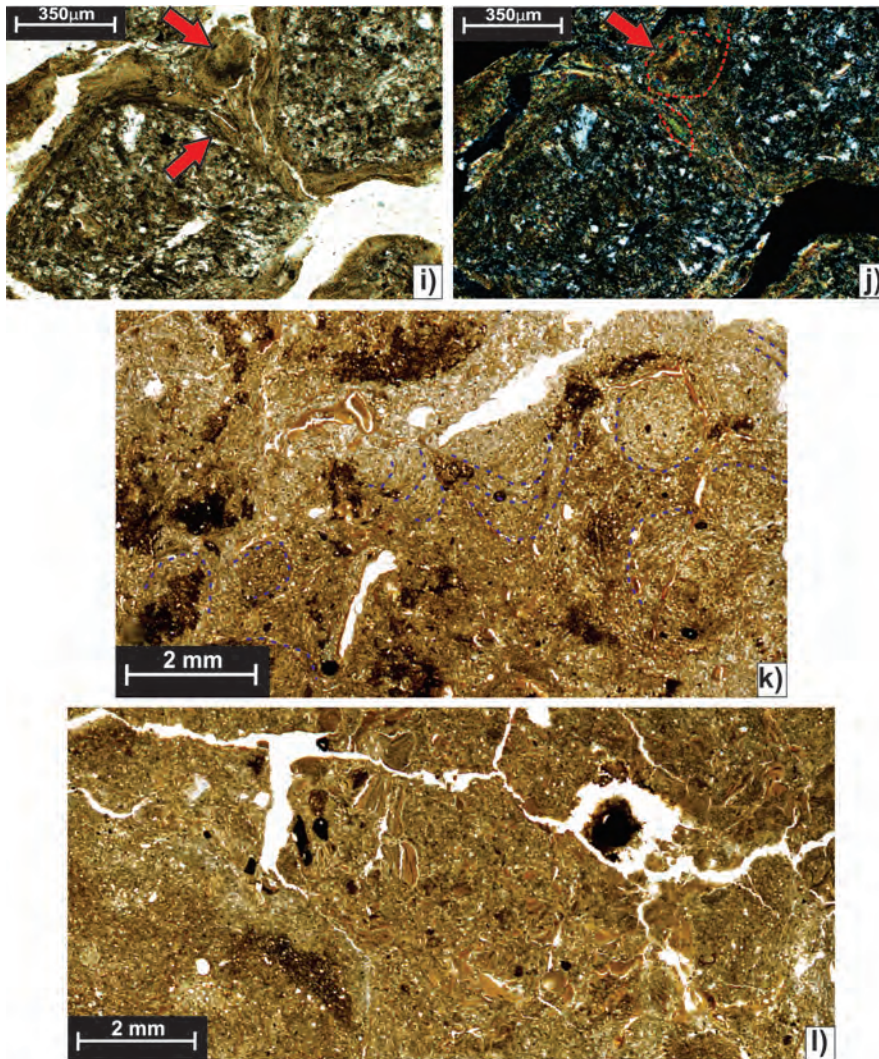


Fig. 2: Micromorphology of the study section: i. Deformed pedofeatures. Deformed clay coatings showing a "twisted" morphology and displacing the groundmass (red arrow) around blocky microstructure in Btg2 horizon of OL3, photomicrograph (PPL).

j. same as 2j, (XPL). The red discontinuous lines show the "twisted" morphology with the displacing of the groundmass. Note the high interference color inside and outside of the blocky microstructure, related to a primary phase of clay illuviation (2:1 clay structure). k. Orbicular fabric (blue discontinuous line) in Bg2 horizon in OL4. High resolution image (4600 dpi) of scanned thin section. l. Fragmented clay cutans possibly associated to bioturbation in Bg2 horizon of OB4. High resolution image (4600 dpi) of scanned thin section.

Abb. 2: Mikromorphologie des untersuchten Abschnitts:

i. Wie h., gekreuzte Polarisatoren. Man beachte die hohen Interferenzfarben aufgrund der 2:1-Struktur der Tonminerale.

j. Fragmente illuvialer Tonüberzüge verdrängen die Matrix im Btg2-Horizont von OL3. Mikroskopaufnahme, (PPL).

k. Wie 2j., gekreuzte Polarisatoren. Auch hier ist die hohe Interferenzfarbe innerhalb des primär illuvial verlagerten Fragments zu beachten. Der rote Pfeil kennzeichnet teilweise in der Matrix assimilierte Toncutane.

l. In der Bildmitte Fragmentierung von Toncutanen in Bg2 (OB4), möglicherweise durch Bioturbation. Hochauflösende Darstellung (4600 dpi) von gescanntem Dünnschliff.

yellowish brown with mottles varying from gray to brownish gray and red. The structure is subangular blocky and platy. Vertical cracks filled by grayish material cross the horizon. Fe/Mn-concretions are frequent (0.1 to 0.5 in diameter).

OL4. Third paleosol complex (depth 900–1200 cm). This unit is constituted by Bg1-Bg2-BCg1-BCg2-BCTg1- BCTg2-G horizons (Figure 1), which are very compact and have silty loam textures. The unit is crossed by vertical fissures filled by dark gray, stagnic materials. In the central part of the fissures we observe clay and silt coatings (the latest distinguish by their coarser texture). The Bg1 horizon is strongly patchy and wavy (900–950 cm). Dark brown material alternates with light grey vertically oriented mottles. The exterior part of the mottles is surrounded by yellow-brown oxidized zones. Structure is platy and angular blocky. Abundant Fe/Mn- concretions (0.5–1cm in diameter) are found. Dark coatings are present, especially on the surface of platy blocks. Bg2 (950–990 cm) clearly differs from the previous horizon by the increase of gray areas (stagnic) on the ped surfaces. Fe/Mn-concretions are also present but in lesser amounts. BCg1 (990–1030 cm) is patchy, paler than the previous Btg, but with a similar structure. In this horizon the vertical fissures are thicker than in the upper part. Fe and Mn coatings are frequent. BCg2 (1030–1100 cm) is very similar to BCg1 but less patchy. BCTg1 (1100–1140

cm) and BCTg2 (1140–1180 cm) are alike, but BCTg1 is more clayey and shows dark clay coatings. Round vertical channels are common. G (1180–1200 cm) is more homogeneous in color with a grey matrix, showing small brown rusty mottles. Structure is well developed and consists of small subangular blocks. It is clear that vertical channels do not enter into this horizon.

OL5. Fourth paleosol sequence (1200–1520). This lowest stratigraphic level is clearly a pedocomplex, composed by Bg-G-2Bg-2BCg-3Ah-3EBg-3Bg-3Cg horizons (Figure 1). The upper Bg (1200–1250 cm) is pale brownish yellow with rusty and grey mottles. It is constituted by very fragile subangular blocks with a silty clay texture. Its transition to the G horizon is wavy and gradual. The G horizon (1250–1290 cm) is bluish gray with black and rusty mottles. Due to the occurrence of small subangular blocks the structure is better developed than in the Bg horizon. Mn segregations are common. The boundary with the next 2Bg horizon (1290–1370 cm) is clear and wavy. It is pale brown with brown, rusty and black mottles. It has a silty loam texture. Friable subangular blocky structure is dominant with a tendency to break into crumbly, granules. Mn is segregated in pores and channels. Very few clay coatings are present in channels and on ped surfaces. The boundary with the less structured 2BCg (1370–1415 cm) is wavy and gradual. Rounded rock fragments are present in this

horizon marking a clear and straight limit to the next 3Ah horizon (1415–1430 cm), which has a very dark gray color with rusty and brown mottles. It is a silty clay loam with a very fragile subangular blocky-granular structure. Charcoal fragments and few clay coatings are found in channels and pores. The transition to the 3EBg horizon is clear and slightly wavy. 3EBg (1430–1450 cm) is pale gray. Rusty, brown, and black mottles are present. Its structure is constituted by fragile, small subangular blocks and gravels are frequent. 3Bg1 (1450–1490 cm) is yellowish brown with grayish mottles. It is a silty clay with a fragile structure composed of subangular blocks. Mn segregations and ferruginous hypocoatings are abundant in pores and channels. The lower part of this horizon is more grayish. Its contact with the underlying 3Cg (1490–1520 cm) is abrupt. It is build up by gravels with abundant Mn segregations and dark-brown mottles around the gravels, which increase in size with depth.

### 3.2 Micromorphology

The main micromorphological features observed in the Oberlaab section can be divided into four kinds: biogenic components and features, redoximorphic pedofeatures, clay coatings, and deformed pedofeatures. In most cases the proportions are obtained by using the scanned thin sections and the Image Pro software.

Biogenic components and features. These materials include root traces, humus, coprolites, channels, and charcoal. They are clearly recognizable in both the modern soil and the OL4 profiles. The Ah and AE horizons of the modern soil (OL1) are characterized by strong bioturbation including the presence of chambers with loose and discontinuous coalescent excremental infillings (Figure 2a) associated with root traces (in some cases more or less degraded). Dark humus impregnates the matrix, in cases where a granular structure dominates. Biogenic materials in OL4 are less diverse. They are presented only by charcoal fragments with a well preserved cellular structure, as well as passage features of mesofauna.

Redoximorphic pedofeatures. These kinds of pedofeatures include ferruginous concretions, hypocoatings, and bleached areas. Paleosols contain the highest concentrations in comparison to the modern soil (Figure 3). However, Fe concretions are frequent in the G horizons of the modern soil, where they show sharp boundaries to the matrix (Figure 2b). Fe concretions and hypocoatings reach the maximum (19%) in the BCg1 horizon of OL4. The highest proportion of bleached areas (65%), which permit the estimation of zones where iron has been removed, was found in the Bg2 horizon of OL4 (Figures 2c and 2d, Figure 3). It is interesting that the contact between bleached zones and groundmass is diffuse in surficial horizons, however, in the lower part of the profiles, boundaries are more abrupt. Figures 2e and 2f clearly

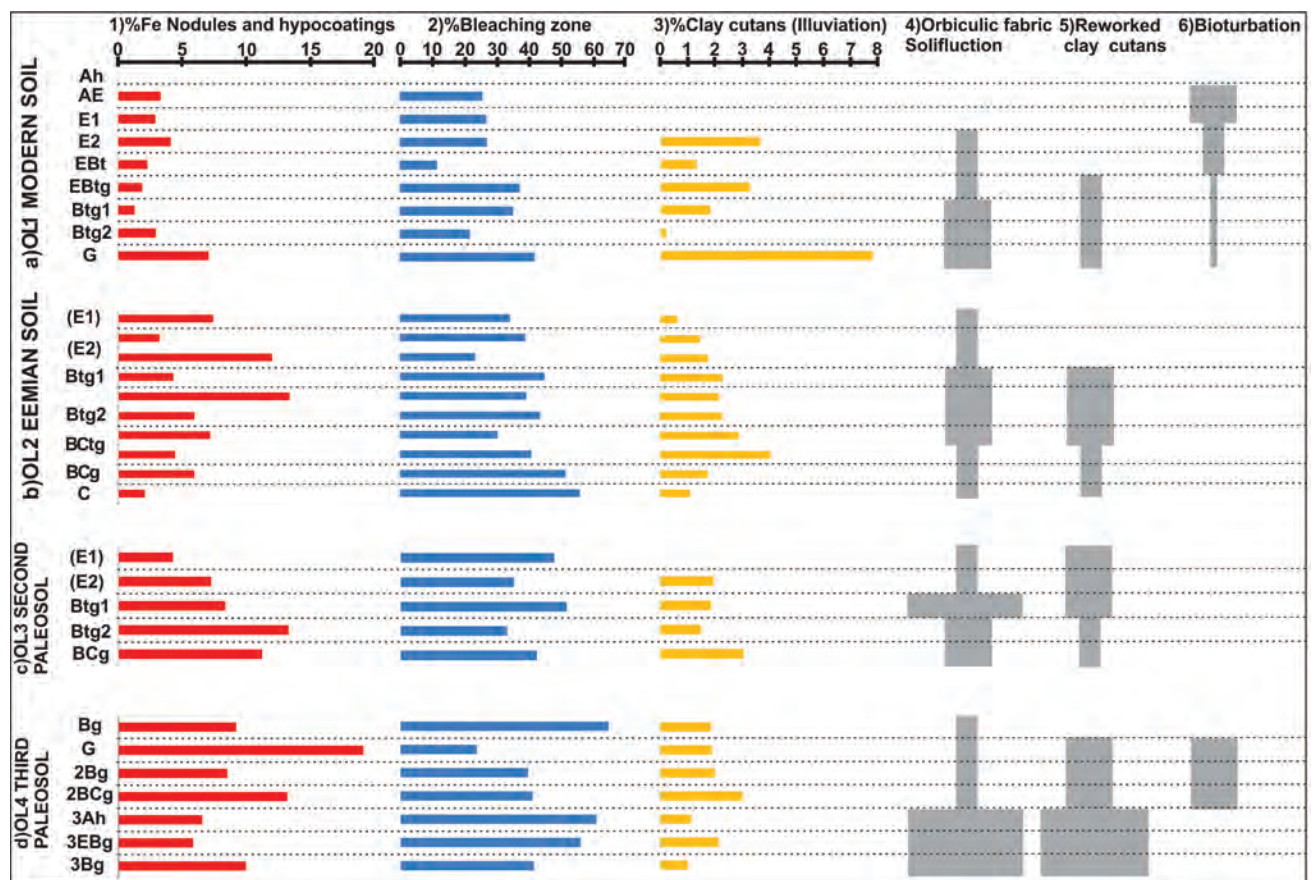


Fig. 3: Proportion of micropedofeatures in the Oberlaab sequence, according to the image analysis using the software Image Pro in high resolution image (4600 dpi) of scanned thin section.

Abb. 3: Quantifizierung von Mikrobodenmerkmalen, basierend auf der Bildanalyse mit der Software Image Pro, in hochauflösender Darstellung (4600 dpi) von gescanntem Dünnschliff.

show this kind of contact in the BCg horizon of OL3.

**Clay coatings.** Clay coatings are common in the modern soil; coating content varies from 1 to 3.5%. Surprisingly, the G horizon reaches a 7% the maximum in the whole sequence (Figure 3). In the Eemian paleosol (OL2), particularly in the Btg and BCtg horizons, the amounts are also high (2.1 to 4%). The lowest concentration of clay coatings is present in the OL3 and OL4 profiles (Figure 3). The “impurity” of the clay coatings varies in the profiles. In the modern soil most of the coatings are pure and limpid (with yellowish colors), meanwhile in the paleosols coatings show Fe microlaminations and pigmentations (Figure 2g and 2h).

**Deformed pedofeatures.** All studied paleosols, as well as the modern soil show deformation of some pedofeatures. This deformation is more evident in Btg horizons, where

the clay coatings are assimilated inside the bleached, silty groundmass and exhibit a “twisted” morphology (Figure 2i and 2j). The presence of fragmented and displaced groundmass blocks with “cloud” structures as well as orbiculate fabrics are also common (Figure 2k). Additionally, in OL3 and OL4 small blocks are fragmented and displaced. In such blocks we observe two different generations of clay coatings: the first one (thin and discontinuous clay coatings) is inside the blocks; the second one (with thicker coatings) covers the aggregates. Another kind of “deformed” coatings is present in BCg horizons (particularly in BCg horizon of OL4). These coatings are fragmented and rounded, indicating reworking processes (Figure 2l).

Besides the micromorphological descriptions made for the thin sections we analyzed the bulk mineral com-

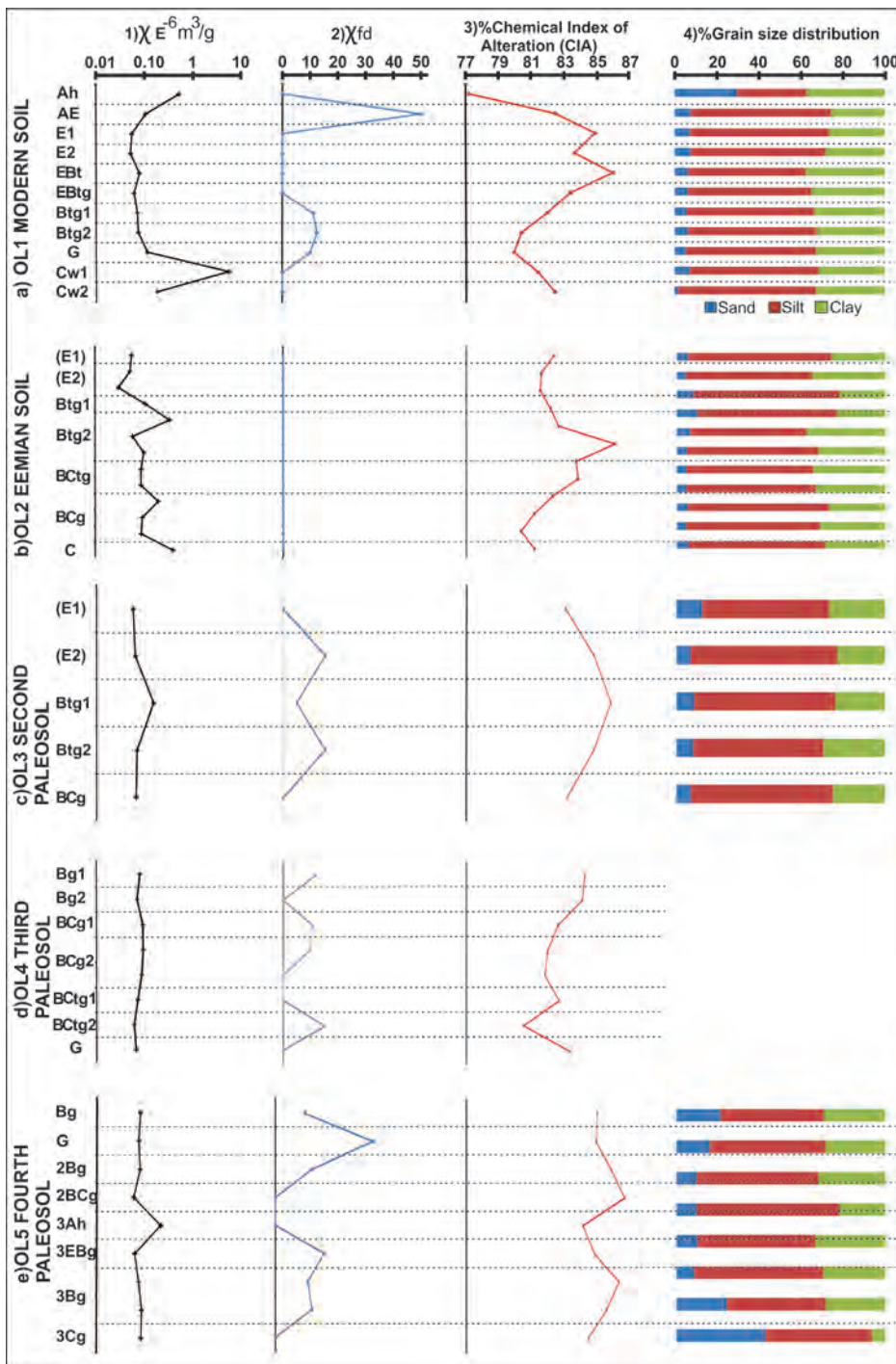


Fig. 4: Selected analytical properties of the study sequence: Magnetic susceptibility ( $\chi E^{-6} m^3/g$ ); frequency dependence of susceptibility ( $\chi fd\%$ ); grain size distribution (%).

Abb. 4: Ausgewählte Analyseergebnisse des untersuchten Abschnitts.

position of sand and silt fractions. The mineral content is mostly constituted by quartz, potassium feldspar, and mica (muscovite), which are moderately to well sorted. Biotite is present in low quantities and mostly concentrated in the sand fraction. Heavy minerals such as amphiboles (in some cases with “serrated” morphology), epidote, and garnet are concentrated in the silt fraction and demonstrate angular shape. Some schist fragments are recognizable in primary sedimentary layers. This mineralogical assemblage coincides well with the reports from other loess-paleosols sequences of Upper Austria (TERHORST et al. 2011).

### 3.3 Analytical properties

The silt fraction in the modern soil ranges from 33 to 67%. The surficial Ah horizon shows a high amount of sand (30%), contrasting to that obtained in the rest of the profile (Figure 4). The clay fraction exhibits similar proportions in all horizons, however, it increases in EBT and Btg horizons. The Cw horizon contains the highest proportion of silt (64–66%). Magnetic susceptibility ( $\chi$ ) is generally low in all horizons, but with high values in Cw1 and Ah horizons. The low values of the frequency dependence of susceptibility  $\chi_{fd}\%$  indicate a null contribution of super paramagnetic (SP), ultrafine grains ( $<0.05 \mu\text{m}$ ). To the contrary, SP particles dominate in AE horizon (Figure 4). CIA values (chemical index of alteration) are high, increasing from 77% in Ah to 85% in EBT (Figure 4). The lowermost part of the profiles records slightly lower percentages.

In the Eemian soil (OL2) the silt fraction is dominant (55–70%). The whole profile shows a low proportion of sand. Clay content reaches a maximum in Btg2 horizon, although (E1) and (E2) horizons on the top show a high proportion as well (Figure 4). The values of magnetic susceptibility ( $\chi$ ) are very low in (E1) and (E2) but increase in Btg.  $\chi_{fd}\%$  values are very low in the whole profile, revealing that none of the horizons has SP contribution (Figure 4). CIA pattern shows a maximum in the Btg2 horizon (87%) and reach the lowest value in BCg (80.3%).

OL3 has a higher percentage of sand (7 to 12%) than the Eemian soil, however, silt still dominates (61 to 70%). In this unit the clay amount is lower (25–29%) and there is no significant difference among the horizons, although a slightly higher content is obvious in Btg2 (Figure 4). In case of magnetic parameters,  $\chi$  has similar values, while  $\chi_{fd}\%$  shows strong differences varying from zero in the (E1) to 14% in (E2), 5% in Btg1, 14% in Btg2, and zero in BCg. This contrast evidences the presence of discontinuities (Figure 4). CIA values are also very similar (83–85%). The highest weathering degree corresponds to Btg1 (Figure 4).

OL4 profile shows very low and homogeneous  $\chi$  values, but high percentages of  $\chi_{fd}\%$  indicate an elevated contribution in SP particles. However, this behavior is not continuous, because in Bg2, in the lowermost part of BCg2 (in 180–200 cm depth), and in the G horizon,  $\chi_{fd}\%$  drops to zero (Figure 4). CIA varies from 84% in Bg horizon decreasing to 82% in BCTg2.

OL5 has high variability in the proportions of the different size fractions, which reveal the influence of alluvial sedimentation. Sand varies from 8% in the Bg horizon to 42% in the lowermost part of the profile, where gravel appears. OL5

has very low values in  $\chi$ , although an enhancement without contribution of SP grains is noted in the 3Ah horizon, which is detectable in the uppermost part of the profile in the Bg and G horizon (Figure 4). CIA is equally variable. The upper Bg as well as G horizon are slightly less weathered than 2Bg and 2BCg and 3Ah and 3Cg show the lowest CIA percentages.

## 4 Discussion

### 4.1 Chronostratigraphy and correlation of the Oberlaab profile

Although no instrumental absolute dates are available in Oberlaab it is possible to establish a local chronostratigraphic scheme taking into account the stratigraphic classification of SCHOLGER & TERHORST (2011) and TERHORST et al. (2011) developed for the Middle Pleistocene profiles in the Northwest of Austria.

TERHORST (2007) points out that the Oberlaab sequence contains four interglacial paleosols (1st, 2nd, 3rd, 4th) thus covering at least the last five glacial periods (thus the development of the Mindel terrace (Younger Deckenschotter) occurs at least during MIS 12). According to the stratigraphic schemes, Oberlaab records completely the main paleoclimatic stages from MIS 12 to MIS 5e, parts of the last glacial, as well as the Holocene soil. This interpretation is in good agreement with the pedostratigraphical schemes of the cover layers in Wels-Aschet, Neuhofen, and Oberlaab and our studies can clearly be positioned in the regional stratigraphic context of the area. This contradicts to the view of PREUSSER & FIEBIG (2009), who propose younger ages for the Wels-Aschet sequence including the Older Deckenschotter of the Günz glaciation, based on luminescence dating. For instance, they suggest an age of  $252 \pm 29\text{ky}$  for the 4th interglacial paleosol (MIS7). Paleomagnetic studies done in the same sequence also contradict the chronostratigraphy of PREUSSER & FIEBIG (2009) documenting the presence of the Calabrian Ridge excursions: CR1 (325–315 ka) and CR2 (525–515 ka) inside the 3rd and in the base of the 4th interglacial pedocomplex, respectively (SCHOLGER & TERHORST 2011). Consequently, OL5 can be attributed to MIS 11 with three phases of soil formation: Bg-G-2Bg-2BCg-3Ah-3EBg-3Bg-3Cg. The upper Bg horizon correlates with the 4th interglacial paleosol, described by TERHORST (2007) and TERHORST et al. (2011) in Oberlaab, Neuhofen, and Wels-Aschet, but in our case clay coatings are not observed in this unit as in other sequences. TERHORST (2007) mentions an underlying G horizon in Neuhofen and Wels-Aschet, but without the 2Bg-2BCg horizons, which are present in our section. Furthermore, the lower paleosol has not been described before in the area. There are clear differences in analytical properties, which reflect soil formation cycles: CIA values are low in the Bg-G, 3Ah, 3EBg, and high in 2Bg-2BCg, 3Bg horizons (Figure 4).

The subsiding stratigraphic levels of OL4 and OL3 correlate with the 3th and 2th interglacial paleosols described for Upper Austria (TERHORST 2007; 2013, this volume; TERHORST et al. 2011). OL4 is developed on sediments which are classified as MIS 10 deposits and represent an intensive glacial phase. Therefore paleosol formation could correspond best to the MIS 9 (c.f. LISIECKI & RAYMO 2005). On

the base of the pedostraphic observations, the next sedimentation events can be correlated to the MIS 8 period, which records a weaker glacial phase (c.f. LISIECKI & RAYMO 2005). Pedogenesis of the OL3 unit can be related to the interglacial period of MIS 7.

Both profiles OL3 and OL4 are pedocomplexes (TERHORST 2007; TERHORST et al. 2012). According to our field observations significant differences and discontinuities are not detectable between both units. However, some quantitative analytical characteristics clearly indicate them. In OL4 we can separate three phases of pedogenesis: the first phase corresponds to the development of the lowermost G horizon, reflected by different properties: CIA values are higher than in the overlying BCtg2,  $\chi_{fd}\%$  is zero with no contribution of SP magnetic particles. The second phase comprises the horizons BCtg1 and BCg2, which tend to have similar characteristics although CIA values are higher in BCtg1, a fact which is related to a higher amount of clay. The third phase corresponds to the formation of the upper horizons Bg1-Bg2-BCg1- BCg2. These results are not in good agreement with those obtained by TERHORST (2007), who reports the presence of a pedocomplex in Oberlaab with only 2 paleosols, one described as a Bw horizon with stagnic properties superimposed by a Btg horizon. In Wels-Aschet, TERHORST et al. (2012) find the 3th paleosol, containing two Bt horizons with moderate weathering degree. Coincidences are in terms of the gleying process, because in Oberlaab as in other sections in Lower Austria, elevated amounts of Mn have been found in different forms.

On the other hand, OL3 is another pedocomplex. Repeatedly, the profile has been characterized as monogenetic during field survey, however, analytical results could highlight specific differences, although the clay content is similar (Figure 4). CIA index and  $\chi$  reveal maxima in the Btg1 horizon, which can be related to the high amount of clay (around 30%).  $\chi_{fd}\%$  values record significant differences in the contribution of magnetic particles, and thus can be explained by the presence of discontinuities (Figure 4). We have even found charcoal in Btg1 – probably because this horizon was close to the surface. Our results exhibit a discrepancy to those obtained by TERHORST et al. (2012) for the Wels-Aschet sequence, because on the basis of mineralogical analyses the authors conclude that the 2th paleosol is less weathered than the 3th one. Similar mineralogical results were obtained for earlier studied Oberlaab profiles (TERHORST et al. 2003). However, the presented CIA values, clay contents (40–32%), as well as thick clay coatings observed in our study indicate a more advance weathering in the OL3 unit in comparison to OL4, and that means 2nd paleosol is more weathered than the 3rd one).

The multiphase pedogenesis detected in both units, OL3 and OL4, could be linked to the climatic fluctuations within MIS 7 and MIS 9. These stages demonstrate a complex pattern in the  $\delta^{18}O$ -record with warm intervals alternating with cold episodes. Earlier BRONGER et al. (1998) supposed that pedocomplexes PKII and PKIII in loess-paleosol sequences of Tadjikistan reflect climatic fluctuations within MIS 7 and MIS 9, respectively. More detailed correlation of pedocomplexes and marine isotope curve is limited by the lack of a reliable age control.

The OL2 profile, regarded as the Eemian soil (1st inter-

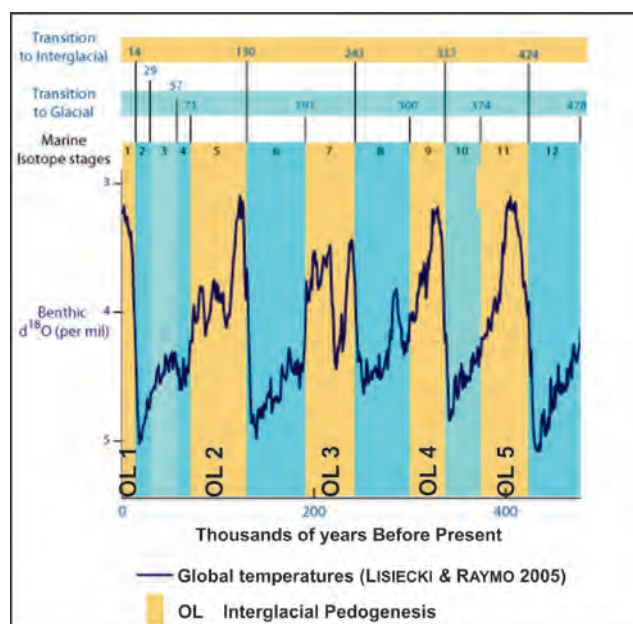


Fig. 5: Correlation of paleosols in Oberlaab with the Marine Isotope Curve.  
Abb. 5: Korrelation von Paläoböden in Oberlaab mit der marinen Isotopenkurve.

glacial paleosol), occupies the next stratigraphic level. Its age has been well established by several instrumental dates: OSL and paleomagnetism in other sections, in particular in the Wels-Aschet sequence (PREUSSER & FIEBIG 2009; SCHOLGER & TERHORST 2011), but in the case of Oberlaab such results are not available. In Wels-Aschet, the Eemian soil is developed on loess sediments of the Riss glaciation, which is  $142 \pm 18$  ky (MIS 6) according to PREUSSER & FIEBIG (2009). On top of the MIS 5e-Eemian soil the Blake event ( $\sim 117$  ka) is proved by SCHOLGER & TERHORST (2011; 2013, this volume) in Wels-Aschet. The Eemian soil in unit OL2 is correlated to MIS 5e in this study, based on the results of further sequences in Austria. In general, the paleosol displays fewer stagnic properties and less compaction than the older paleosols. Furthermore, its stratigraphic position is situated below the Würmian sediments. From the analytical point of view the interglacial pedogenesis is proved by intensive clay illuviation, coinciding with higher CIA values.

OL1, which correspond to the Holocene soil, is separated from the Eemian paleosol by loess-like deposits characteristic for the Würmian glaciation. There is no evidence for well-developed soils in these deposits, only reworked layers are present.

According to the proposed pedostratigraphy, Figure 5 shows the correlation of the study section with the Marine Isotope Curve (acc. to LISIECKI & RAYMO 2005).

#### 4.2 Pedogenic trends in Oberlaab induced by climate change or/and geomorphic processes

The loess-paleosol sequence in Oberlaab contains several units, which developed different specific properties. Although some of the specific patterns are found in the whole section, their intensity is strongly related to pedogenic and geomorphic processes. All units are characterized as pedo-

complexes, and can be welded, thus difficulties in the recognition of every single soil cycle can emerge.

In the sequence we studied it is clear that the lower units, especially OL4 and OL5, developed under conditions produced by water saturation, which gave rise to redoximorphic processes. The lowest OL5 unit shows the formation of a diagnostic G horizon with a pale reduced soil matrix and relatively few iron segregations. These properties correspond to gleyzation in a permanently reduced soil environment, which lets us suppose a shallow groundwater table. OL4 demonstrates patchy Bg horizons, Fe-Mn concretions and hypocoatings with bleached groundmasses. In fact, the highest concentrations of ferruginous pedofeatures as well as bleached zones are found in OL4 (Figure 3). These characteristics correspond to stagnic properties related to temporary saturation with surface water. Very limited development of clay illuviation agrees with the hypothesis of pedogenesis in a poorly drained soil environment.

The stagnic and gleyic features become less frequent in the upper paleosols. Here, clay illuviation is dominant, with thick and dark clay coatings. From micromorphological results the Eemian (OL2) and Holocene soils have the highest percentages of illuvial clay (2.1 to 4%). On the other hand, coatings in OL3, OL4, and partly in OL2 (for details see SEDOV et al. 2013, this volume) are deformed and displaced, thus the material has been also affected by gleyzation and cryogenesis. In fact, only in the upper part of the modern soil clay coatings are pure with a yellow color. Its inferior part (G horizon) contains the highest amount of clay coatings, which are “dirty”, laminated or having reddish colors, with pigmentations. This horizon has also a low magnetic susceptibility value and considerably lower weathering degree. TERHORST (2007) suggests it corresponds to a previous phase of soil formation that probably occurred during MIS2 to 3.

In summary, the main trend of change of pedogenic processes in the Oberlaab sequence is the following: gleyzation at the base (OL5-OL4), stagnic processes in the central part (OL4-OL3) and clay illuviation in the upper units (OL2-OL1).

Which soil forming factors control these pedogenic trends? It seems that climate is only one factor due to the fact that all units are formed under humid climates evidenced by the formation of stagnic and gleyic features, clay neoformation and illuviation, and silt and/or iron illuviation. Differences between the pedocomplexes are well expressed in the intensity of the involved processes. Additionally, changes in temperature may be somewhat responsible. Lower units could reflect the influence of cooler environments, while the upper ones are affected by warmer climates.

The geomorphic position of the studied paleosols also evidently affected pedogenesis. The lowermost paleosols are situated at a lower level in the geoform, and were more affected by waterlogged conditions (high groundwater table or flooding). This is evident in OL5, where fluvio-glacial processes formed a gravelly alluvial deposit with a high content of Mn concretions.

The upper paleosols are located in better drained positions away from the influence of groundwater and inundations. This can be explained by the aggradation of the terrace and by lower levels of the younger terraces. In particular, the Eemian soil shows a well-developed formation stage charac-

terized by a high weathering status. The well-drained geomorphological position favors clay illuviation.

Another important process observed in the lower paleosols is related to deformation of pedofeatures, which can be caused by bioturbation, as well as by cryogenesis and gelifluction. Orbiculate fabrics, silt concentrations, and platy structures are related to such kind of processes. According to our observations these features are more frequent in OL5 and OL4 (Figure 3) due to a higher intensity of glacial periods. However, lower sedimentation and/or more intensive erosion processes cannot be neglected. According to LISIECKI & RAYMO (2005) during MIS 12 and 10 severe glacial phases occurred, after which OL5 and OL4 are formed (Figure 5). During MIS 8 the glacial phase is much weaker, thus characteristics found in OL3 are clearly less affected by cryogenesis.

## 5 Conclusion

The Oberlaab sequence on top of a Mindel terrace (Younger Deckenschotter) is constituted by four interglacial paleosols, which developed most probably during the Marine Isotope Stages MIS 11, MIS 9, MIS 7, and MIS 5, respectively. Similar to other interglacial paleosols of the same age in Upper Austria, the paleosols are formed as pedocomplexes in Oberlaab. Studied properties reveal the dominance of humid conditions in the whole section, but with differences in the intensity of pedogenic processes involved in the formation of each pedostratigraphic level. We assume that the oldest paleosol OL5 can be correlated to the 4th interglacial paleosols found in Neuhofen, Wels-Aschet, and in earlier studied profiles in Oberlaab. We correlate this soil to MIS 11. The pedocomplex is affected by strong gleyzation. OL4, which is the 3<sup>rd</sup> interglacial paleosol, is characterized by the signs of strong reductomorphic features, partly deformed and with orbiculate fabrics, which are associated with later cryogenesis, probably during MIS 8. OL3 also has properties related to stagnic processes, but clay illuviation and a higher degree of weathering are evidence of its formation under warmer conditions during MIS 7. The Eemian soil (OL2) is the best developed and reflects warm interglacial conditions.

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# Grain size and mineralogical indicators of weathering in the Oberlaab loess-paleosol sequence, Upper Austria

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**Abstract:** Grain size analyses, bulk and clay mineralogical data were used to characterize weathering within the loess-paleosol-sequence of Oberlaab in Upper Austria. Soil horizons can be clearly identified by the calculation of weathering index Kd from granulometric parameters.

The mineralogical composition of the bulk samples shows increasing weathering intensity from the top to the bottom. The weakest weathering stage 1 is not present in Oberlaab, because all samples are free of carbonate minerals. Weathering stage 2 can be found in the upper part of the profile, whereas stage 3 is mainly present in the lowermost horizons. The highest weathering stages 4 and 5 are not present in Oberlaab. The clay mineral distribution in the profile is dominated by the disappearance of primary chlorite in the upper part of the profile and the neoformation of vermiculites from illite by pedogenesis in the lower part.

Two different types of mixed layer minerals were found in the pedocomplexes. An illite/chlorite mixed layer mineral occurs following the disappearance of chlorite and is present in the Eemian luvisol. The second mixed layer mineral consists of illite/vermiculite and is present in the whole profile.

The weathering stages obtained from the clay mineral composition are slightly lower than that of bulk mineralogy, but reach as well stage 3 in the lower part of the profile.

## Korngrößen und mineralogische Verwitterungsintensitäten in der Löss-Paläoboden-Sequenz von Oberlaab, Oberösterreich

**Kurzfassung:** Untersuchungen des Gesamt- und Tonmineralbestandes sowie der Korngröße wurden zur Charakterisierung der Verwitterungsstadien in einer Löss-Paläoboden-Sequenz in Oberlaab (Oberösterreich) verwendet. Durch den aus der Korngrößenzusammensetzung berechneten Verwitterungsindex Kd können Bodenhorizonte eindeutig identifiziert werden.

Der Gesamtmineralbestand zeigt, dass die Verwitterungsintensität mit der Tiefe zunimmt. Das schwächste Verwitterungsstadium tritt in Oberlaab nicht auf, da alle Proben karbonatfrei sind. Verwitterungsstadium 2 kommt in den oberen Bodenhorizonten vor und ist durch das Fehlen von primärem Chlorit in der Tonfraktion gekennzeichnet. Die untersten Horizonte entsprechen mit der Neubildung von Vermikuliten aus Illit Verwitterungsstadium 3. Die intensivsten Verwitterungsstadien 4 und 5 treten in diesem Profil nicht auf.

In den Pedokomplexen wurden zwei verschiedene Mixed-layer-Mineralen gefunden. Ein Illit-Chlorit-Mixed layer kommt in der Parabraunerde des Eem-Interglazials vor, ein Illit-Vermikulit-Mixed layer ist im gesamten Profil nachweisbar.

Die aus der Zusammensetzung der Tonfraktion ermittelten Verwitterungsstadien sind etwas niedriger als die aus dem Gesamtmineralbestand, erreichen aber im untersten Teil des Profils ebenfalls Stufe 3.

**Key words:** *Paleosols, clay minerals, vermiculite, secondary chlorite, weathering index Kd*

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## 1 Introduction

The weathering status of Pleistocene loessic paleosols provides an important proxy for interglacial paleoenvironments as the processes of primary mineral breakdown and neoformation of secondary components are strongly dependent upon the bioclimatic conditions of pedogenesis, (TERHORST, 2013, this volume). Geochemical indicators (various coefficients based on ratios of different major and trace element concentrations) are widely used to evaluate weathering (BUGGLE et al., 2011). For the Oberlaab sequence this

approach was applied by SOLLEIRO-REBOLLEDO et al. (2013, this volume). Paleomagnetic stratigraphy of the Middle to Upper Pleistocene loess/paleosol sequence of Wels-Aschet is presented by SCHOLGER & TERHORST (2013, this volume). A detailed overview, as geographical position and geology of Oberlaab as well as the detailed profile description can be found in TERHORST (2013, this volume).

An important additional set of weathering indicators is related to the grain size distribution and mineralogical composition of total soil samples and clay fraction. The products of pedogenic silicate alteration are concentrated mostly in

the fine material. This justifies to use the ratios of different size fractions as an estimate of this process. PÉCSI & RICHTER (1996) proposed the weathering index Kd as an integral measure of weathering intensity in paleosols and pedocomplexes. Additional information can be obtained from the proportion of the clay fraction and specifically of the fine clay fraction. The latter consists mostly of products of more advanced transformation and synthesis of secondary minerals, whereas coarse clay is produced mostly from physical breakdown accompanied by moderate chemical transformation (CHAMLEY, 1989).

The mineralogical composition of total soil samples (although only semi-quantitative) provides hints to estimate the weathering status, through the relative abundance of weatherable components (e.g. chlorite, amphiboles, etc.) and stable minerals (especially quartz). Clay mineral assemblages are sensitive indicators of paleopedogenesis and paleoenvironments. The formation of different clay minerals in modern soils is highly dependent upon the soil processes and regimes, which in turn are controlled by climatic conditions (DIXON & WEED, 1989). Clay mineral studies in loess-paleosol sequences thus provide important proxies for the Pleistocene climatic history (BRONGER & HEINKELE, 1990; BRONGER et al., 1998).

The aim of this study is to estimate the weathering status of the Pleistocene paleosols in the Oberlaab loess profile using grain size, bulk mineralogy and clay mineral assemblages. We further compare these data with other proxies available for this profile to incorporate them into the integral paleoecological interpretation.

## 2 Methodology

### 2.1 Grain size distribution

The grain size distribution was determined by combination of wet sieving of the fraction >20 µm and automatic sedimentation analysis with a Micromeritics SEDIGRAPH III 5120.

50 g of the dried sample were treated with 200 ml 10% H<sub>2</sub>O<sub>2</sub>. The purpose was the oxidation of organic components and a proper dispersion of the sample. After approximately 24 hours reaction time the unused H<sub>2</sub>O<sub>2</sub> was removed by heating. After ultrasonic treatment the sample was sieved with a set of 2 mm, 630 µm, 200 µm, 63 µm and 20 µm sieves. The coarse fractions were dried at 105°C and stated in mass percent. The <20 µm portion was treated with 0.05% sodium polyphosphate and analyzed in the sedigraph by X-rays, according to Stoke's Law. From the cumulative curve of the sedigraph and the sieving data the grain size distribution of the entire sample was calculated.

A general overview of the grain size distribution in Oberlaab is presented by SOLLEIRO-REBOLLEDO et al. (2013). In this paper we present mostly the data relevant for the evaluation of the soil weathering index Kd. Silt (2–63 µm) including coarse silt (20–63µm) and clay (<2 µm) including fine clay (<0.2 µm) fractions are evaluated.

### 2.3 Total mineral analysis

The dried samples, ground in a rock mill to analytical size, were prepared according to the backloading method and X-rayed with a Panalytical XPert Pro MPD diffractometer with

an automatic divergence slit, Cu LFF tube, 45 kV, 40 mA, and an X'Celerator detector. The samples were measured from 2° to 70° 2Θ. The X-ray diffraction patterns served as the basis for calculating the qualitative mineral content.

### 2.4 Clay mineral analysis

The samples were dispersed with 10% hydrogen peroxide. After the reaction had subsided and the extra H<sub>2</sub>O<sub>2</sub> was removed, they were exposed to ultrasonic sounding for 15 minutes. The <63 µm fraction was obtained by wet sieving, and out of this the 2 µm fraction by centrifugation.

The next step was cation exchange. 40 ml of the clay suspension were each mixed with 10 ml 4 N KCl solution and 4 N MgCl<sub>2</sub> solution, respectively, and shaken for 12 hours. The texture specimens were placed on ceramic platelets, onto which the clay suspension was sucked via low pressure.

After treatment with ethylene glycol, DMSO (dimethyl sulfoxide) and heating 2 hours at 550 °C the samples were measured in the diffractometer and evaluated according to the same principle as the total mineral analysis. In general, the identification of the minerals and clay minerals was carried out according to BRINDLEY & BROWN (1980) and MOORE & REYNOLDS (1997).

### 2.5 Weathering intensity

In the context of this study, the data collected from the total as well as the clay mineral analysis serve as a basis for estimation of the weathering intensity of the individual horizons. It is assumed that the most sensitive minerals, such as carbonates and chlorite, will be dissolved or replaced first, and, along with increasing weathering, the more stable minerals, such as mica and feldspars.

This method was used for the first time in the loess paleosol profile Wels-Aschet (TERHORST et al., 2012) and slightly modified for the Oberlaab profile (Tables 1 and 2).

Based on the clay mineral analysis, a classification of the weathering intensity was carried out with the presence of primary chlorite on the one hand, and vermiculite varieties on the other hand TERHORST et al. (2012).

## 3 Results and discussion

### 3.1 Grain size analyses of silt and clay


The variation in clay content (<2 µm) is relatively high. The lowest clay content was found in a loess sample of OL4 with 22.7 mass%, the highest amount of clay was found in clay cutans from the same OL4 profile and is 64.3 mass% (Table 3). Apart from the clay cutans the highest clay contents occur in the Middle Pleistocene paleosols of profile OL5 (Table 3 and Fig. 1).

The Eemian interglacial Luvisol (OL2) in 120–140 cm has a very high clay content of 51.3 mass%, and especially in the deepest part of the profile the clay content is high.

The paleosols of the OL4 profile section are characterized by slightly lower clay contents; only in one sample more than 40 mass% of the clay fraction were detected. The more recent loess samples of the upper profile sections (OL1) have a clay content of less than 30 mass%.


Tab. 1: Weathering intensity according to bulk mineralogy after TERHORST et al. (2012).

Tab. 1: Verwitterungsintensität auf der Basis der mineralogischen Ergebnisse nach TERHORST et al. (2012).

Degree of weathering		Index minerals
Stage 1		Carbonate minerals present
Stage 2		Carbonates weathered, chlorites traceable
Stage 3		Micas present, chlorite absent
Stage 4		Micas absent [weathered]
Stage 5		Micas and feldspars absent [weathered]

Tab. 2: Weathering intensity according to the clay mineralogy, after TERHORST et al. (2012), slightly modified.

Tab. 2: Verwitterungsintensität auf der Basis der tonmineralogischen Ergebnisse nach TERHORST et al. (2012), leicht verändert.

Degree of weathering		Index Minerals
Stage 1		Primary chlorite present
Stage 2		Primary chlorite absent, illite or vermiculite 14Å dominant
Stage 3		Vermiculite 18 Å dominant, smectite from lessivation dominant
Stage 4		Vermiculite 18Å dominant, vermiculite 14 Å completely transformed
Stage 5		Illite totally weathered

The fine clay content (< 0.2 µm) varies much more than the total clay. The clay cutans from the OL4 profile section contain 54.6 mass% fine clay, the Eemian paleosol from 120–140 cm (OL2) 31.7 mass%. In the younger loess samples the value for the fine clay is below 10 mass%, the lowest value of 3.7 mass% was found in a loess sample of OL3 (Table 3 and Fig. 2).

As expected, the silt fraction is predominant in most samples and reaches values up to 74.0 mass% (OL4). By contrast, contents of the silt fraction in the paleosols are significantly lower. In the basal parts of the sequence (OL5), the silt fraction diminishes far below 50 mass%. The lowest silt value (33.6 mass%) can be found in the clay cutans of profile section OL4.

The decrease of the silt fraction within the paleosol horizons is significant. This trend is particularly obvious in the Eemian interglacial paleosol (OL2), in which the silt fraction is only 45.6 mass% (Table 3).

Generally speaking, the silt content is higher in the upper part of the profile and lower in the deeper parts because of the known clay enrichment during pedogenesis (Table 3).

### 3.2 Kd-values and weathering intensity

The grain size distribution can be used as an indicator for the degree of weathering. Clay illuviation in Bt horizons causes an enhancement of the fine fraction at the expense of the coarse fraction. The individual soil horizons can be recognized either by the increased values of the clay and fine clay fraction respectively (Figs. 1 and 2), or by significantly lower values of the silt and coarse silt fraction.

According to PÉCSI & RICHTER (1996), the weathering index Kd allows to measure the weathering intensity of pale-

osols and pedocomplexes. It is calculated by dividing of the coarse and middle silt content by the fine silt and total clay content. In general, the lower the Kd-value is, the more weathered is the soil horizon (Fig. 3). The lowest Kd-value (0.32), and thus by far the highest degree of weathering, is obtained in the paleosols of profile OL5 and in the clay cutans.

In the Eemian paleosol (OL2) the Kd-value of 0.55 shows that the weathering process – caused by an intensive pedogenesis – is remarkable in this part of the profile as well.

Figure 3 shows clearly the less weathered upper part, a mixed part in the middle and the highly altered deepest part of the OL5 profile section.

Summarizing, paleosols are indicated by the lowest Kd-values in the studied profile. The enrichment of clay at the expense of the coarse grain fraction can be observed in many samples with low Kd-values.

### 3.3 Mineralogical results

#### 3.3.1 Bulk mineral analysis

All examined samples are free of carbonate (Table 4). Neither the uppermost more recent loess samples nor the samples from the Mindel terrace contain any calcite or dolomite.

Chlorite occurs in higher amounts only in section OL1 and partly in OL2. In OL3 hardly any chlorite is detectable, but traces were present in OL4. In OL5 due to strong weathering processes no chlorite was found (Table 4).

Hornblende - also a mineral relatively sensitive to weathering - could be detected in traces in the profile sections OL1 and OL2, but not in the Eemian paleosol. In older horizons hornblende is not present because of strong weathering (Table 4).

Generally, quartz occurs as a main component throughout the profile in uniform distribution and without recogniz-

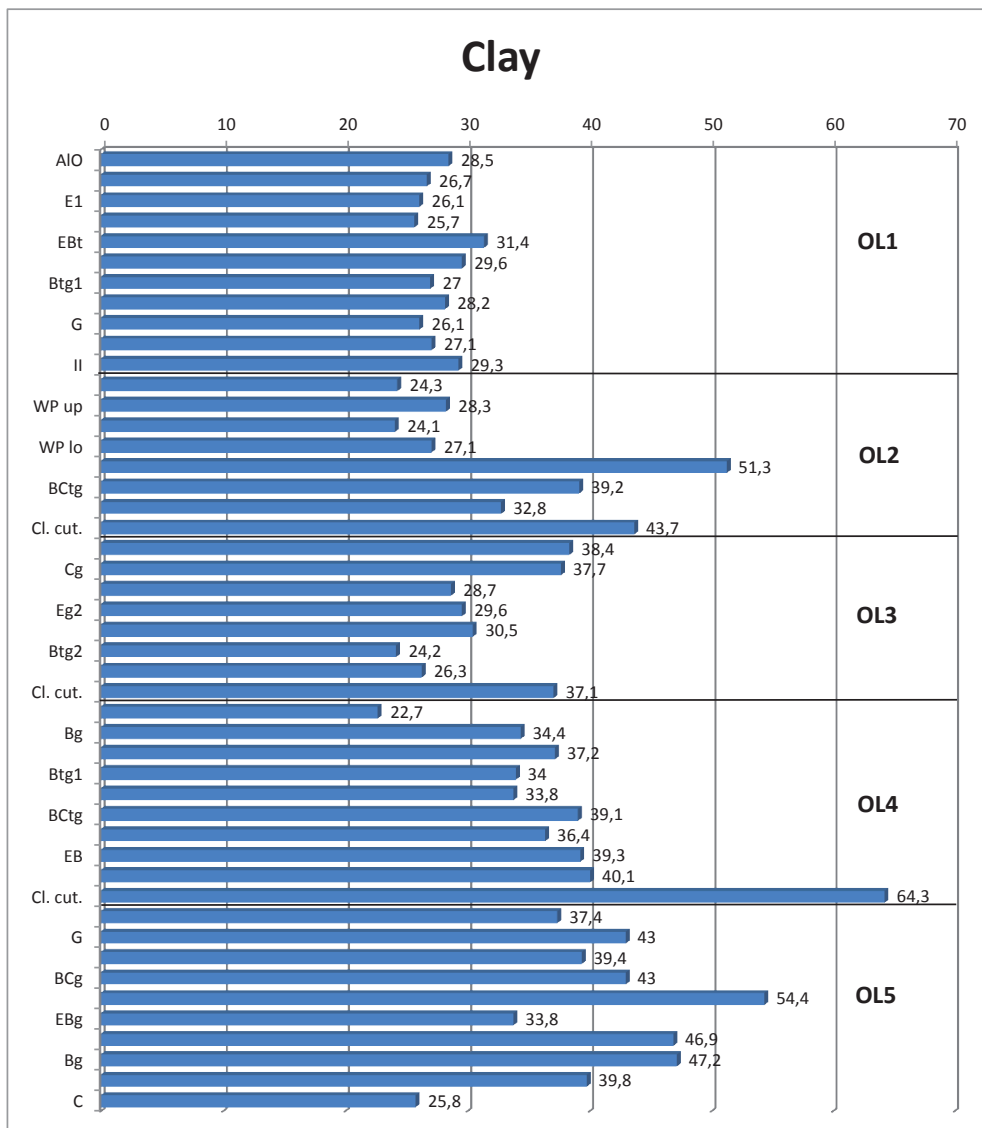


Fig. 1: Total clay content in mass% in the profile Oberlaab, WP = pedosediments.  
 Abb. 1: Gesamt-Tongehalt im Profil Oberlaab in Masse%, WP = Bodensedimente.

able trends. Plagioclase is detectable throughout the profile. Potassium feldspar occurs in traces only. In Table 4 the values for both types of feldspars are combined in the column “feldspars”.

The distribution of mica is also more or less homogeneous, except for clay cutans, where the amount is higher and in some samples of profile section OL5 it was found only in traces. Layer silicates, which represent the sum of all clay minerals and mica are present in all samples, whereby the highest content can be found in the Bt horizon of the Eemian interglacial paleosol (OL2). The peaks of about 14 Å originate from the clay minerals smectite and vermiculite, which will be discussed in detail in the following chapter.

The total mineral content can be used to estimate the weathering intensity of the individual horizons as based on the presence or absence of indicator minerals (cf. Tab. 1 and 2). Weathering stage 1 represents the horizons with the lowest weathering intensity. In this stage the horizons still contain carbonate minerals. In Oberlaab this stage is not present, as all samples are free of carbonate.

Most horizons can be ascribed to weathering stage 2. These horizons are characterized by the presence of chlorite with a simultaneous absence of carbonate. All samples from profile section OL1 and OL2, and few samples of section OL3

and OL4 belong to this weathering stage. Only section OL5 is totally free of chlorite.

Weathering stage 3, in which primary chlorite is absent but mica is still present, comprises the samples from OL 5 and some horizons from OL3 and 4.

The highest weathering stages 4 and 5, which are characterized by the absence of mica and feldspars and by the occurrence of high amounts of layer silicates, are not developed in the Oberlaab profile.

In all horizons, the iron hydroxide goethite could be found in varying amounts.

### 3.3.2 Clay mineral analysis

Clay minerals of the profile show strong dynamics in terms of rearrangement, particularly triggered by vermiculite.

Classical vermiculite, which expands only to 14 Å, is present in the lower part of the profile in small amounts with only few exceptions (Table 5). It occurs mainly in the moderately to stronger weathered horizons. The younger profile sections are almost free of vermiculite 14 Å.

Advanced weathering leads to the formation of the more strongly expanding vermiculite variety (vermiculite 18 Å). This mineral is present in intensely weathered horizons, in ad-

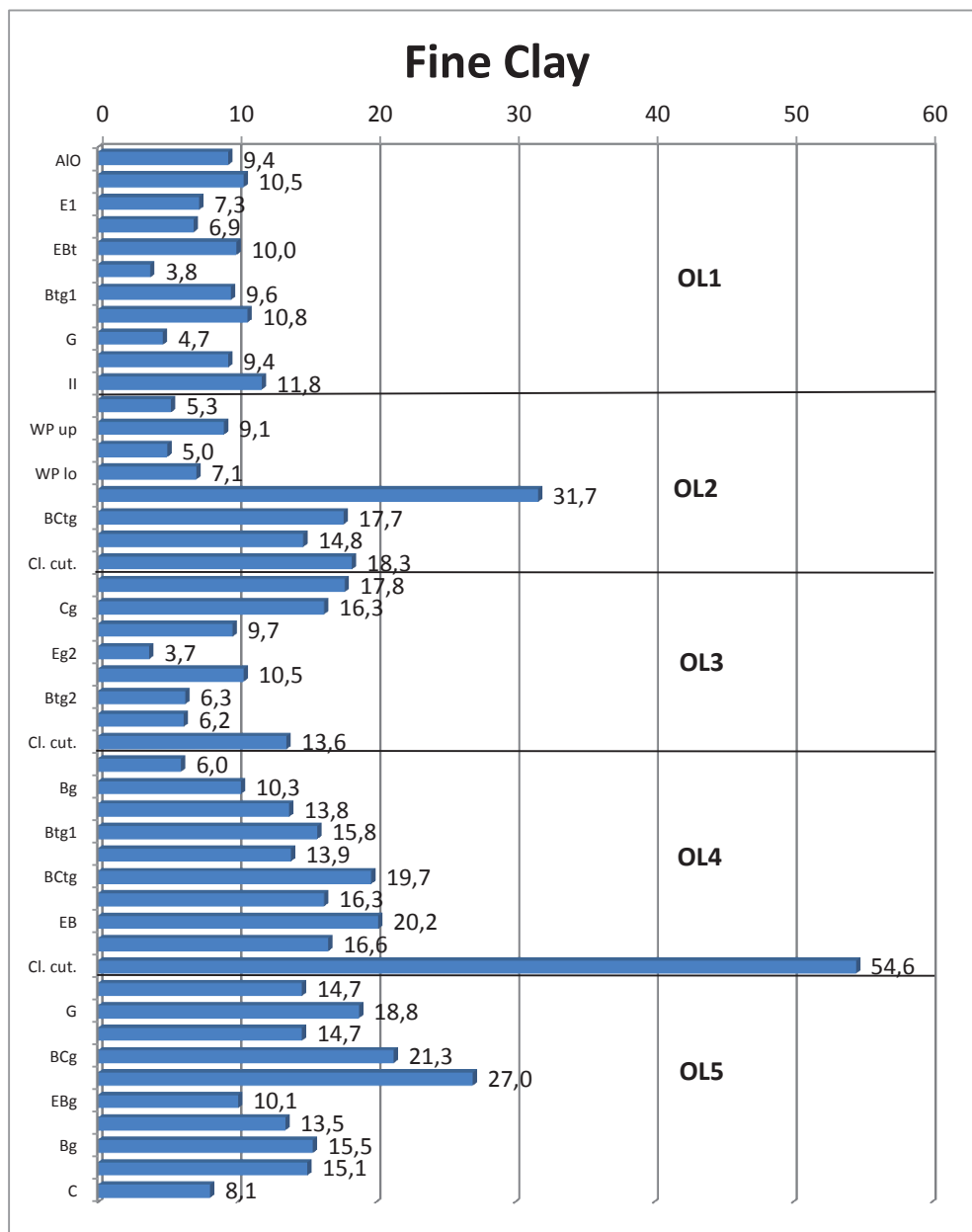


Fig. 2: Fine clay content in mass% in the profile Oberlaab, WP = pedosediments.

Abb. 2: Feintongehalt im Profil Oberlaab in Masse%, WP = Bodensedimente.

dition to vermiculite 14 Å. In the case of increasing weathering intensity it replaces vermiculite 14 Å completely. In the studied profile it is most frequently present in the older profile sections OL4 and OL5. Smaller amounts are also present in the rest of the profile but without regular distribution or trends.

In all horizons, smectite can also be found in relatively small amounts, in some parts also as a main component of the clay fraction. It is detectable both in soil horizons and in loess sediments. Its occurrence in less weathered loess layers gives reason to believe that smectite was not newly formed within the profile, but rather originates from previously weathered source material. In contrast, the occurrence of the highest amounts of smectite in the Eemian paleosol may originate in neof ormation of smectite in those horizons or is caused by lessivation.

The amount of illite, the source material for more weathered clay minerals, does not vary much in the profile, only small amounts are detectable.

Kaolinite can be found in all horizons in small amounts and does not show any recognizable trends. Kaolinite occurs

in the well crystallized form expandable with DMSO and in a poorly crystallized form (fire clay).

Primary chlorite occurs occasionally and in small amounts. It is clearly detectable only within the young loess sediments (OL1), while it is absent in all older horizons.

Secondary chlorite, which is found occasionally in paleosols, is not clearly detectable in the Oberlaab profile.

Most horizons contain two distinguishable mixed layer minerals, which consist of the components illite, chlorite or vermiculite, respectively. The chlorite containing mixed layer mineral occurs mostly in horizons of OL1 following the disappearance of chlorite. It is also present in OL2 and disappears almost completely in the lower horizons. In the clay fractions of all other horizons an additional variety of mixed layer minerals can be found, which is composed of vermiculite and most likely illite (Table 5).

The results from the clay mineral analyses were also classified according to their weathering intensity. This was mainly based on the presence of the vermiculite varieties (Tables 2 and 5).

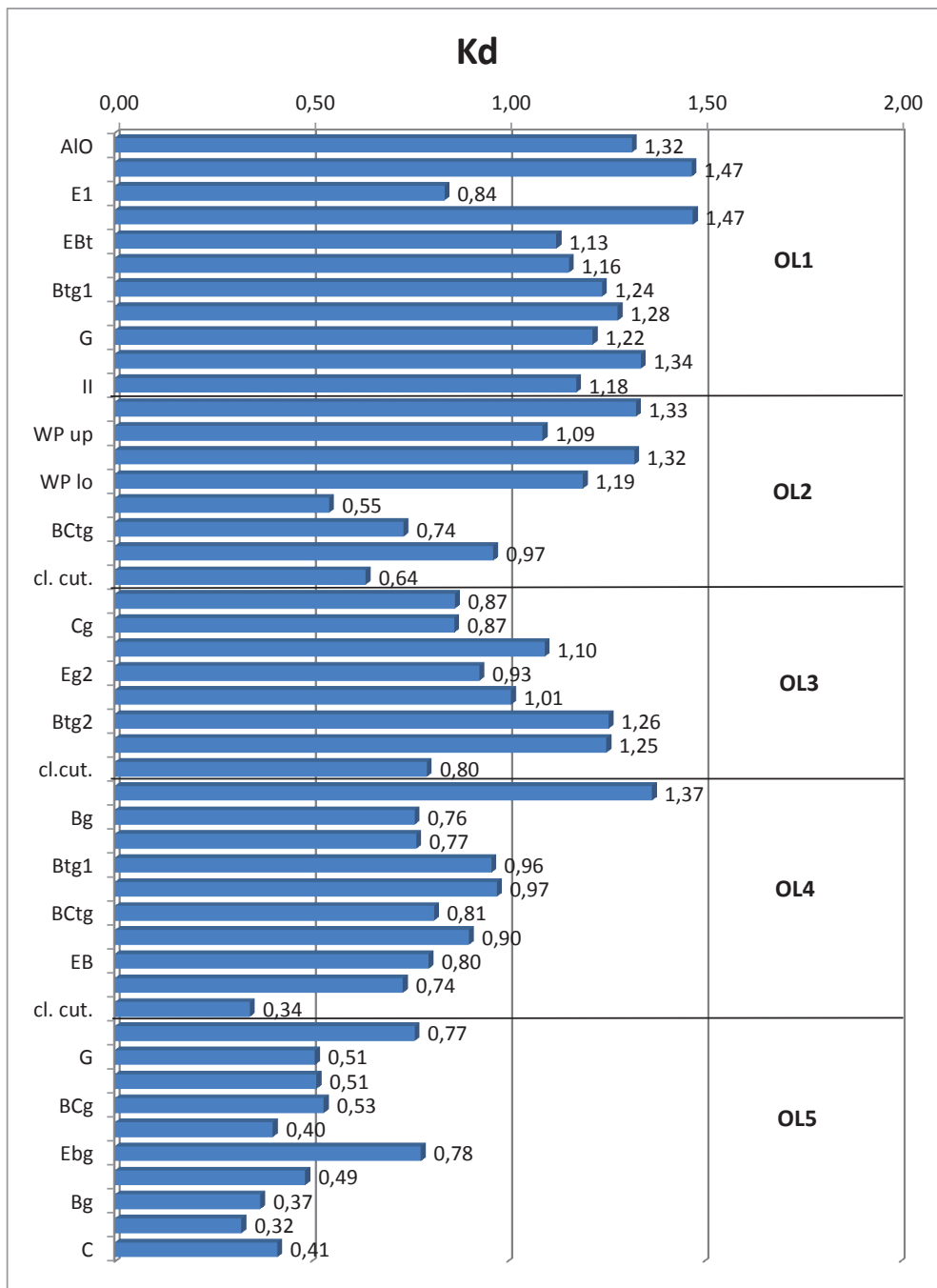


Fig. 3: Distribution of weathering index Kd in the profile Oberlaab, WP = pedosediments.  
 Abb. 3: Verteilung des Verwitterungsindex Kd im Profil Oberlaab, WP = Bodensedimente.

The weakest weathering stage 1, which is characterized by the presence of primary chlorite, only occurs in the uppermost horizons of OL1. Below that, all horizons are at least at weathering stage 2, where by definition illite or vermiculite 14 Å are the dominant clay minerals.

Most horizons below OL2 contain higher amounts of vermiculite 18 Å, which means they already belong to weathering stage 3. By means of lessivation processes, clay cutans were transported into the larger pores and fissures of the underlying loess loams. This is also reflected by the occurrence of weathering stage 3.

Weathering stage 4 is only developed in the upper part of the Middle Pleistocene paleosol OL5.

#### 4 Discussion

The results show that the paleosol units of Oberlaab are de-

veloped as pedocomplexes. The basal parts of the sequence show the highest weathering degree, although the intensity of weathering never reached the highest stages as presented for the nearby loess/paleosol sequence in Wels-Aschet (TERHORST et al., 2012).

Grain size and mineralogical indicators of weathering provide quite similar estimates of the weathering status for the different pedocomplexes of Oberlaab. All these indicators point to the lower weathering grade of the recent Luvisol as compared to all Pleistocene paleosols. The Holocene soil has a lower content of total and fine clay, a higher Kd index, and at the same time demonstrates a lower mineralogical and clay mineralogical weathering stage. The presented results show that the Kd index proposed by PÉCSI & RICHTER (1996), fits quite well to the mineralogical weathering stages used in this paper. It is remarkable that among Pleistocene paleosols the most weathered according to grain size characteristics

Tab. 3: Distribution of silt and clay fractions in the Oberlaab loess-paleosol sequence; mass%, WP = pedosediments.

Tab. 3: Verteilung der Schluff und Tonfraktionen in der Löss-Paläoboden-Sequenz Oberlaab, Masse%, WP = Bodensedimente.

IAG Lab. Nr	Horizon	Position	Silt	Clay	Fine clay
Profile OL1 Holocene Luvisol					
12401	AfD	0-5 cm	67.1	28.5	9.4
12402	EAh	5-15 cm	68.9	26.7	10.5
11212	E1	15-30 cm	71.7	26.1	7.3
11213	E2	30-40 cm	72.2	25.7	6.9
11214	EBt	45-65 cm	66.2	31.4	10.0
11215	EBtg	75-95 cm	67.7	29.6	3.8
11216	Btg1	100-120 cm	70.4	27.0	9.6
11217	Btg2	140-150 cm	69.2	28.2	10.8
11218	G	150-160 cm	70.8	26.1	4.7
11219	I	150-165 cm	70.7	27.1	9.4
11220	II	170-190 cm	61.5	29.3	11.8
Profile OL2 Eemian Luvisol					
11221	WP upper	0-20 cm	71.6	24.3	5.3
11222	WP upper	20-40 cm	68.1	28.3	9.1
11223	WP lower	40-50 cm	72.0	24.1	5.0
11224	WP lower	70-90	66.3	27.1	7.1
11225	Btg2	120-140 cm	45.6	51.3	31.7
11226/27	BCtg	160-180 cm	58.3	39.2	17.7
11228	BCtg	180-190 cm	64.1	32.8	14.8
11229	clay cutans		53.0	43.7	18.3
Profile OL3 Middle Pleistocene Paleosol					
11230	BCtg	5-20 cm	57.9	38.4	17.8
11233	Cg	25-40 cm	58.4	37.7	16.3
11234	Eg1	40-50 cm	64.0	28.7	9.7
11235	Eg2	70-90 cm	66.7	29.6	3.7
11236	Btg1	120-140 cm	66.0	30.5	10.5
11231	Btg2	180-190 cm	70.1	24.2	6.3
11232	BCg	210-220 cm	70.5	26.3	6.2
11238	clay cutans		60.7	37.1	13.6
Profile OL4 Middle Pleistocene Paleosol					
11432	BCg	10-30 cm	74.0	22.7	6.0
11433	Bg	60-70	62.6	34.4	10.3
11434	Bg	85-100 cm	60.6	37.2	13.8
11435	Btg1	120-140 cm	63.6	34.0	15.8
11436	Btg2	140-150 cm	63.9	33.8	13.6
11437	BCtg	180-200 cm	58.7	39.1	19.7
11438	BCtg	230-240 cm	61.2	36.4	16.3
11439	EB	270-280 cm	55.1	39.3	20.2
11440	G	290-300 cm	55.8	40.1	16.6
11441	clay cutans		33.6	64.3	54.6
Profile OL5 Middle Pleistocene Paleosol					
11442	BG	0-50	51.1	37.4	14.7
11443	G	50-90 cm	46.1	43.0	18.8
11444	Bg	90-170 cm	43.0	39.4	14.7
11445	BCg	170-215 cm	45.7	43.0	21.3
11446	Ah	215-230 cm	38.8	54.4	27.0
11447	EBg	230-250 cm	59.0	33.8	10.1
11448	Bg	250-260 cm	48.1	46.9	13.5
11449	Bg	265-290 cm	41.9	47.2	15.5
11450	Cg	290-310 cm	34.6	39.8	15.1
11451	C	310+	30.9	25.8	8.1



Tab. 4: Mineral composition of the bulk minerals in the Oberlaab loess-paleosol sequence, WP = pedosediments.

Tab. 4: Mineralogische Zusammensetzung der Löss-Paläoboden-Sequenz Oberlaab, WP = Bodensedimente.

IAG Lab. Nr		14Å	Mica	Chlorite	Layer sili-cates	Quartz	Feldspar	Hornbl	Calc + Dolo	Weath. stage
Profile OL1 Holocene Luvisol										
12401	AIO	.	*	*	*	**	*	.		2
12402	EAh	.	*	*	*	**	*	.		2
11212	E1	.	.	*	*	**	*	.		2
11213	E2	.	.	*	*	**	*	.		2
11214	EBt	*	*	*	*	**	*	.		2
11215	EBtg	*	*	**	*	**	*	.		2
11216	Btg1	*	*	*	*	**	*	.		2
11217	Btg2	*	*	*	*	**	*	.		2
11218	G	*	*	*	*	**	*	.		2
11219	I	*	*	*	*	**	*	.		2
11220	II	*	*	*	*	**	*	.		2
Profile OL2 Eemian Luvisol										
11221	WP upper	*	*	*	*	**	*	.		2
11222	WP upper	*	*	.	*	**	*	.		2
11223	WP lower	*	*	*	*	**	*	.		2
11224	WP lower	*	*	*	*	**	*	.		2
11225	Btg2	*	*	.	**	**	*			2
11226	BCtg	*	*	.	**	**	*			2
11228	BCtg	**	*	.	**	**	*			2
11229	clay cut.	*	**	.	**	**	*	.		2
Profile OL3 Middle Pleistocene Paleosol										
11230	BCtg	*	*	.	*	**	*	.		2
11233	Cg	**	*	.	*	**	*			2
11234	Eg1	*	*		*	**	*			3
11235	Eg2	*	*		*	**	*			3
11236	Btg1	**	*		*	**	*			3
11231	Btg2	*	*		*	**	*			3
11232	BCg	*	*		*	**	*			3
11238	clay.cut	.	.		**	**	*			3
Profile OL4 Middle Pleistocene Paleosol										
11432	BCg	.	*		*	**	*			3
11433	Bg	*	*	.	*	**	*			2
11434	Bg	*	*	.	*	**	*			2
11435	Btg1	*	*	.	*	**	*			2
11436	Btg2	**	*	.	*	**	*			2
11437	BCtg	**	*	.	*	**	*			2
11438	BCtg	**	*	.	*	**	*			2
11439	EB	*	*	.	*	**	*			2
11440	G	*	*		*	**	*			3
11441	Clay cut.	*	*		**	*	*			3
Profile OL5 Middle Pleistocene Paleosol										
11442	BG	*	*		**	**	*			3
11443	G	*	.		**	**	*			3
11444	Bg	*	.		**	**	*			3
11445	BCg	*	.		**	**	*			3
11446	Ah	*	*		**	**	*			3
11447	EBg	*	*		**	**	*			3
11448	Bg	*	*		**	**	*			3
11449	Bg	*	*		**	**	*			3
11450	Cg	*	*		**	**	*			3
11451	C	*	.		**	**	*			3

Legend:

\*\*\* : mineral in large quantity                      \* : mineral in low quantity  
 \*\* : mineral in intermediate quantity            . : mineral in traces

Tab. 5: Mineral composition of the clay fraction from the Oberlaab loess-paleosol sequence; well and poorly crystallized kaolinite are together in one column, WP = pedosediments.

Tab. 5: Mineralogische Zusammensetzung der Tonfraktion in der Löss-Paläoboden-Sequenz Oberlaab, WP = Bodensedimente.

IAG Lab. Nr	Horizon	Smectite	Vermiculite 18Å	Vermiculite 14Å	Illite	Kaolinite	Chlorite	Mixed Layer		Weath. stage
								I/Vc	I/Chl	
Profile OL1 Holocene Luvisol										
12401	Ai0				*	*	*	*		1
12402	EAh				*	*	**	*		1
11212	E1	*	*		*		**	*		1
11213	E2	*			*	*	**	*		1
11214	EBt	*			*	*	*	*		1
11215	EBtg	*			*	*	*	*		1
11216	Btg1	*			*	*	.	*	.	1
11217	Btg2	*	*		*	*		*	*	2
11218	G	*			*	*		*	*	2
11219	I	*			*	*	.	*	*	1
11220	II	*			*	*		*	.	2
Profile OL2 Eemian Luvisol										
11221	WP upper	*	*		*	*		*	*	2
11222	WP upper	*	*		*	*		*	.	2
11223	WP lower	**			*	*		*	.	2
11224	WP lower	**			*	*		*	.	2
11225	Btg2	***			*	*		*	.	3
11226	BCtg	***			*			*	.	3
11228	BCtg	***			*	.		*	.	3
11229	clay cut	**			*	*		.	.	2
Profile OL3 Middle Pleistocene Paleosol										
11230	BCtg	*	*	*	*	*		*	.	3
11233	Cg	*	*	*	*	*		*		3
11234	Eg1	*	*	*	*	*		*		3
11235	Eg2	*	*	*	*	*		*	.	3
11236	Btg1	*			*	*		*		3
11231	Btg2	*		*	*	*		*		3
11232	BCg	*	**		*	*		*		3
11238	clay.cut	*	*	*	*	*		*		3
Profile OL4 Middle Pleistocene Paleosol										
11432	BCg	*	**	*	*	*		.		3
11433	Bg	*	**		*	.		.		3
11434	Bg	*	**		*	.		.		3
11435	Btg1	*	**	*	*	*		.		3
11436	Btg2	*	**	*	*	.		.		3
11437	BCtg	*	**	*	*	*		.		3
11438	BCtg	**	**	*	*	*		.		3
11439	EB	*	**	.	*	*		.		3
11440	G	*	**	*	*	*		.		3
11441	Clay cut.		**	*	*	.		.	.	3
Profile OL5 Middle Pleistocene Paleosol										
11442	BG	**	**		*	*		.	.	4
11443	G	**	**		*	.		.		4
11444	Bg	*	**		.	.		.		4
11445	BCg	*	**		*	.		.		4
11446	Ah	*	**		*	.		.		4
11447	EBg	*	**	*	*			.		3
11448	Bg	**	**	*	*			.	.	3
11449	Bg	**	**	*	*	.		.		3
11450	Cg	**	**	*	*	.		.		3
11451	C	**	**	*	*			.	.	3

Legend:

\*\*\* : mineral in large quantity      \* : mineral in low quantity  
 \*\* : mineral in intermediate quantity      . : mineral in traces

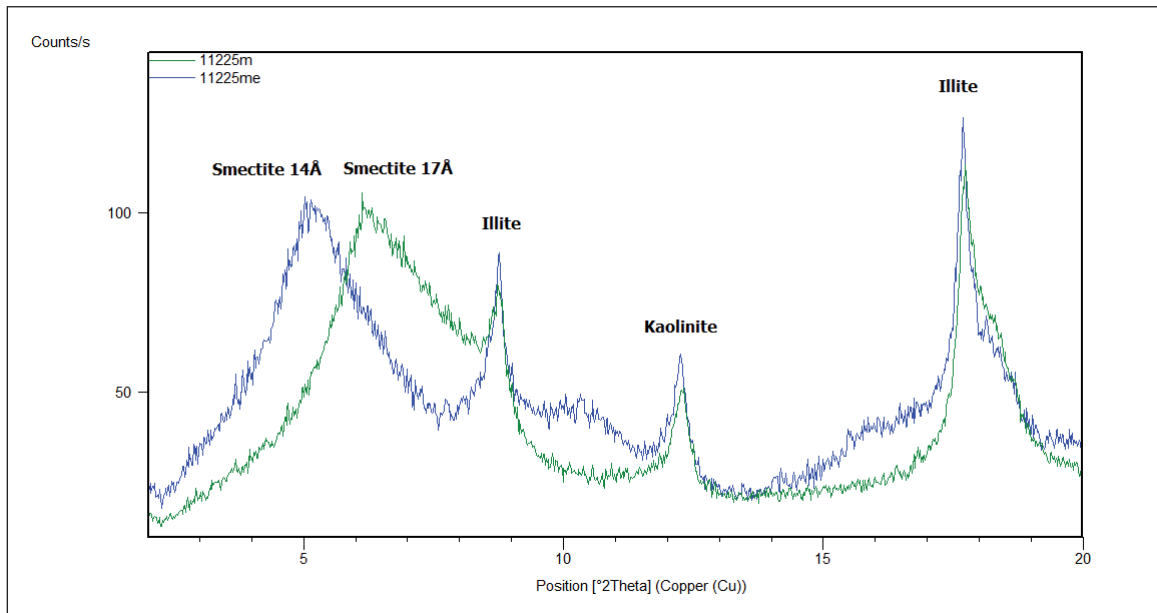


Fig. 4: X-ray diffractogram of the clay fraction <math><2\mu\text{m}</math> from OL2 Btg2 horizon. The green line is Mg treated and air dried, the blue line is Mg-treated and glycolated. The expansion of smectite from 14Å to 17Å is quite visible.

Abb. 4: Diffraktogramm der Tonfraktion <math><2\mu\text{m}</math> aus dem Btg2 Horizont von OL2. Die grüne Kurve zeigt die Mg-behandelte und lufttrockene Tonfraktion, die blaue Kurve die Mg-behandelte und glykolisierte Kurve. Deutlich ist die Aufweitung von Smektit von 14Å auf 17Å zu sehen.

are the clay illuvial Bt horizons of the Eemian paleosol OL2 and the basal gleyed paleosol OL5. We suppose that some strong chemical weathering processes should be taken into account when interpreting these maxima (CHAMLEY, 1989; SHELDON & TABOR, 2009). In the Eemian argic horizons the maximum of illuvial clay was registered by micromorphometric studies (SOLLEIRO-REBOLLEDO et al., 2013). Furthermore, detailed micromorphological observations revealed abundant multiphase clay illuvial pedofeatures (SEDOV et al., 2013, this volume). We suppose that besides weathering, strong clay illuviation could contribute to the maximum of fine fractions observed in these horizons. This is confirmed by the abundance of fine clay and smectites, which are the most mobile components in suspensions. The lower paleosol unit OL5 represents a quite different soil environment. Both abundant morphological features of redoximorphic processes and minimal values of magnetic susceptibility point to water-logged anoxic conditions. Usually such conditions hamper silicate weathering. We should pay attention to the changes in the composition of parent material in this unit – higher sand and lower silt content indicate an increase of the fluvial component. A part of the clay could be provided by slow fluvial sedimentation (overbank alluvium) being derived from the older pre-existing weathered materials of the Alpine Foreland.

In the Oberlaab sequence, grain size and mineralogical indicators of weathering provide more variable and sensitive signals of mineral transformations than the geochemical index CIA, presented by SOLLEIRO-REBOLLEDO et al. (2013).

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# Last Interglacial paleosols with Argic horizons in Upper Austria and Central Russia – pedogenetic and paleoenvironmental inferences from comparison with the Holocene analogues

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## Abstract:

The paleosols of the Last Interglacial are presented in many loess sequences of the European temperate zone by soils with Argic horizon, that are considered to be the pedological response to the bioclimatic conditions of that period. We studied micromorphological, physical/chemical (bulk chemical composition, texture and dithionite-extractable iron) and mineralogical characteristics of two profiles – an Eemian Luvisol in Upper Austria (Oberlaab) and a Mikulino Albeluvisol in Central Russia (Alexandrov Quarry near Kursk) to compare them with recent analogous soils and to make further paleoecological and chronological inferences. Both profiles showed a set of characteristics indicative for weathering of primary minerals, clay transformation illuviation and surface redoximorphic (stagnic) processes. Paleosols demonstrate more advanced development than the Holocene analogues manifested however in different pedogenetic characteristics. The Eemian Luvisol in Upper Austria is characterized by stronger clay illuviation manifested in higher clay content and abundance of illuvial clay pedofeatures in the Bt horizon. Mikulino Albeluvisol in Central Russia is more strongly affected by eluvial and stagnic processes evidenced by deeper and more intensive accumulation of bleached silty material and clay depletion. We suppose that the properties of parent material are responsible for these differences. Russian Albeluvisol is formed on the Dnepr loess poor in weatherable minerals and having limited capacity for buffering acidity and clay formation. The higher development status of the Last Interglacial paleosols compared to the Holocene soils having however same type pedogenesis implies longer soil formation period, that agree with some of the paleobotanical proxies and could include besides MIS 5e part of MIS 5d; the warmer and moister paleoclimate during MIS 5e could also account for more advanced paleosol development. Several phases of clay illuviation interrupted by frost structuring and deformation are detected in the Eemian Bt horizon in Upper Austria. It suppose even longer development that could extend to the Early Würmian interstadials (late substages of MIS5).

## Letztinterglaziale Paläoböden mit Bt-Horizonten in Oberösterreich und Zentral-Russland: pedogenetische und paläoumweltbezogene Schlussfolgerungen im Vergleich mit den holozänen Analogon

## Kurzfassung:

In vielen Löss-Paläoboden-Sequenzen der gemäßigten Breiten Europas ist das letzte Interglazial (dt.: Eem, russ.: Mikulino) durch einen Bt-Horizont (Argic horizon nach WRB) repräsentiert, der als pedologisches Resultat entsprechender bioklimatischer Bedingungen dieser Zeit gedeutet wird. Wir untersuchten mikromorphologische, physikalische/chemische (Gesamtelementzusammensetzung, Korngrößenverteilung und dithionitlösliches Eisen) und mineralogische Charakteristika im Profilabschnitt des Eem-Luvisols in Oberösterreich (Oberlaab) und des Mikulino-Albeluvisols in Zentralrussland (Alexandrov Grube nahe Kursk). Aus dem Vergleich der Paläoböden mit entsprechenden rezenten Böden ergeben sich paläoökologische und chronologische Schlussfolgerungen. Beide Profile zeigen eine Reihe von Charakteristika, die auf Verwitterung primärer Minerale und Tonminerale sowie Tonverlagerung und redoximorphe Prozesse hinweisen. Die Paläoböden zeigen jeweils eine weiter fortgeschrittene Entwicklung im Vergleich zu entsprechenden holozänen Böden, jedoch anhand unterschiedlicher pedogenetischer Merkmale. Der Eem-Luvisol in Oberösterreich weist ein höheres Maß an Tonverlagerung auf, was durch einen höheren Tongehalt und zahlreiche Toncutane im Bt-Horizont gezeigt wird. Der Mikulino-Albeluvisol in Zentralrussland ist stärker von Auswaschung und Stauwasser betroffen, was sich in tieferer und stärkerer Ansammlung von gebleichtem schluffigem Material und Tonverarmung zeigt. Wir nehmen an, dass das Ausgangsmaterial diese unterschiedliche Entwicklung hervorruft. Der russische Albeluvisol bildete sich auf Dnjepr-Löss, welcher im Vergleich zu den Riss-Lössen Oberösterreichs ärmer an leicht verwitterbaren Mineralen ist und daher eine geringere Kapazität Säure zu puffern sowie Ton neu zu bilden aufweist. Das fortgeschrittenere Entwicklungsstadium des letztinterglazialen Bodens im Vergleich zu holozänen Böden, die allgemein demselben Bodentyp entsprechen, spricht für eine längere Bodenentwicklungsphase, was mit paläobotanischen Ergebnissen in Einklang steht. Einerseits könnte der letztinterglaziale Paläoboden neben dem MIS 5e auch Teile des MIS 5d umfassen, andererseits könnte eine intensivere Paläobodenentwicklung durch das wärmere und feuchtere Paläoklima während der Interglazialphase des MIS 5e verursacht werden. Mehrere Phasen der Tonverlagerung, unterbrochen durch frostdynamische Strukturierung und Deformation sind im eemzeitlichen Bt-Horizont in Oberösterreich nachweisbar. Die noch weiter reichende Entwicklung könnte bis in noch jüngere Frühwürm-Interstadiale gereicht haben.

## Keywords:

*Last Interglacial, paleosol, Argic horizon, paleoclimate*

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Fig. 1: Geographical situation of the studied profiles. O – Oberlaab, A – Alexandrov.

Abb. 1: Geographische Lage der untersuchten Profile. O – Oberlaab, A – Alexandrov.

## 1 Introduction

The perspective of global climate change increased the interest to the paleoecology of earlier warmer periods, in particular to the Quaternary interglacials. The last interglacial (known in the regional stratigraphies as Riss-Würm, Eemian, Mikulino, Sangamonian) drew special attention as far as the corresponding geological objects and materials are frequent and well preserved. During the last decades a significant progress was achieved in defining the duration of this warm phase and its paleoenvironmental characteristics based on the correlations between marine and terrestrial records (SHACKLETON et al. 2003; SIER et al. 2011).

Paleosols of loess-paleosol sequences are among the most important terrestrial records of interglacial environments, especially since these sequences were thoroughly correlated with the deep sea isotope record (KUKLA 1977; BRONGER et al. 1998). In the contemporary temperate zone of Europe the paleosols of the last interglacial are often represented by soil bodies with strong clay illuviation (HAESAERTS & MESTDAGH 2000), which were considered to be, with certain limitations, the most significant response of pedogenesis to the interglacial bioclimatic conditions (STEPHAN 2000). The examples are the Ipswichian paleoargillic soils in England (CATT 1996), Rocourt Luvisol in France (ANTOINE et al. 1999) and Belgium (HAESAERTS et al. 1999; VANDENHAUTE et al. 2003; VANCAMPENHOUT et al. 2008), Erbacher Boden – Luvisol/Alfisol/Parabraunerde in Germany (MAHANAY et al. 1993; ANTOINE et al. 2001; TERHORST et al. 2001), lower Luvisol of the PK3 pedocomplex in Czechia (DEMEK & KUKLA 1969; FRECHEN et al. 1999) and Nietulisko I - GJ1 pedocomplex in Poland (JARY 2009, 2011), “brown-podzolic soil” of Pryluky-Kaydaky pedocomplex in Ukraine (ROUSSEAU et al. 2001; GERASIMENKO 2006) and Salyn paleosol, the Luvisol of Mezin pedocomplex in Central Russia (VELICHKO 1990; YAKIMENKO 1995; VELICHKO et al. 2006).

Only in areas with drier continental climate the last interglacial paleosols do not show clay illuviation: in Lower Aus-

tria (FINK 1956; BRONGER 1976; PETICZKA 2009; SPRAFKE et al. 2013, this volume) it is represented by Cambisols, whereas in the south-eastern part of the East European Plain (RUTTER et al. 2003) and the Carpatian Basin (BRONGER 1976; MARKOVIC et al. 2008) the last interglacial soils correspond to Chernozems. In fact, in the European temperate zone Luvisols occupied a larger areal in the Eemian soil cover than they do in the Holocene.

This area differs especially in the East European Plain where the zone of forest soils was much broader during the Eemian and occupied a large portion of the present day steppe zone (MOROZOVA 1981; MOROZOVA et al. 1998). In a number of sites in Central Russian and Ukrainian steppe regions one can see a striking difference between Holocene Chernozem on the surface and buried Mikulino Luvisol, separated by Valday loess, all exposed in the same cut (this is also the case in Alexandrov quarry exposure).

In the present day temperate humid forest areas both, last interglacial and Holocene soils show clay illuviation. However, they are often clearly distinguished quantitatively: stronger development of the last interglacial Argic horizons compared to the recent ones was reported in a number of case studies (BOARDMANN 1985; BULLOCK & THOMPSON 1985; MAHANAY et al. 1993). The reasons for this distinction are not so obvious: differences in bioclimatic conditions as well as in duration of pedogenesis could be involved.

We have to conclude that both paleoecological and chronological aspects of the last interglacial Luvisols development are far from being clear. Concerning the paleoenvironmental setting the assumption that the Bt horizon is a reliable sign of interglacial conditions “defined by the return of the temperate deciduous woods” (STEPHAN 2000) could be too ambitious: we know that in the present day soil cover these horizons occur in a much broader spectrum of ecosystems. In the classic zonal sequence of the Eastern Europe soils with strong clay illuviation are formed on loamy substrates from the broadleaf forests and forest-steppe till boreal coniferous northern taiga (GEROSIMOVA et al. 1996). In fact within the Eurasian forest belt only in the regions with drier extra-continental climate and permafrost in Middle and Eastern Siberia clay illuviation is hampered (GEROSIMOVA et al. 1996).

Broad ecological limits of the recent Argic horizon formation have implications for determining its timing in Europe. In the course of soil evolution during and after Last Late Glacial, clay illuviation is detected not only in the Holocene but also in the late MIS 2 interstadials (KÜHN 2003). Some authors attribute the development of Argic horizons in the European Luvisols mostly to the late glacial (BØLLING) boreal paleoenvironment (VAN VLIET LANOË 1990).

Thus the possibility should be considered that also in Mikulino/Eemian paleosols the Luvisol pedogenesis occurred not only during the interglacial *sensu stricto* but also in transitional phases of colder forest or forest-steppe ecosystems before and after it. In fact, this scenario has been already proposed by various researchers who detected several clay illuviation phases in the Eemian to Early Würmian pedocomplexes in Belgium (HAESAERTS et al. 1999; HAESAERTS & MESTDAGH 2000) and Germany (ANTOINE et al. 2001; TERHORST et al. 2001). All these authors coincided in discriminating between forest Luvisol development corresponding to the warm humid interglacial and forest-steppe

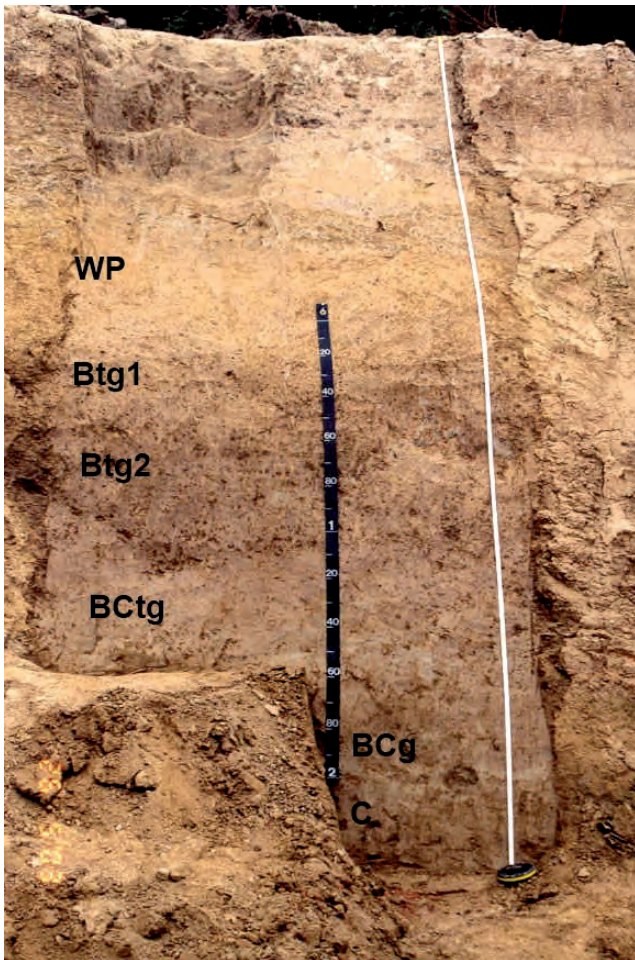


Fig. 2: Profile morphology of the Eemian paleosol of the Oberlaab profile, with the overlying Würmian pedosediment (WP).

Abb. 2: Eemzeitlicher Paläoboden im Profil Oberlaab mit darüber liegendem würmzeitlichem Bodensediment.

Grey Forest Soils attributed to cooler continental climate of the Early Weichselian interstadials. This approach revives the earlier ideas of BOARDMANN (1985) about longer pedogenesis of the Pleistocene interglacial paleosols compared to the Holocene.

However, in most works on loess stratigraphy, chronology and paleoecology the information on paleosol pedogenesis is rather limited: the conclusions are mostly based on field morphological observations, grain size and magnetic susceptibility distribution. Deeper understanding of paleoenvironmental significance of the paleosol of the last interglacial in Europe requires more detailed data about the types and succession of pedogenetic processes – which can be provided by detailed micromorphological and clay mineralogical investigation accompanied by soil chemical analyses (BRONGER 1976; BRONGER & HEINKELE 1990; KEMP 1999).

Such detailed study of the buried Luvisols promise to be effective also concerning their comparison with modern surface soils. Temperate forest soils with clay illuviation are provided with one of the most variable and detailed datasets of morphological, physical/chemical and mineralogical characteristics. In fact, these soils were one of the main “experimental polygons” of pedology to construct scenarios of pedogenesis (including polygenetic models) based on these complex datasets (e.g. TARGULIAN et al. 1974 a,b; KÜHN et

al. 2006). Thus, the actualistic approach towards paleosol interpretation was expected to be most fruitful in case of the Luvisols.

This research intends to contribute to the knowledge about the last interglacial soil evolution in Europe by carrying out a comparative pedogenetic analysis of two buried paleosols with Argic horizon formed on loess in geographically distant and ecologically dissimilar regions: Upper Austria (Oberlaab section) and Central European Russia (Alexandrov quarry section near Kursk). The final objective is to develop scenarios of paleosol genesis and evolution for both profiles and from this derive conclusions about their paleoenvironmental significance and timing.

## 2 Materials and methods

Two profiles of the last interglacial paleosols from the loess-paleosol sequences were subjected to detailed morphological analysis and sampling being selected as the key sites for the ICSU-2003 project “Polygenetic models for Pleistocene paleosols” (Fig. 1). The profile in Oberlaab (Upper Austria) is located on the fluvio-glacial terrace attributed to the Mindel glaciation, close to its bench (TERHORST 2013, this volume). The present day climate is humid providing conditions for broad leaf forest vegetation and for a development of Stagnic Luvisols as dominant modern soil type. The profile Alexandrov quarry (near Kursk, Central Russian Plain) is located at the watershed position close to the slope towards the valley of river Mlodat. The modern climate is temperate semi-humid, the surface soils are Haplic Chernozems formed under meadow-steppe vegetation. Thus, both study sites present two major cases of interrelation between last interglacial Luvisols and the Holocene soil: they belong to the same type in case of Oberlaab and they are genetically dissimilar in case of Alexandrov section.

The stratigraphic, chronological, and paleoecological aspects of both paleosol-sedimentary sequences were published earlier. For Oberlaab these data are available in the papers included in this Special Volume and some earlier publications (TERHORST 2007, 2013, this volume; SOLLEIRO et al. 2013, this volume). The Eemian soil in Oberlaab is the uppermost of 5 buried paleosols; it is separated from the modern soil by colluvial and loessic sediments with signs of weak pedogenesis, attributed to the Würmian glacial. Alexandrov section was studied by Sycheva and co-authors during the last 20 years, the results are presented in several papers (c.f. SYCHEVA 1998; GOLYEVA & SYCHEVA 2010; SYCHEVA & SEDOV 2012). Mikulino (Salyn) paleosol is developed on the Dnieper (Rissian Glacial, MIS 6) calcareous loess. The important advantage of Alexandrov section is an extensive and detailed part corresponding to the last (Valday) glaciation, overlying the Mikulino soil. The lower segment of this part is comprised by three Early Valday paleosols clearly separated by colluvial strata. Above them the Middle Valday paleosol is developed (Bryansk soil). Finally, modern Chernozem is separated from Bryansk soil by the Late Valday loess.

Detailed morphological description of the last interglacial paleosol profiles was followed by sampling of the genetic horizons: bulk samples for physical, chemical and mineralogical characteristics and blocks with undisturbed structure for thin sections were collected. Micromorpho-

logical observations were performed under petrographic microscope and the descriptions followed the terminology of BULLOCK et al. (1986). We gave special attention to the pedofeatures related to the clay eluvial-illuvial redistribution, redoximorphic processes as well as the signs of aggregation and mixing by biotic and cryogenic processes. The properties selected for paleosol laboratory analyses were those indicative of ancient pedogenetic processes and not subjected to major diagenetic changes after burial. Bulk chemical composition was estimated by X-ray fluorescence analysis. Grain size distribution – intended to estimate clay accumulation due to weathering and illuviation – included separation of sand fractions by sieving and quantification of silt and clay by pipette method. The content of free iron oxides – an important product of soil weathering – was evaluated by measuring Fe in the dithionite-citrate-bicarbonate extract ( $Fe_d$ ). Clay mineral assemblages in the soils with eluvial-illuvial differentiation are known to be specific for different horizons and strongly influenced by pedogenetic processes (TARGULIAN et al. 1974b; TONKONOGOV et al. 1987). We studied the composition of clay fraction by X-ray diffraction using a Shimadzu XRD-6000 diffractometer in oriented specimen subjected to different pretreatments: air-dry, saturated with ethylene-glycol, heated to 375 °C and 500 °C. Clay species content was estimated in semi-quantitatively, using simple peak weighting factors; for area estimation we used Fityk (WOJDYR 2010) program. Illite, smectite and mixed-layered components were identified by separating their peak areas in the glycolated XRD-trace and calculating their proportions (MOORE & REYNOLDS 1997).

All results were further integrated with other datasets – morphological, geochemical, rock magnetic, etc. published for Oberlaab by SOLLEIRO et al. (2013, this volume), for Kursk by SYCHEVA (1998), RIVAS et al. (2003) and SYCHEVA & SEDOV (2012).

The key element of interpretation consisted in the comparison of obtained results with the existing knowledge about the pedogenetic characteristics of the clay-illuvial soils formed under different environmental conditions: temperate forest of Western and Central Europe (JAMAGNE et al. 1984; KÜHN et al. 2006; SAUER et al. 2009), taiga (TARGULIAN et al. 1974 a,b; TONKONOGOV et al. 1987) and forest-steppe (MIEDEMA et al. 1999) ecosystems of European Russia. Reference profiles for comparison were modern analogues (surface soils) as close as possible to the respective studied profile for which the necessary pedogenetic characteristics were available: surface Holocene Luvisol for Oberlaab (Solleiro et al. 2013, this volume) and modern Albeluvisol of the Moscow region (TARGULIAN et al. 1974 a,b).

### 3 Results

#### *Paleosol morphology from macro- to microscale*

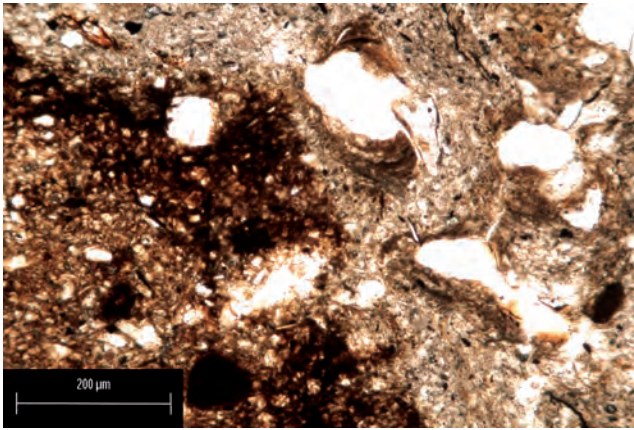
The Eemian soil in Oberlaab is located at a depth of about 3m below the Holocene Luvisol, formed on Würmian sediments. The lower part (60 cm) of the latter directly above Eemian Bt horizon has yellow color, silty loam texture and contains ferruginous concretions. This morphology together with gradual wavy boundary suggests that this layer could be a pedosediment that incorporated material of the Eemian eluvial horizon (indicated in the tables and

figures as WP – Würmian pedosediment). The underlying 90 cm thick brown Btg horizon (subdivided into Btg1 and Btg2) is darker, more compact and clayey than the overlying pedosediment, with platy/subangular blocky structure. Brown clay coatings cover ped surfaces and fill vertical cracks. Furthermore, few pale coatings of bleached silt, iron concretions and mottles are present. Clay illuvial features decrease in the underlying BCtg and BCg horizons (joint thickness about 1m), where larger visible cutans are restricted to very few biogenic pores – channels and chambers. Simultaneously, stagnant colour pattern appears – pale areas, often of subhorizontal lenticular shape are neighboring with yellow mottles of ferruginous pigment. Below follows the pale brown-yellow homogeneous C horizon – the loessic sediment of Riss glaciations separating the Eemian profile from the underlying Middle Pleistocene paleosol (TERHORST 2007; SOLLEIRO et al. 2013, this volume). The soil was classified in the field as Stagnic Cutanic Luvisol (Fig. 2). Under the microscope, the groundmass in all horizons is dominated by coarse silt, having rich and variable mineralogical composition: besides quartz feldspars, amphiboles and micas are abundant. Micas are represented both by colorless muscovite and dark brown biotite; in the latter at stronger magnification signs of weathering are observed – weak interference colors and pleochroism, deformation, precipitation of iron oxides.

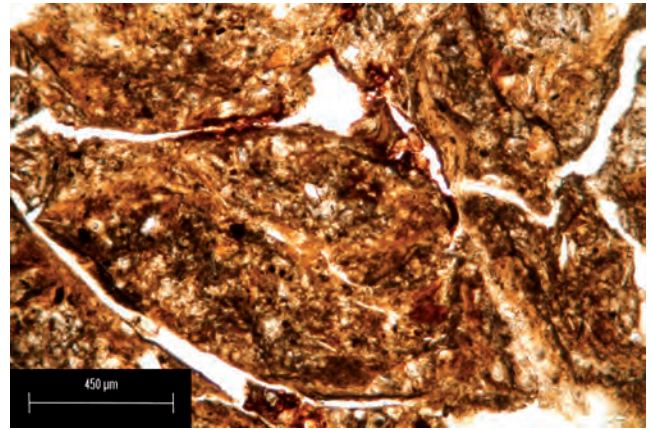
The Würmian  $\gamma$ 0pedosediment (WP) has areas with banded fabric demonstrating stripes of coarse silt and fine material. Frequent ferruginous nodules – rounded, circular, compound are distributed within the groundmass, few thin clay coatings are present. Areas with banded structure were also observed in the Btg1 horizon. Clay illuvial pedofeatures become more abundant in this horizon, being presented both by undisturbed clay coatings (mostly thin and impure) (Fig. 3a) in the pores and fragmented and deformed clusters of illuvial clay incorporated into groundmass. Ferruginous nodules and mottles, mostly with irregular shape and diffuse boundaries are also abundant. Rather big anorthic iron-clay nodules of subangular fragmented shape (Fig. 3b) and charcoal fragments with the tissue structure of a conifer tree were encountered.

The major abundance and variety of clay illuvial materials was found in the Btg2 horizon. Undisturbed clay coatings are present, however, they are not dominant. Fragments of clay illuvial pedofeatures of different size and morphology are more frequent, some of them are larger than any of the undisturbed coatings (Fig. 3c). The Btg2 horizon is characterized by strong fracturing giving rise to angular blocky and coarse lenticular peds (Fig. 3d) and being accompanied by displacement along the cracks. The walls of these cracks support thin undisturbed coatings. Bleached irregular shaped silt concentrations depleted of clay are very rare (Fig. 3e). Few infillings with bow-like structure are of zoogenic origin (Fig. 3f) (KOOISTRA & PULLEMAN 2010).

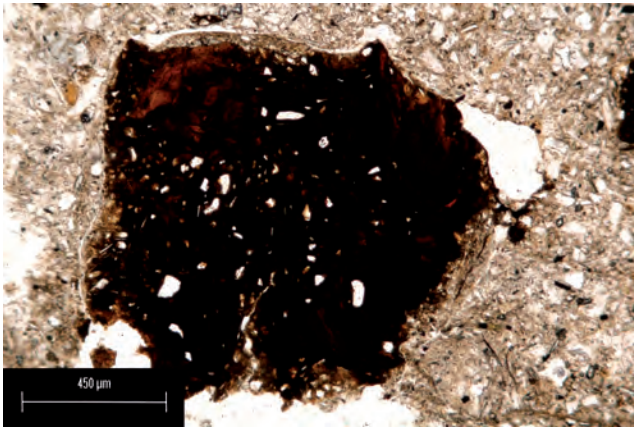
Deeper in the BCtg horizon the number of clay illuvial pedofeatures decreases but their preservation increases – they are mostly undisturbed coatings in voids, some of them large and laminated (alternating limpid with strong interference colors and impure microlayers) (Fig. 3g,h). Further down the profile in the BCg horizon clay coatings are



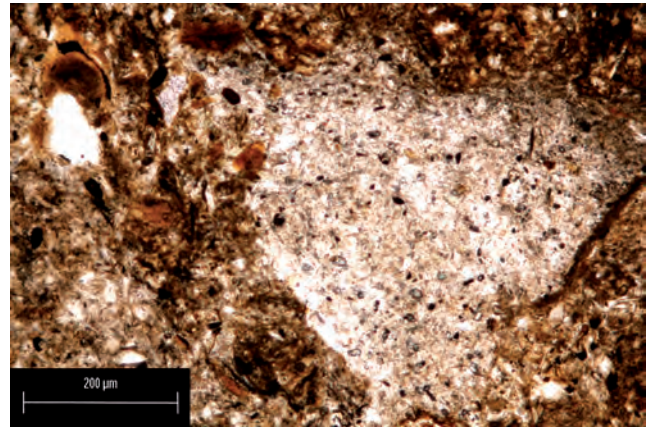
a) Impure clay coatings, Btg1 horizon, plain polarized light.



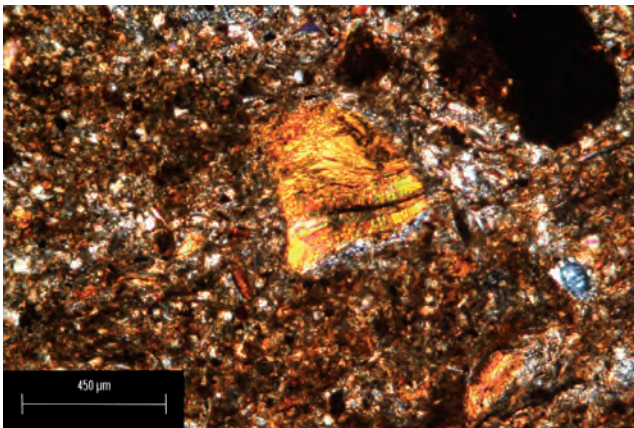
d) Angular blocky – lenticular structure produced by fracturing; note thin clay coatings on the aggregate surface. Btg2 horizon, plain polarized light.



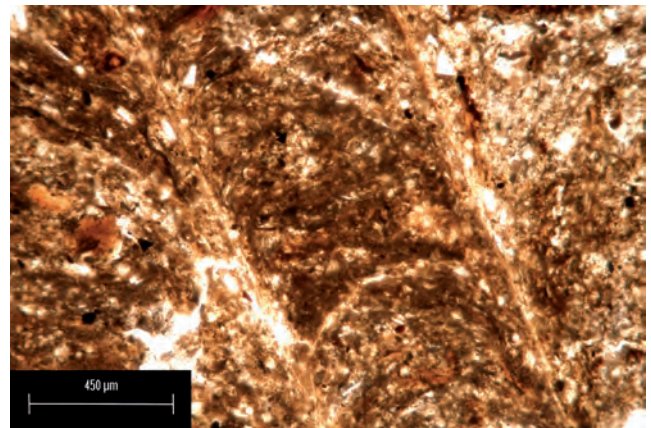
b) Anorthic ferruginous nodule, Btg1 horizon, plain polarized light.



e) Concentration of bleached silt particles. Btg2 horizon, plain polarized light.



c) Fragment of clay illuvial pedofeature, note strong interference colors and microlamination. Btg2 horizon, crossed polarizers.

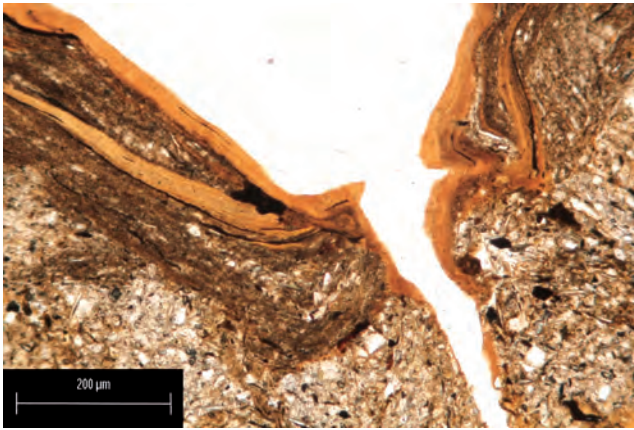


f) Infilling with bow-shaped structure. Btg2 horizon, plain polarized light.

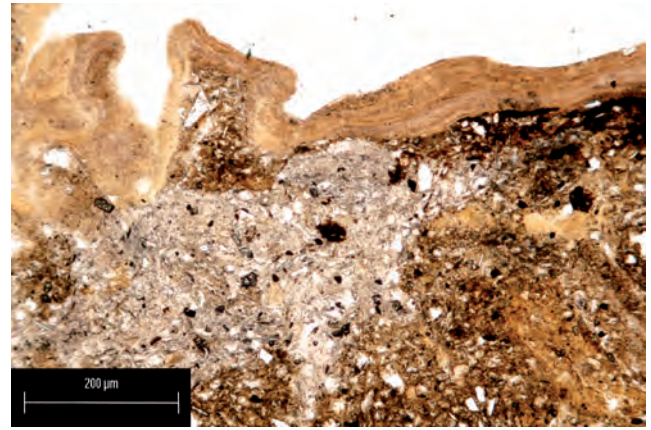
Fig. 3: Micromorphology of the Eemian paleosol in Oberlaab.

Abb. 3: Mikromorphologie des eemzeitlichen Paläobodens in Oberlaab.

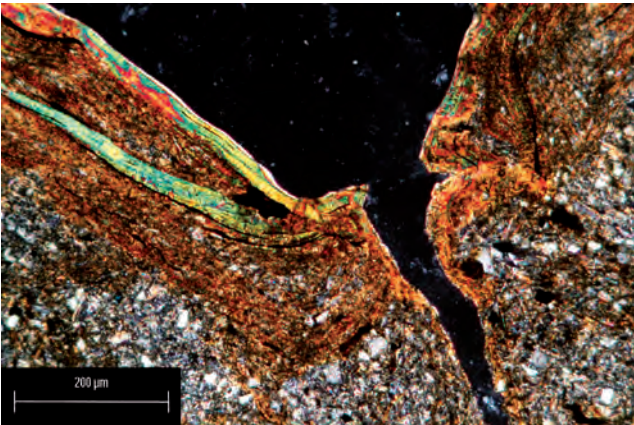




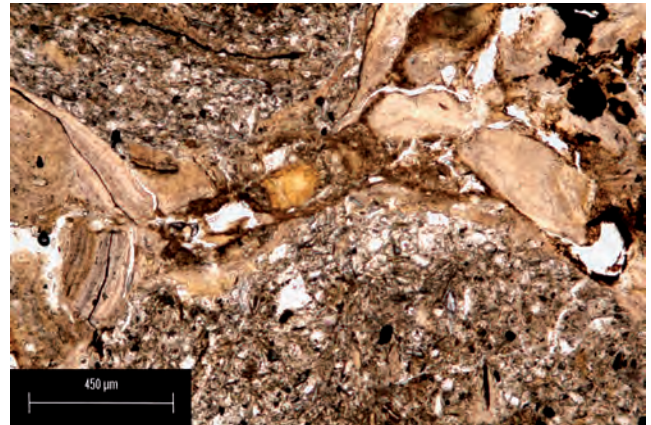
*g) Laminated clay coatings with limpid and impure microlayers. BCtg horizon, plain polarized light.*



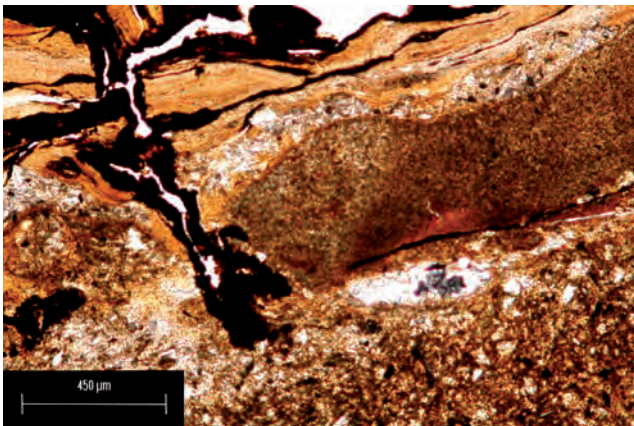
*j) Bleached area in the groundmass under clay coating. BCg horizon, plain polarized light.*



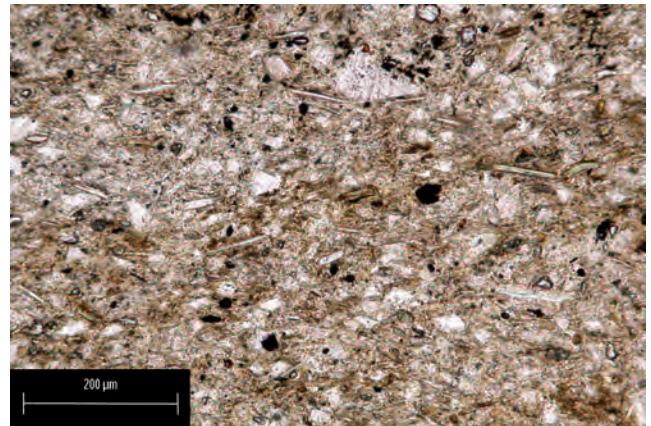
*h) Same as g), crossed polarizers; note high birefringence of limpid layers.*



*k) Disturbed clay illuvial pedofeature, fragments covered by a thin undisturbed clay coating. BCg horizon, plain polarized light.*



*i) Clay illuvial pedofeature with ferruginous coatings on the surface and in the cracks. BCg horizon, plain polarized light.*



*l) Parallel orientation of elongated mineral particles, C horizon, plain polarized light.*

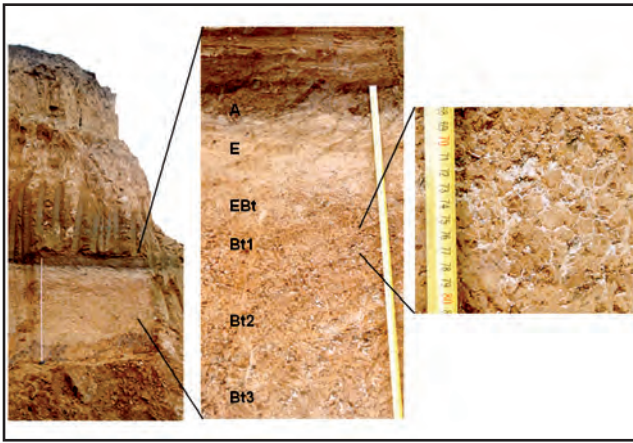


Fig. 4: Profile morphology of the Mikulino paleosol in Alexandrov: position in the sequence, eluvial-illuvial differentiation, detail of the Bt1 morphology, showing bleached areas related to the pores.

Abb. 4: Morphologie des Mikulino Paläobodens in Alexandrov: Lage im Profil, Differenzierung von Eluvial- und Illuvialbereichen, Detailaufnahme Bt1-Horizont mit gebleichten Bereichen, die im Zusammenhang mit Poren stehen.

even fewer and usually combined with ferruginous nodules and films penetrating into illuvial clay (Fig. 3i). Frequently, clay coatings in the pores overly bleached microareas adjacent to the pore walls, however, these areas although lacking ferruginous pigment are not depleted of fine clay material (Fig. 3j). In the BC horizons few clay pedofeatures are fragmented but usually not incorporated into groundmass – the fragments stay in the pores being only slightly displaced and covered by the new generation of thinner cutans (Fig. 3k). Finally in the C horizon all pedofeatures become less abundant and primary sedimentary structures are more evident – among them parallel subhorizontal orientation of the elongated mineral particles (Fig. 3l).

The profile of Mikulino paleosol as mentioned above is overlain by 6 m of Walday sediments with a Holocene Chernozem on top. The profile demonstrates a remarkable preservation of all horizons including the topsoil. It is delimited from the overlying laminated colluvial sediments by a thin (less than 1 cm) black layer of burned litter. Below it a grey-brown Ah horizon (8 cm thick) is located; it contains numerous charcoal particles (having tendency to parallel orientation) and few soft ferruginous nodules, sometimes associated with the charcoal. It is underlain by the very pale 20 cm thick E horizon with well-developed multi-order platy structure. Iron oxide nodules (soft and hard) become more abundant, whereas charcoal particles are fewer. An important diagnostic feature is the type of the lower boundary of this horizon. It is gradual with the bleached material penetrating into the underlying layer along pores and fissures and it forms thin glossic fills in the vertical cracks – known as “albeluvic tonguing”. The downward “invasion” of bleached silt material defines the 15 cm thick transitional EB horizon. It still has platy structure, however, platy aggregates are thicker and break into small subangular blocks. The bleached material concentrates on the aggregate surface whereas the internal part is darker and contains more clay. This distribution produces a heterogeneous color pattern: the network of pale veins is spread among the brown “islands”. Iron-manganese pedofeatures

are presented by small nodules and films mostly associated with the thin fissures.

The illuvial part of the profile (Bt1, 30 cm thick – Bt2, 35 cm thick – Bt3, 65 cm thick) is marked by a contrasting difference of main morphological characteristics. It has clayey loamy texture, subangular blocky to prismatic structure and predominantly brown color. However, silty bleached material is still abundant: in the Bt1 horizon it forms films on most ped surfaces. In many cases these films overly brown clay coatings. In Bt2 the pale silt component decreases and is restricted mostly to the larger vertical cracks. Aggregates are covered with continuous clay coatings. The prismatic blocky peds become coarser and clay coatings on their surfaces turn patchy in the Bt3 horizon. All Argic horizons have also ferruginous pedofeatures, in Bt1 these are mostly small nodules, whereas in Bt2 and Bt3 dendritic iron-manganese spots occur.

The BC horizons have brighter brown-yellow color and weaker pedogenic structure. Where the paleosol lies on Dnieper loess we observed a clear banded pattern in these horizons: darker clayey stripes were alternating with the pale-yellow silty bands.

All these horizons are free of carbonates and the latter appear only in the underlying loess, at the depth of 250 cm from the Mikulino surface. The paleosol was defined as Haplic Albeluvisol (Fig. 4).

Microscopic observations reveal predominance of silt grains in the groundmass. Their mineralogical composition is rather poor – quartz is clearly dominant, K-Na feldspars are present in minor quantities, micas and heavy minerals are few.

The A horizon has complex microstructure with combination of platy and isometric crumbly aggregates. Dark organic pigment with heterogeneous distribution causes interchanging of dark-colored and light-colored micro-zones. There are few signs of faunal activity: some zoogenic channels with excrement infillings were found; groundmass contains numerous charcoal particles (Fig. 5a).

In the E horizon dominant fine platy aggregates demonstrate clear textural differentiation with prevailing bleached coarse silt in the upper parts and close to the pores and fine silt and plasmic material in the central and lower parts. Frequent compact rounded simple iron nodules are widespread in the groundmass, sometimes charcoal is observed in their nucleus (Fig. 5b). We encountered fragments of clay coatings embedded in the bleached silty matrix (Fig. 5c,d).

The EB horizon also has well developed platy structure with even stronger intra-pedal differentiation of particle size. Fine silt and clay is concentrated inside the plates, while the peripheral parts consist of bleached silt. Iron pedofeatures are frequent and variable: we observed rounded nodules and incrustation of platy aggregates mainly on the bottom side (fig. 5e).

Traces of platy aggregation are still observed in the Bt1 horizon but plates are frequently neighbouring subangular blocks (Fig. 5f). The pores have common bleached silty coatings as well as infillings, whereas deformed and fragmented illuviated clay pedofeatures were found in the interior of the peds (Fig. 5g). Undisturbed clay illuvial pedofeatures still related to the pores were not observed. Small ferruginous nodules are found both, in the groundmass and silty infillings.

Horizon	Total element concentrations, %							
	SiO <sub>2</sub>	TiO <sub>2</sub>	Al <sub>2</sub> O <sub>3</sub>	Fe <sub>2</sub> O <sub>3</sub>	MgO	CaO	Na <sub>2</sub> O	K <sub>2</sub> O
Eemian paleosol [Oberlaab, Austria]								
WP upper	74.14	1.12	14.86	4.95	1.03	0.81	0.97	1.94
WP lower	74.46	1.12	14.75	4.68	1.01	0.85	1.00	1.92
Btg1	76.16	1.03	13.15	4.66	1.09	0.66	0.94	2.11
Btg2	70.26	0.94	16.49	6.38	1.60	0.67	0.79	2.60
BCtg	70.58	0.95	16.02	6.30	1.63	0.81	0.88	2.58
BCg	71.37	0.94	15.33	6.01	1.59	0.91	1.04	2.56
C	71.48	0.96	15.47	5.86	1.55	0.93	1.03	2.45
Mikulino paleosol [Alexandrov quarry, Russia]								
A	82.30	0.90	9.44	2.68	0.61	0.81	0.65	2.34
E upper	85.37	0.90	7.69	1.88	0.43	0.62	0.56	2.38
E lower	83.72	0.94	8.54	2.33	0.55	0.58	0.64	2.52
EBt	81.97	0.95	9.30	3.14	0.71	0.60	0.58	2.62
Bt1	81.30	0.92	9.70	3.40	0.73	0.62	0.55	2.63
Bt2	81.01	0.81	10.23	3.44	0.78	0.64	0.53	2.37
Bt3	80.67	0.79	10.49	3.45	0.76	0.67	0.59	2.39
BC	82.17	0.77	9.60	3.02	0.64	0.68	0.49	2.43

WP – Würmian pedosediment

Only the Bt2 and Bt3 horizons have undisturbed layered illuvial clay coatings on the pore walls (Fig. 5h). However, bleached silt concentrations are still present and concentrated mainly in fissures and large channels. They have abrupt boundaries to the groundmass, and sometimes they are juxtaposed on the clay coatings or occur in the neighboring pores (Fig. 5i,j).

In the thin sections of the banded BC horizons we observed a clear laminar structure with alternating bands enriched in coarse silt and fine clay materials. Even in these horizons we observed few thin clay coatings, which showed no clear preference for specific bands.

#### Physical and chemical characteristics

Bulk chemical composition demonstrates similar tendencies of profile differentiation in terms of element concentrations. The upper eluvial horizons have higher quantities of Si at the expense of Al and also Fe and Mg. However, it should be noted that absolute values are quite different in the two studied profiles: the material of Mikulino paleosol in Alexandrov has much higher Si content (80–85%) than that of the Eemian soil in Oberlaab (70–76%), whereas the latter is more rich in Al, Fe as well as alkaline and alkaline earths (except K that has similar concentrations in both paleosols) (Tab. 1).

Clay content shows a clear and similar distribution trend in both profiles: the maximum is observed in the Bt horizons, whereas lower values are found in the eluvial E horizon in Alexandrov and upper Würmian pedosediment in Oberlaab (in the latter – with contrasting fluctuations between upper and lower parts). The highest values of the dithionite extractable iron also correspond to the illuvial part of both profiles. However, iron and clay maxima do not coincide: iron content is highest in the upper Bt1 horizons,

whereas clay has its maximum below in the Bt2 horizons. Again, both clay and Fe<sub>d</sub> contents although having very similar distribution tendencies in both profiles, demonstrate different absolute values. For both components they are considerably higher in the Oberlaab profile (Tab. 2).

Clay mineral assemblages show very clear differentiation in both studied profiles. Interlayered illite-smectite with variable proportion of both components – (1.4–1.5 nm peak, shifting to smaller angles on glycolation) decreases strongly in the eluvial E horizons and is much higher in the illuvial stratum and parent material. In the Eemian Luvisol of Oberlaab this tendency is rather weak and expressed mostly in different proportions of smectite-rich and smectite-poor mixed layered components. The latter increase in the upper part of the profile. In the Mikulino Albeluvisol of Alexandrov this tendency is much stronger: all smectitic components decrease in the eluvial part, whereas on the contrary, illite and kaolinite (respectively 1.0 nm and 0.7 nm peak, the latter disappearing after 550 °C heating) are more abundant (see Table 3 and Fig. 6). Only in the pedosediments, eluvial and upper illuvial horizons (EB and Bt1) we found mixed-layered minerals with chloritic component (“shoulder” of 1.0 nm peak towards smaller angles after heating indicating incomplete contraction of 1.4–1.5 nm maximum), which we identify with the hydroxyl-interlayered vermiculite and smectite (HIV and HIS) (BAMHISEL & BERTSCH 1989).

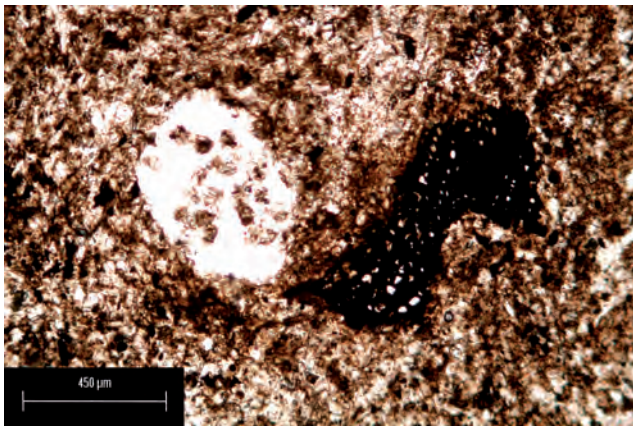
## 4 Discussion

### *Pedogenesis of paleosols – comparison with the regional modern analogues*

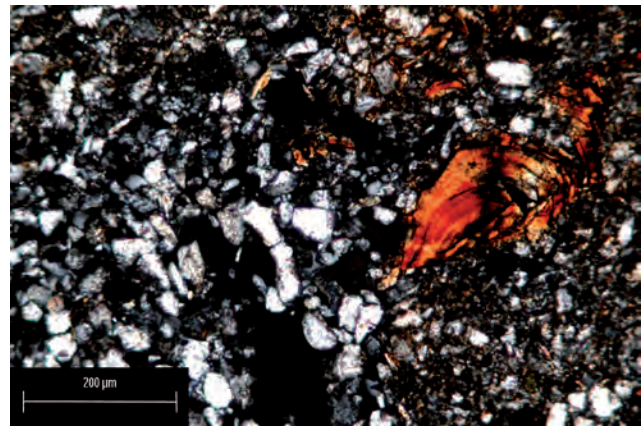
The question to be answered before soil forming processes in the paleosol can be assessed is how complete is their profile

Tab. 1: Bulk chemical composition: selected major elements.

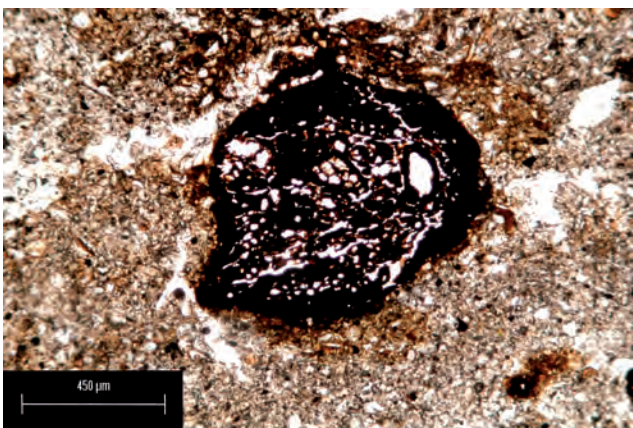
Tab. 1: Ergebnisse der Elementaranalysen: Ausgewählte Hauptelemente.



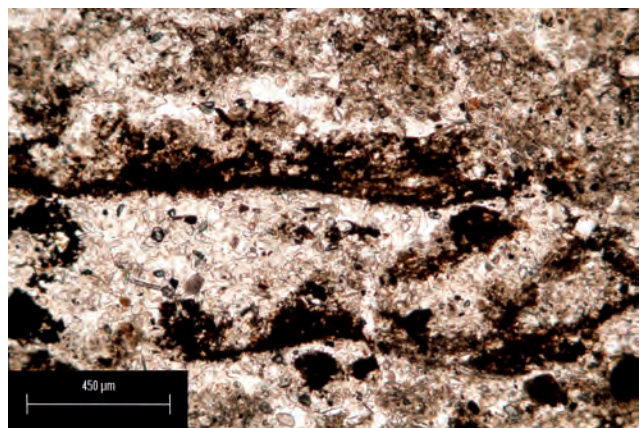
a) Pore with the excremental infilling, charcoal particle. A horizon, plain polarized light.



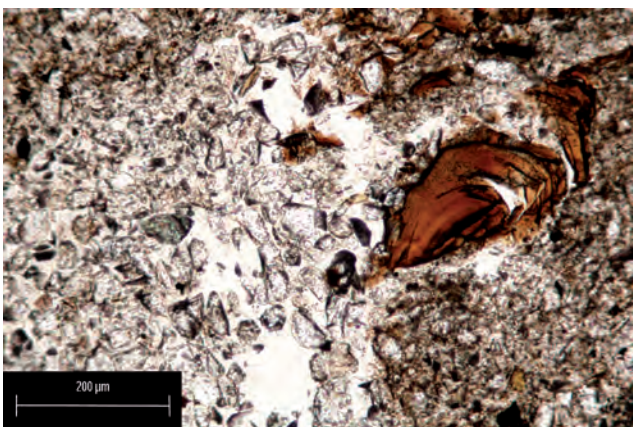
d) Same as c), crossed polarizers; note strong interference colors and undulating pattern of extinction.



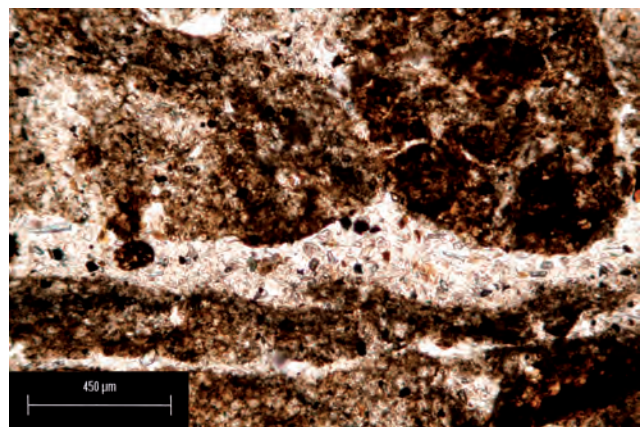
b) Ferruginous nodule with charcoal in the center. E horizon, plain polarized light.



e) Platy structure, fissures between the plates are filled with bleached silt, ferruginous hypocoatings are located on the lower surfaces of the plates. EBt horizon, plain polarized light.

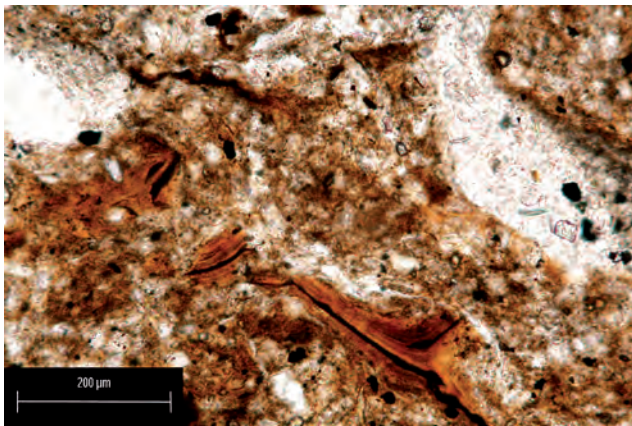


c) Fragment of clay coating surrounded by bleached silt. E horizon, plain polarized light.

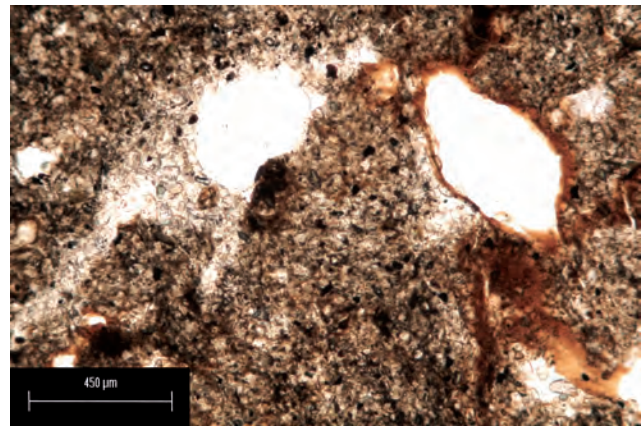


f) Platy and blocky aggregates, pores are filled with bleached silt, intrapetal matrix enriched with fine material. Bt1 horizon, plain polarized light.

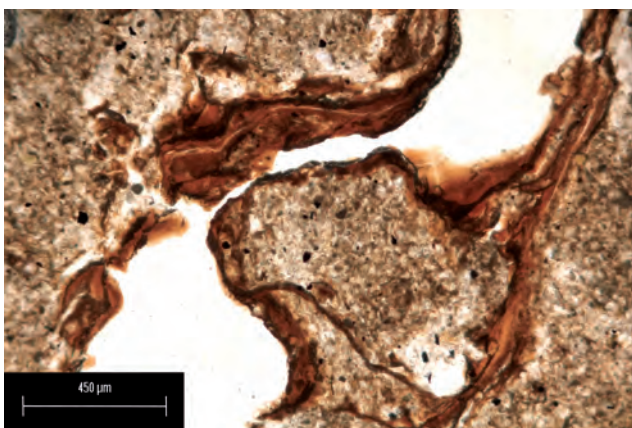
Fig. 5: Micromorphology of the Mikulino paleosol in Alexandrov.  
Abb. 5: Mikromorphologie des Mikulino Paläobodens in Alexandrov.



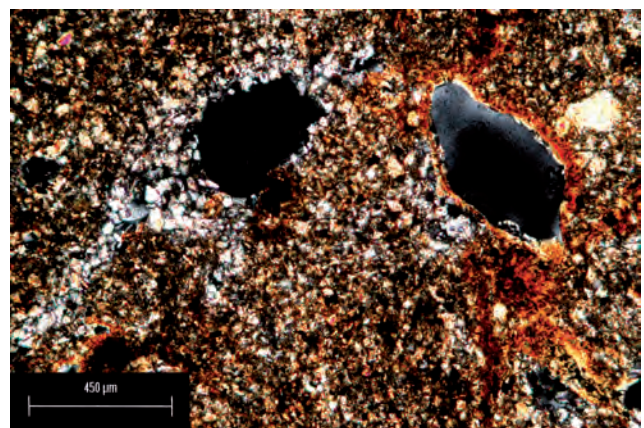
g) Fragmented and deformed clay illuvial pedofeatures incorporated into groundmass. Bt1 horizon, plain polarized light.



i) Bleached silt (left) and limpid clay coating (right) in the neighboring pores. Bt3 horizon, plain polarized light.



h) Continuous undisturbed clay coating in the pore. Bt3 horizon, plain polarized light.



j) Same as i), crossed polarizers, note strong interference colors of the clay coating.

and to what extent it is affected by post-burial (diagenetic) processes.

In the Oberlaab profile the layer directly above the Eemian Bt horizon has clear signs of sedimentation and mixing. These signs are stronger in the upper part of this horizon, where grain size fluctuations mark a lithological discontinuity; in this layer few thin clay coatings are most probably the lowermost traces of clay illuviation from the Holocene Luvisol. However, stagnic ferruginous pedofeatures and presence of the specific clay component – hydroxyl interlayered vermiculites and smectites suggest that material of the eluvial horizon, although redeposited is present in this layer. Minor signs of disturbance (anorthic iron-clay nodules, charcoal) are present even in the Bt1 horizon, and the paleosol is supposed to be intact only below the Bt1 (TERHORST et al. 2003).

In case of Alexandrov quarry the grade of preservation is high: a complete sequence of genetic horizons including the burned litter is present. Only the A horizon has signs of redeposition: oriented charcoal particles. Moreover, no effects of posterior pedogenesis are detected in this paleosol, due to its protection by a thick overlying sediment layer.

In both profiles a similar set of pedogenetic processes is detected. Clay illuviation is evidenced by clay coatings (observed both on macro- and microscale) as well as a maximum of clay content in the Bt horizon compared both to the underlying C and overlying E horizon (in situ or redepos-

ited). Clear micromorphological signs of silicate weathering are observed in the biotite particles. Increase of dithionite extractable iron, which mostly includes fine oxides produced by weathering is an important analytical indicator. An enhanced Si content in the eluvial horizons of Luvisols is often interpreted as a result of decay of all unstable silicates (including part of clay minerals) and residual accumulation of quartz (TARGULIAN et al. 1974b). Loss of smectites in the same horizons is also considered to be a result of selective destruction of the most unstable clay components (TONKONOGOV et al. 1987). However, it should be taken into account that clay illuviation could also contribute to these properties. Removal of clay from E horizon also could cause an increase of Si, due to relative accumulation of coarse components richer in this element. Smectites are most fine and dispersible components, that could be preferentially mobilized by suspension transport and removed from the eluvial horizons (VAN RANST et al. 1982). Thus, geochemical and mineralogical indicators give a mixed signal from both processes, a fact that unfortunately is rarely taken into account when elemental composition of loess-paleosol sequences is interpreted through various “weathering indices”.

More clear evidence of clay mineral transformation is provided by the HIV accumulation in the upper part of the paleosols. This component is known to be formed from 2:1 clay minerals with labile structure due to synthesis of

Tab. 2: Grain size distribution and dithionite extractable iron content.

Tab. 2: Korngrößenzusammensetzung und dithionitlöslicher Eisengehalt.

Horizon	Grain size fractions, %			Fe <sub>d</sub> , %
	sand	silt	clay	
Eemian paleosol [Oberlaab, Austria]				
WP upper	5.4	68.5	26.1	0.69
WP lower	8.2	69.8	22.0	0.65
Btg1	10.1	66.7	23.2	0.98
Btg2	6.3	55.9	37.8	0.73
BCtg	4.6	60.9	34.5	0.61
BCg	4.3	64.4	31.3	0.37
C	5.7	65.4	28.9	0.50
Mikulino paleosol [Alexandrov quarry, Russia]				
A	1.4	81.8	16.8	0.49
E upper	1.4	88.7	9.9	0.22
E lower	1.2	86.5	12.3	0.32
EBt	1.1	79.3	19.6	0.43
Bt1	1.0	78.2	20.8	0.51
Bt2	1.2	70.3	28.5	0.45
Bt3	1.5	76.1	22.4	0.43
BC	1.2	80.0	18.8	0.37

WP – Würmian pedosediment

fragmentary Al-hydroxyl layer between the 2:1 stacks. This process takes place in acid soil environment and Al for it is provided by hydrolytic alteration of silicates – primary, or the same clay components (BAMHISEL & BERTSCH 1989). Presence of these components in the eluvial horizons (in situ in Alexandrov and redeposited in Oberlaab) point to acid weathering in both interglacial paleosols. However, the alternative explanation of higher content of chloritic component in the upper horizons of acid forest soils on loess also exists. JAMAGNE et al. (1984) suppose that these minerals are not formed through the synthesis, but rather by simple breakdown and partial degradation of primary chlorites from silt fractions.

Surface redoximorphic (stagnic) processes produced ferruginous pedofeatures throughout the profile, but most abundant and variable in the eluvial and upper Bt horizons. There they are presented by nodules as well as by coatings and incrustations, indicative of iron migration (BRONNIKOVA et al. 2000). Besides intra-horizontal short-distance redistribution we suppose certain eluvial-illuvial transport of iron from E to EB and upper Bt horizons due to this process. Dithionite extractable iron is depleted in E horizons and has its maximum in the Bt1; also the abundance of the ferruginous pedofeatures is maximal in the EB and Bt1 horizons. Interestingly magnetic susceptibility also demonstrates maxima in EB (Alexandrov quarry) (RIVAS et al. 2003) and Bt1 (Oberlaab) (SOLLEIRO et al. 2013, this volume) that could also be related to the increase of pedogenic iron oxides. It should be taken into account that surface redoximorphic processes could also contribute to acid weathering and HIV neof ormation through the process of ferrollysis (BRINKMAN 1970).

Being similar regarding the set of the pedogenetic processes the studied paleosols differ greatly in their relative development. The Austrian paleosol demonstrates much stronger clay illuviation in the Bt horizon, whereas the Russian profile has much stronger evidences of eluvial and surface redoximorphic processes. This tendency becomes even clearer by comparison with the modern soils. The Eemian soil in Oberlaab has higher clay content and more abundant clay illuvial pedofeatures than the overlying Holocene Luvisol (SOLLEIRO et al. 2013, this volume). The Mikulino Albeluvisol in Alexandrov quarry compared to the local Holocene soil, a Haplic Chernozem, shows contrasting qualitative differences in the set of key pedogenetic processes. The Chernozem is characterized by high accumulation of dark humus and development of coprogenic granular structure resulting in the development of Mollic topsoil horizon. Below, in the Bk horizon abundant neof ormated carbonates are observed. Morphological features of clay translocation are very few (if any), clay content is rather uniform throughout the profile showing no tendency towards eluvial-illuvial differentiation (GERASIMOVA et al. 1996; BRONGER 2003). This set of pedogenetic properties typical for steppe pedogenesis under Ustic soil moisture regime (BRONGER 2003) provides undoubted evidence that on the local scale the Holocene climate has been drier than that of the last interglacial – in agreement with the above mentioned reconstruction for Eastern Europe by MOROZOVA (MOROZOVA 1981; MOROZOVA et al. 1998). For current research it is interesting to compare the studied Mikulino profile with the nearest analogue having similar profile organization, to assess changes in the Albeluvisol pedogenesis on the regional scale. The Mikulino Albeluvisol in Alexandrov quarry compared to the Holocene Albeluvisol of the Moscow region (TARGULIAN et al. 1974b), does not show stronger clay accumulation, which is contrary to the case study in Low Austria. However, it demonstrates major advance of eluvial and stagnic processes – bleaching and clay depletion besides E and EB horizons affected the whole Bt horizon, “attacking” the area of earlier clay illuviation. Fresh illuvial pedofeatures pointing to active clay illuviation appear only in the lower Bt2 and BtC horizons. In profiles of Holocene Albeluvisols (KÜHN et al. 2006) and Grey Forest Soils (MIEDEMA et al. 1999) this “aggression” of eluviation against degrading illuvial areas is also observed but usually is not so deep. It is strong in EBt horizon and decreases in the upper part of Bt horizons, where it is represented mostly by deep thin bleached tongues (albeluvisol tonguing), whereas the areas between the tongues are much less affected (SPIRDONOVA et al. 1999; SAUER et al. 2009). Thus, in both cases last interglacial paleosols present higher grade of development than analogous Holocene profiles in the region, however, the particular mode of this tendency is different.

These differences could be explained when linked to the classic evolution scheme for forest soil formation under temperate humid climate (PÉDRO et al. 1978). Following this scheme intensive clay illuviation develops after initial leaching phase and results in the formation of Argic horizon. However, it is succeeded by the subsequent “degradation” stage of further acidification, surface redoximorphic processes and (possibly) ferrollysis. At this stage clay illuviation is hampered and part of accumulated clay is lost due to acid weathering. We suppose that Eemian Luvisol in Oberlaab

Horizons	Types of clay minerals				
	Smectite/ illite	Illite / smectite	Chloritised 2:1 minerals	Illite	Kaolinite
Eemian paleosol [Oberlaab, Austria]					
WP upper	-	xxx	x	xxx	xx
WP lower	-	xxxx	xx	x	x
Btg 1	-	xxxx	xx	x	x
BCtg	xxxx	xx	-	x	x
BCg	xxxxx	x	-	xx	x
C	xxxxx	x	-	xx	x
Mikulino paleosol [Alexandrov quarry, Russia]					
E upper	-	xxx	xx	x	xx
E lower	-	xx	xx	xx	xxx
EBt	-	xx	xx	xx	xx
Bt2	xxxx	xx	-	x	x
C	xxxxx	-	-	x	x

Tab. 3: Clay mineral composition.

Tab. 3: Tonmineralogische Zusammensetzung.

WP – Würmian pedosediment  
 xxxxx – dominant  
 xxxx – abundant  
 xxx – frequent  
 xx – few  
 x – traces  
 - - not detected

still developed within the stage of predominant clay formation and illuviation, whereas Mikulino Albeluvisol in Alexandrov quarry already reached the stage of “degradation”.

This difference in the evolutionary status is most probably related to the composition of the paleosol parent material. The Riss loess in Austria is richer in weatherable minerals (including phyllosilicates like biotite), has more clayey texture, is enriched in bases and has lower Si content. This increases the potential for neutralizing excessive acidity as well as for the neoformation of clay and iron oxides. The Dnieper loess is much poorer, strongly enriched in quartz, thus buffering and clay generating capacities are lower.

#### **Chronological and paleoecological background of the advanced Eemian/Mikulino paleosol development**

On the local scale, especially in the border regions, qualitative differences between Mikulino/Eemian paleosol and modern soil profile are clearly related to climatic shift as shown by the case of Alexandrov exposure. The question arises, what are the reasons why last interglacial paleosols demonstrate more advanced development than Holocene soils belonging, however, to the same type of pedogenesis (proven by similarity of horizonation, main macro- and micromorphological characteristics, clay mineral association etc.)? We suppose that in many cases the duration of pedogenesis rather than “strong interglacial paleoclimates relative to the present-day climate” (as supposed by MAHANEY et al. 1993) could account for these differences. Comparison with the European paleobotanical records with high-resolution chronological scale provides insight into this problem. Detailed correlation with the deep sea isotope records and establishing the precise position of Blake paleomagnetic event allowed reliable dating of the main paleobotanic phases of Eemian established earlier in the pollen spectra from terrestrial (mostly lacustrine) sediments (ZAGWIJN 1996; TURNER 2002a). Earlier the last interglacial was associated with the Marine Isotope Stage (MIS) 5e (SHAKLETON 1969). Recently, it was shown that the spread of forest vegetation in Europe postdated the onset of MIS 5e but on the other hand

extended into the subsequent stage 5d despite the beginning of global cooling, increase of ice volume, and drop of sea level (SHAKLETON et al. 2003; SIER et al. 2011). This extension concerns in particular the last “post-temperate” paleobotanical phase of boreal forests. Thus, although the duration of the last interglacial *sensu stricto* is supposed by many authors to be about 11 000 yr (similar to that of the Holocene) (CASPER et al. 2002; TURNER 2002b), the total succession of the Eemian forest vegetation phases in the Western and Central Europe could reach 20,000 yr- nearly twice longer than the Holocene (KUKLA et al. 2002).

It should be stressed that all phases of humid forest ecosystems including the last one of the boreal forests provide conditions for clay illuviation and development of Argic horizons on loess-like sediments. Thus, the scenario of long-term clay illuviation during the complete succession of different forest phases during MIS5 explains more advanced development of Eemian/Mikulino paleosols compared to the Holocene analogues. Rather cold climatic conditions during the last stages of Albeluvisol profile development in Central Russia are confirmed by micromorphological observations. Typical features of eluvial-illuvial processes are combined with deep frost effects in the upper part of the profile. Lenticular structure with the microscale grain size differentiation is related to ice lensing (VAN VLIET-LANOË 2010), iron incrustations of the platy aggregates are also characteristic for seasonally deeply frozen soils (FEDOROVA & YARILOVA 1972).

An even more complex scenario for interaction of illuvial and frost processes is required for the Eemian Luvisol in Oberlaab. Here, the illuvial and frost features not only co-exist but show complex interaction. Frost fracturing propagates into the Bt horizon disrupting the older generation of the clay illuvial pedofeatures and generating blocky-lenticular structure, whereas the younger generation of clay coatings covers these aggregates. The clay translocation process itself was heterogeneous as evidenced by variable composition of illuvial pedofeatures. Impure dusty and coarse clay coatings (“mud cutans”) are attributed to the

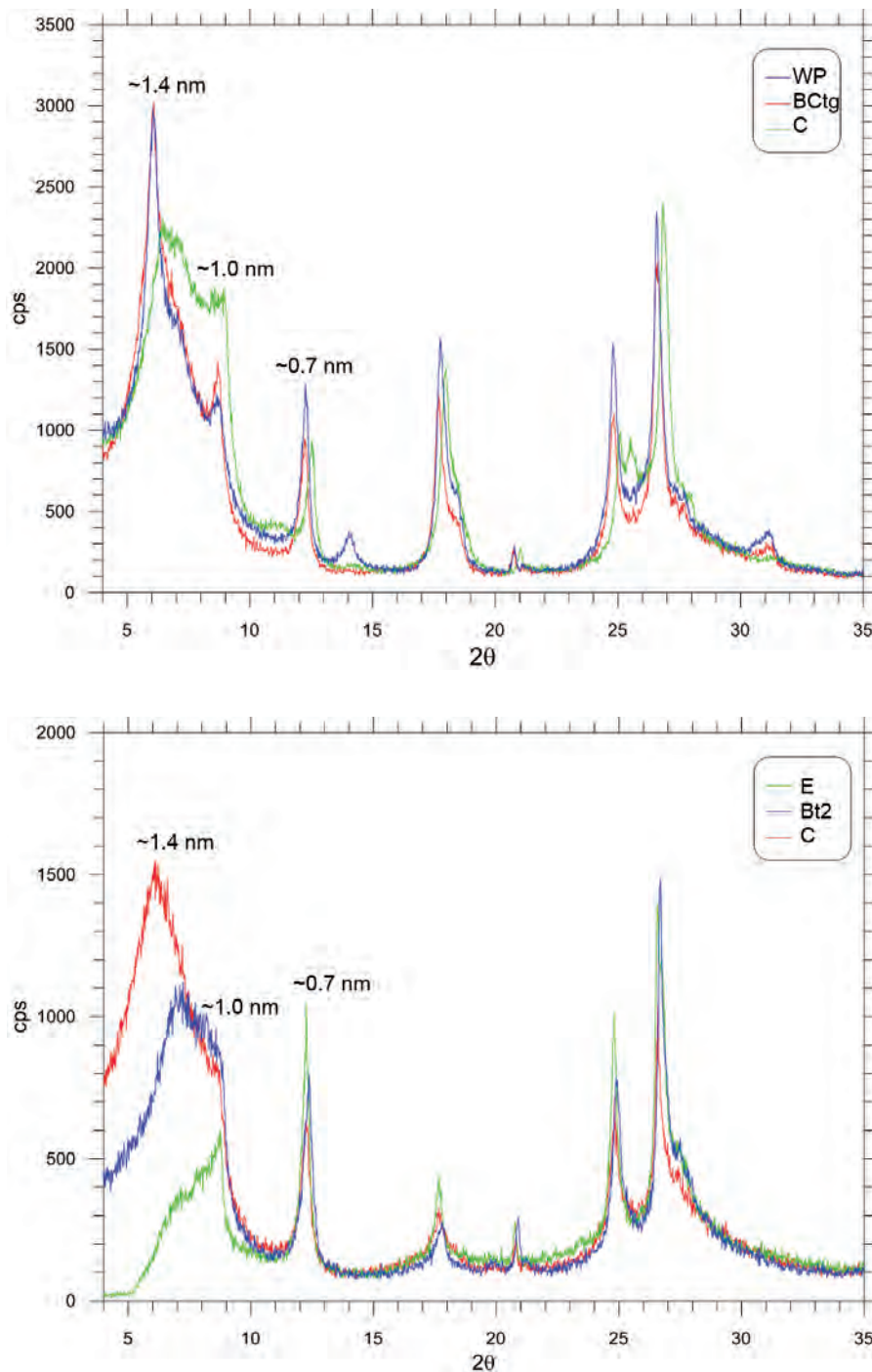


Fig. 6: X-ray diffraction patterns of oriented air-dry specimens of clay fractions from selected paleosol horizons; a) Oberlaab profile, b) Alexandrov profile.  
 Abb. 6: Röntgenbeugungsdiagramme von orientierten luftgetrockneten Proben der Tonfraktion aus ausgewählten Horizonten der Paläoböden; a) Profil Oberlaab, b) Profil Alexandrov.

fast percolation of suspensions after snowmelt in freeze-thaw soils (VAN VLIET-LANOË 2010). The laminated clay coatings (which we observed in the lower part of the Eemian Luvisol of Oberlaab) where pure limpid and impure silty microlayers are juxtaposed, could be interpreted as alternation of “normal” illuviation due to slow percolation and fast deposition from snow melt suspensions (KÜHN et al. 2010).

This suggests that clay illuviation in the Oberlaab profile was polygenetic and discontinuous, interrupted by the stages of deep frost processes. Such interplay of features is well known for Quaternary Bt horizons and could be used for pedostratigraphic inferences (summarized by VAN VLIET-LANOË 2010)

Concerning possible chronological and paleoecological implications we should take into account that no paleosols

or pedosediments clearly correspond to the Early Würmian interstadials. Thus, we could suppose that the pedogenesis of these warmer stages could contribute to the formation of the “Eemian” profile in Oberlaab. In several profiles of Europe the soils of these interstadials are characterized by dark humus accumulation and correspond to steppe environment (e.g. Mosbacher Humuszonen in Germany, Krutitsa soil in Russia, etc.). From such type of pedogenesis one would not expect contribution to the clay illuviation. However, this type of soil and landscape conditions seems to be not sole during the Early Würmian interstadials. Forest ecosystems have been also registered by various European paleoecological proxies (BIBUS et al. 2002). The warmer climate and reforestation could provide conditions for Luvisol pedogenesis, although interrupted by cold phases of the early glacial. Early Würmian Bt horizons clearly separated from the Eemian



Luvisol and correlative to Mosbacher Humuszonen were reported in the Rhine Area (LESER 1970). As mentioned above (see 'Introduction') several authors described superimposing of the Early Würmian clay illuviation (associated with Grey Forest Soil formation) over the pre-existing Eemian Luvisol producing a pedocomplex.

We suppose that Early Würmian clay illuviation could develop in the "Eemian" Argic horizon interrupted by frost processes imposed by the cold stadials in Oberlaab. Thus, its morphological and analytical (clay content, geochemical differentiation) characteristics could be an integral of longer Luvisol pedogenesis including Eemian Interglacial sensu lato (MIS 5e and part of MIS 5d) and Early Würmian interstadials (correlative to MIS 5c and MIS 5a). Such cumulative effect could be present in a number of the Eemian Luvisol profiles where Early glacial sedimentation was not sufficient to separate interglacial and interstadial soil bodies. In this way we return to the earlier ideas of BOARDMAN (1985) who stated: "Geological and palynological evidence from terrestrial, midlatitude sites suggests that temperate conditions conducive to soil formation existed for periods greatly in excess of 10,000 yr during the middle and late Pleistocene. Available data on soils formed during this interval from Western Europe and midwestern United States are best explained by relatively long interglaciations or the development of composite soils over a number of temperate periods".

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# Paudorf *locus typicus* (Lower Austria) revisited – The potential of the classic loess outcrop for Middle to Late Pleistocene landscape reconstructions

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## Abstract:

The more than 12 m thick loess-paleosol sequence in Paudorf, Lower Austria, has been known for decades as *locus typicus* of the “Paudorfer Bodenbildung” (Paudorf paleosol). The upper section of the outcrop contains an up to 1 m thick pedocomplex that developed during MIS 5. The differentiated sequence of loess-like sediment below, including a more than 2 m thick pedocomplex in its basal part, is an exceptional archive of landscape evolution from the Middle Pleistocene. Herein we present detailed paleopedological and sedimentological surveys, as well as first micromorphological observations to address the sequence in its entirety and the processes leading to its genesis. Furthermore, high resolution color and carbonate analyses, as well as detailed texture analyses, have resulted in a substantial database.

The studies show that the loess sediments were subject to a polygenetic development under periglacial conditions reflected in eolian silt and fine sand accumulation, admixture of local material during (mostly solifluidal) redeposition and in situ processes. Horizons with signs of pedogenesis, particularly the two pedocomplexes, document longer phases of stability; the stages of development can be correlated to equivalent sequences and seen as paleoclimatic signals where chronological data are available. The upper pedocomplex is a Chernozem of the early last glacial (MIS 5c–[a?]), which developed in a solifluidal redeposited (MIS 5d) interglacial Cambisol (MIS 5e). Cryosols, typical for MIS 6 sequences, are present in the loess sediment below. The lower pedocomplex formed during several warm stages of varying intensities, with interruptions caused by colluvial processes and admixture of eolian sediment during colder stages.

## Paudorf *locus typicus* (Niederösterreich) – Das Potential des klassischen Lössaufschlusses für die Rekonstruktion mittel- bis spätpleistozäner Paläoumwelten

## Kurzfassung:

Die über 12 m mächtige Löss-Paläoboden-Sequenz in Paudorf (Niederösterreich), ist seit Jahrzehnten bekannt als *locus typicus* für die „Paudorfer Bodenbildung“. Dieser 1 m mächtige Pedokomplex im obersten Profilabschnitt entwickelte sich eem- bis frühwürmzeitlich. Das differenzierte Lösssediment im Liegenden, mit einem über 2 m mächtigen Pedokomplex im basalen Bereich, stellt ein außergewöhnliches Archiv mittelpleistozäner Landschaftsentwicklung dar. Die paläopedologischen und sedimentologischen Untersuchungen sowie erste mikromorphologische Analysen widmen sich erstmals der Gesamtabfolge und den Prozessen, die zu ihrer Entstehung beitragen. Farb- und Karbonatanalysen in hoher Auflösung sowie detaillierte Korngrößenanalysen dienen dabei als hilfreiche Datenbasis.

Die Untersuchungen zeigen zum einen, dass die Lösssedimente einer komplexen Genese unter periglazialen Bedingungen unterlagen. Neben äolischer Akkumulation spielten (zumeist solifluidale) Umlagerungsprozesse unter Einmischung von lokalem Material ebenso wie in situ Prozesse eine Rolle. Morphodynamisch stabile Phasen sind insbesondere in den beiden Pedokomplexen nachweisbar. In Bereichen, in denen Datierungen vorliegen, können Entwicklungsphasen mit jenen anderer Sequenzen korreliert werden und als klimatische Signale interpretiert werden. Die „Paudorfer Bodenbildung“ ist vermutlich eine frühglaziale Humuszone (MIS 5c–[a?]), die sich in einer solifluidal umgelagerten (MIS 5d) eemzeitliche Braunerde (MIS 5e) gebildet hat. Im Lösssediment konnten Nassböden nachgewiesen werden, wie sie typisch für spätrisszeitliche (MIS 6) Sequenzen sind. Der untere Pedokomplex wurde offenbar über mehrere Warmphasen verschiedener Intensitäten hinweg gebildet und wiederholt von Umlagerungsprozessen und äolischer Sedimentation während Kaltphasen unterbrochen.

## Key words:

Loess, paleopedology, landscape formation, Lower Austria, Paudorf, Middle Pleistocene, micromorphology

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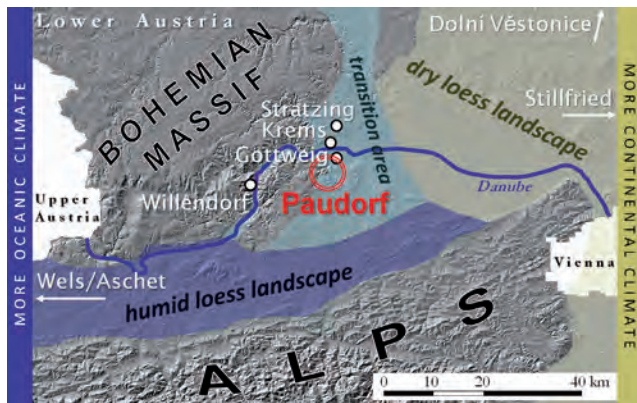


Fig. 1: Paudorf and other important loess-paleosol sequences in the loess landscapes of Lower Austria after FINK (1956). Base map: GEOinfo NÖ.

Abb. 1: Paudorf und weitere wichtige Löss-Paläoboden-Sequenzen in den österreichischen Lösslandschaften nach FINK (1956). Kartengrundlage: GEOinfo NÖ.

## 1 Introduction

Loess-paleosol sequences are the most suitable terrestrial archives to reconstruct landscape responses to Quaternary climate changes on regional to continental scale in the temperate latitudes with several  $10^3$  to  $10^4$  years resolution (cf. BRONGER 2003). For periods as far back as Marine Isotope Stage (MIS) 5 various types of records and dating methods already allow relatively detailed reconstructions. In contrast, earlier landscape responses to past climate changes remain a subject of controversial discussions (e.g. TERHORST 2007; PREUSSER & FIEBIG 2009; SCHOLGER & TERHORST 2011), despite the accurate record of global temperature fluctuations from deep-sea oxygen isotope measurements (LISIECKI & RAYMO 2005).

Reconstructions of past landscapes and climates require data from numerous localities and regions respectively, as well as reliable dating methods. For Europe, on the mentioned time scales of climate change, posterior to the Matuyama-Brunhes boundary, the most complete terrestrial archives exist in the plateau loess sediments of the Middle Danube Basin that are relatively well suited for correlations and quantitative research (BRONGER 1970, 1976, 2003; MARKOVIC et al. 2006, 2009, 2011). Geographically, Lower Austria has a key position related to paleoclimatology, between Central Europe, influenced by more marine climate, and the more continental Eastern Europe. Most thick loess deposits in Lower Austria are found on slopes in leeward positions and therefore are often influenced by hillslope processes. Due to the greater complexity of these archives, straight forward correlations and the application of quantitative methods is hampered. The need for detailed assessment of the processes leading to the formation of the archives is evident, and will give at least qualitative to semi-quantitative information about past environmental conditions.

The northeast of Lower Austria is sheltered from the moisture of the Atlantic and the Mediterranean Sea by the Bohemian Massif in the west and the Alps in the south respectively, which results in an increased continentality. Holocene Chernozems versus Eemian Bw horizons in the area give evidence for a pronounced sensitivity to climate change. This is especially true for the “transition area” around the city of

Krems a.d. Donau between the “dry loess landscape” further east and the “humid loess landscape” west of the Bohemian Massif (cf. FINK 1956; Fig. 1). The “humid loess landscape” has a present mean annual precipitation around 700–900 mm and exhibits mostly soils of the Cambisol-Luvisol group, whereas the “dry loess landscape” has an annual precipitation around 500–550 mm and exhibits mostly soils of the Phaeozem-Chernozem group.

Despite some general spatial relationships, a regional stratigraphy, as attempted by BAYER (1909, 1927), GÖTZINGER (1936), LAIS (1951), BRANDTNER (1954) and most importantly FINK (1954, 1956, 1960, 1961, 1976), has never been satisfactorily established due to a lack of suitable dating techniques as well as sequences that are polygenetic and contain gaps.

Paleomagnetic studies were carried out in the 1970s to establish stratigraphical models in the longest records of glacial-interglacial cycles of Krems Schießstätte (shooting range) and Stranzendorf (FINK 1976; FINK & KUKLA 1977; VERGINIS 1993). More recently, loess research in Lower Austria has mainly dealt with luminescence dating of some previously studied sequences (ZÖLLER et al. 1994; NOLL et al. 1994; TERHORST et al. 2011; THIEL et al. 2011a; THIEL et al. 2011b). Research is also focused on high resolution records of MIS 2 to 3, most notable for Paleolithic findings, such as at Stratzing, Willendorf and Krems-Wachtberg (NEUGEBAUER-MARESCH 1993; HAESAERTS et al. 1996; EINWÖGERER et al. 2006; THIEL et al. 2011a). Aside from that only a small number of studies have tried to unravel the very complex Middle and Late Pleistocene paleoenvironmental archives (KOVANDA et al. 1995; PETICZKA et al. 2009).

One important loess-paleosol sequence in this area is the renowned section of Paudorf, first described by GÖTZINGER (1936). In this sequence, a pedocomplex that can be correlated to MIS 5, is exposed in the upper part (ZÖLLER et al. 1994; THIEL et al. 2011b). On the basis of detailed paleopedological and sedimentological data this study aims to understand the complex pedogenetical and geomorphological processes in the interplay of local influences and climatically induced environmental changes.

## 2 The locus typicus Paudorf

Paudorf (257 m a.s.l.) is situated in the hilly transition area between the Bohemian Massif and the Molasse Basin to the east, 7 km south of the city of Krems (203 m a.s.l.). The surrounding hills consist predominantly of granulate (mostly quartz, feldspar, garnet and some mica) with sporadic lenses of ultrabasic rock (GEOLOGISCHE BUNDESANSTALT 2002; MATURA 2006). At the foot of the Waxberg (500 m a.s.l.), northwest of the village of Paudorf, the >12 m thick loess-sequence is exposed at 265–270 m a.s.l. in an approximately 70 m long outcrop facing SE and belonging to a former loampit. In Figure 2, a sketch of the outcrop with the two prominent pedocomplexes is presented. The thickness of the loess sediments between the 1 m thick upper and the > 2 m thick lower pedocomplex is increasing to the right hand side (NE).

Based on similarities to the famous paleosol of the nearby Göttsweig/Furth outcrop, GÖTZINGER (1936) correlated the pedocomplex in the lower part of the Paudorf sequence with the “Göttsweiger Verlehmungszone” sensu BAYER (1927). Following the German terminology, “Verleh-

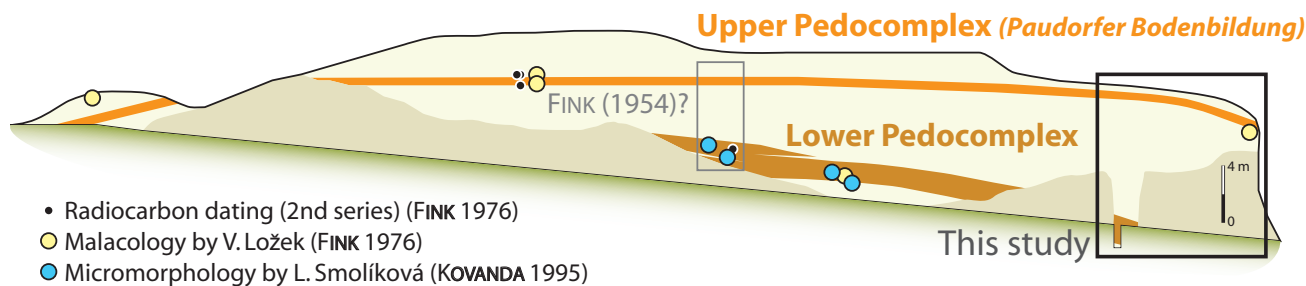


Fig. 2: Overview of the outcrop and former studies (black square: see Fig. 3). Redrawn and modified from FINK 1976, Fig. 31.

Abb. 2: Überblick über den Aufschluss und frühere Arbeiten (Schwarzes Quadrat: diese Arbeit, vgl. Abb. 3). FINK 1976, Abb. 31, neu gezeichnet und verändert.

mung” is the pedogenic process of clay neo-formation, resulting in a Bv horizon (cf. AD-HOC-ARBEITSGRUPPE BODEN 2005; Bw horizon after IUSS WORKING GROUP WRB 2006). For the upper pedocomplex, GÖTZINGER (1936) introduced the term “Paudorfer Leimenzone/Verlehmungszone”, later named “Paudorfer Bodenbildung” (Paudorf soil/paleosol) by BRANDTNER (1954) and FINK (1956). It has to be noted that the term “Paudorfer Bodenbildung” sometimes refers only to the prominent 0.3–0.4 m thick, speckled part of the ~1.0 m thick pedocomplex (e.g. in FINK 1956).

Depending on the interpretation, either both pedocomplexes in the Paudorf sequence (LAIS 1951; BRANDTNER 1954, 1956) or at least the upper one (GÖTZINGER 1936; FINK 1954, 1956) were seen to represent an interstadial of the Würm glaciation sensu PENCK & BRÜCKNER 1909. Two series of radiocarbon dating (FINK 1961, 1976) supported the idea of Paudorf being the locus typicus for a pronounced interstadial of the last glacial, the “Paudorf (Interstadial)”. In the 1960s and early 1970s this term was widely accepted and used in loess research. Interglacial mollusk assemblages in the “Paudorfer Bodenbildung”, identified by V. Ložek (FINK 1976), and the revised stratigraphy by FINK (1976), could not prevent the term “Paudorf” still being present in neighboring sciences such as paleontology and archeology as interstadial around 30 ka (PAZONYI 2004; ANGHELINU & NIȚĂ 2012).

Thermoluminescence (TL) ages presented in NOLL et al.

(1994) support the ideas of GÖTZINGER (1936), whereas the TL dating results, in combination with amino acid chronology by ZÖLLER et al. (1994), lead to a correlation of the “Paudorfer Bodenbildung” with MIS 5. Recently published post-IR infra-red optically stimulated luminescence datings (post-IR IRSL; cf. Thiel et al. 2011a) reveal ages between  $106 \pm 12$  and  $159 \pm 20$  ka for the upper pedocomplex (THIEL et al., 2011b). These results confirm its formation during the last interglacial and the early glacial, because this new dating technique is not prone to any significant over- or underestimation at this age range (BUYLAERT et al., 2012). In the loess sediment below the pedocomplex, ages of  $187 \pm 12$  and  $189 \pm 16$  ka do not suppose any major hiatus in the upper 7.5 m of the sequence. Consequently, it is most likely that most parts of the profile are of Middle Pleistocene age.

### 3 Methods

In contrast to FINK (1954, 1976), the studies were conducted in the northeastern part of the outcrop, where the sequence of loess sediments between the two pedocomplexes is thicker, thus providing higher resolution. Here, the layers/horizons are dipping 10–15° to the right hand side of the profile wall (NE). The two profiles presented in THIEL et al. (2011b), which were based on the studies of PETICZKA et al. (2009) were further extended. One composite Paudorf sequence is presented in this study, which involves the neighboring profiles Paudorf I (2.5 m thick, upper pedocomplex and surrounding loess sediments) and Paudorf II (11.8 m thick, sequence below upper pedocomplex). The presence of the upper pedocomplex in both profiles allows the correlation of the two sequences. The setting is illustrated in Figure 3. High resolution sampling of 1 dm<sup>3</sup>-blocks in 1 dm intervals (cf. Continuous Column Sampling [CCS] by ANTOINE et al. 2009) was carried out in both profiles, including some overlap in the upper pedocomplex. The color/carbonate measurements of the statistically taken samples enhance the consistency of the units and subunits clarified during the field survey. The positions of the samples for grain size analyses are indicated in Figure 6.

#### 3.1 Field survey

The description of the pedological and sedimentological parameters is based on the German Mapping Key for Soils (AD-HOC-ARBEITSGRUPPE BODEN 2005) and was adapted to

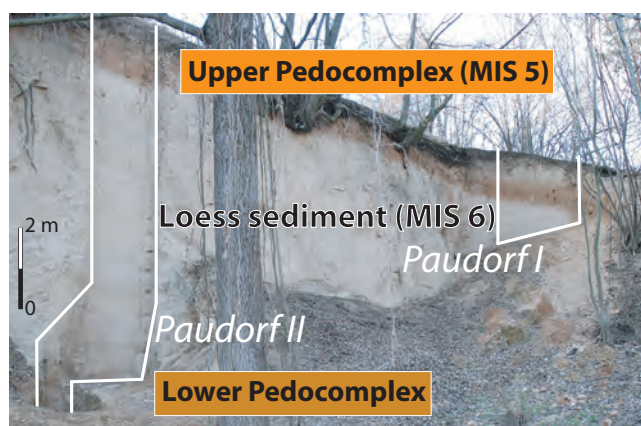


Fig. 3: Overview of the studied profiles. The lowest parts of Paudorf II in the small pit on the left hand side (cf. Fig. 4-3a) are not visible here.

Abb. 3: Überblick über die untersuchten Profile. Der untere Abschnitt von Paudorf II in der Schürfungruben linker Hand (vgl. Abb. 4-3a) ist hier nicht vollständig sichtbar.

the World Reference Base for soil resources (IUSS WORKING GROUP WRB 2006). We note that these guidelines are made for modern soils, and that many horizons exhibit features that were formed under environmental conditions different from today. Bleached horizons in loess sediment are commonly explained by iron reduction in water saturated substrates above the permafrost table in times without extreme eolian input and redeposition. These “Naßböden” sensu SCHÖNHALS et al. (1964) are close to Cryosols (Reductaquic) after WRB (IUSS WORKING GROUP WRB 2006) if assumed environmental conditions during pedogenesis are taken into account.

### 3.2 Laboratory analyses: texture, carbonate content, color

After homogenizing the samples, representative fractions (~400 g) were dried for two days at 45 °C. Half of the samples were dry-sieved to < 2000 µm to produce fine earth which was used for all analyses except the sieving procedure. Water contents were estimated gravimetrically, parallel to weighing out. 30–40 g of each corresponding sample were oven-dried (two days, 105 °C) and the results considered in weight-based calculations.

100 g of each bulk sample were without chemical pretreatment suspended in water and wet sieved for 40 min in a Retesch AS200 basic sieving apparatus (2000/630/200/63/40 µm). The fractions were oven-dried and weighted. The remaining suspension < 40 µm was discarded and its weight-percentage calculated. To measure the fractions of this suspension, of each fine earth sample 10–15 g (depending on the content of fractions < 40 µm) were without further pretreatment wet-sieved to < 40 µm. Treated with  $\text{Na}_4\text{P}_2\text{O}_7 \cdot 10 \text{H}_2\text{O}$  in solution, the fractions of the suspension < 40 µm (40/20/6.3/2 µm) were measured as weight-percent using a Micromeritics Sedigraph III 5120 by means of X-rays during sedimentation; the principle is based on Stokes' Law. Gravel (> 2000) is calculated in weight-percent of the whole sample. Fractions < 2000 µm are calculated in weight percent of total fine earth.

Following DIN ISO 10693 1–2 g of each air-dried fine earth sample were taken for gas-volumetric estimation of carbonate content of the sample after reaction with 10 ml of 10% HCl by means of Scheibler Apparatus.

The oven-dry fine earth used for gravimetric estimation of water content was taken for color measurement, using the Spectrophotometer SPH 850. Before the lens was placed on the samples, the aggregates were thoroughly pounded to prevent shadow effects. The resulting color values (10 nm steps from 400 to 700 nm) are calculated by dividing the ten highest summated intensities (700–600 nm) by the ten lowest values (500–400 nm). The values were normalized to the color of the most loess-like part in P3 (sample in 2.25 m depth). Values > 1.0 are “warmer” (brownish, reddish) colors, representing decalcification, oxidation as well as humification, while values < 1.0 are “colder” colors (bluish, whitish), resulting from iron reduction and carbonate illuviation. A single one-dimensional parameter adequately quantifying qualitative field perception was favored in this study of a polygenetic sequence over several published color proxies that seek for quantification of pedogenetic intensities (e.g. VISCARRA ROSSEL 2006; BÁBEK et al. 2011).

### 3.3 Microscopic studies

Microscopic studies have been carried out in order to obtain further qualitative insight into the complex genesis of the sequence. From several samples each of the five fractions obtained by sieving was studied under binocular microscope to assess their specific composition. Thin sections of the two pedocomplexes have been studied qualitatively under a Leitz DMRB polarizing microscope to principally understand their polygenetical formation. Detailed micromorphological studies are the subject of ongoing studies. The thin sections were produced from resin-impregnated undisturbed block samples.

## 4 Results

### 4.1 Field survey

Overview and detailed photographs of the outcrop are presented in Figure 4. Down to a depth of 11.80 m, the standard Paudorf sequence is differentiated into eight units, with units P2 and P7 covering the two prominent pedocomplexes. Below we highlight the main characteristics of the sequence. For field parameters, soil horizon designations and further descriptions see Figures 5 and 6.

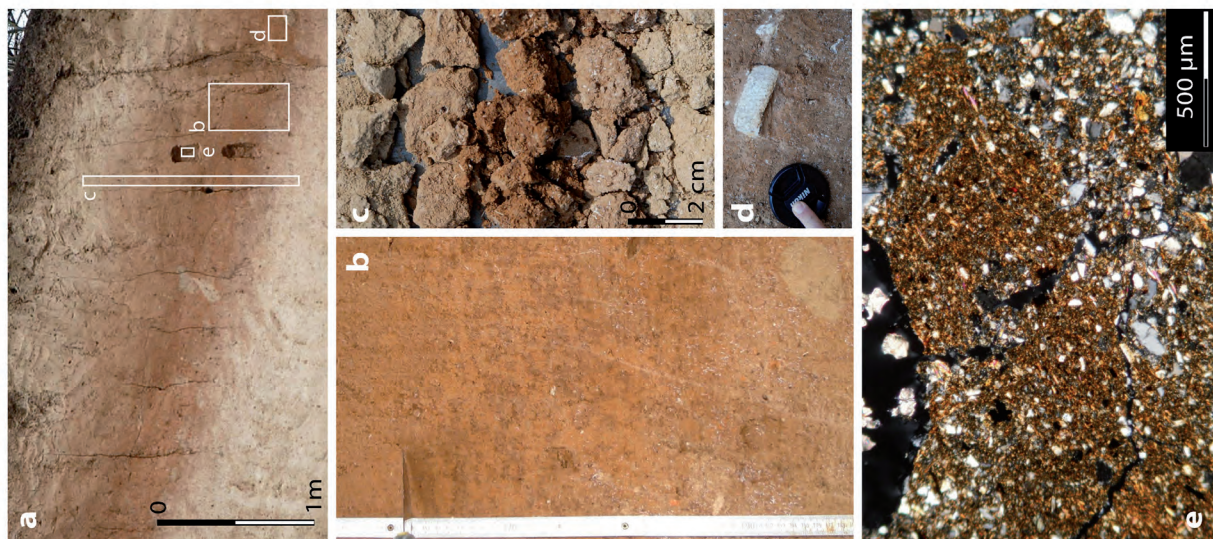
In general, the outcrop shows a stable, wall-forming, differentiated, mostly yellowish-grey substrate dominated by silt-sized particles and a considerable percentage of sand. Both, the particles of the loess-like sediment (P1, P3–6, P8) and of the pedocomplexes (P2 and P7) form blocky to sub-polyhedral aggregates. Carbonate is abundant in all units, except the lower pedocomplex (P7). Each layer/horizon bears granulite fragments of up to gravel size. Several thin bands, rich in gravel and coarse sand, are visible. In the prominent brownish to reddish pedocomplexes rock fragments as large as several cm in length are present.

The top of the profile exhibits an approximately 0.5 m thick disturbed Cambisol (not shown in the standardized profile).

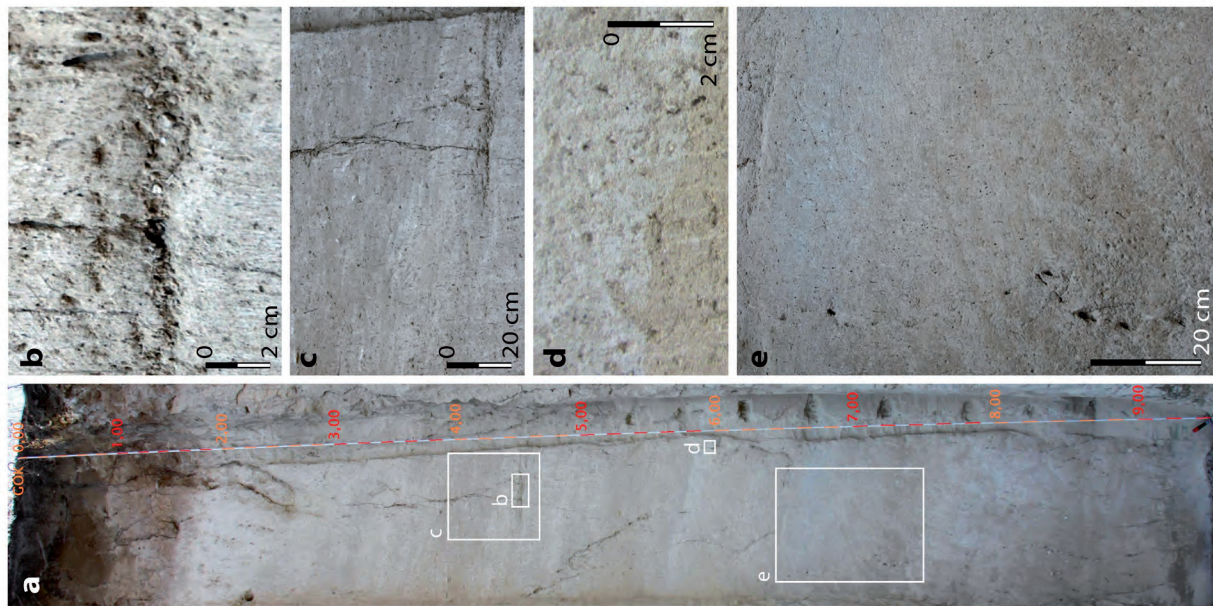
P1 has yellowish gray loess-color and loess-structure, but contains more fine sand than typical loess (cf. PYE & SHERWIN 1999). No macroscopical secondary carbonates are present. The boundary to P2, the upper pedocomplex, is gradual (Fig. 4-1a). Subunit P2a is darker and contains more clay and pseudomycelia. P2b leads into a 0.3–0.4 m thick reddish-brown horizon (P2c); this unit has a relatively regular pattern of brown interconnected irregularly shaped speckles of up to 2 cm in size and is the classical, “Paudorfer Bodenbildung” after FINK (1956) (Fig. 4-1b). The surfaces of the crumbly and porous blocks to sub-polyhedrons are free of clay coatings (Fig. 4-1c). A striking feature is the presence of several granulite cobbles (Fig. 4-1d). In addition to numerous crotovinas, the lower boundary of P2c and the entire dark-grayish subunit P2d are distorted in a complex pattern. The irregular boundaries resemble cryogenic pockets and wedges slightly deformed by hillslope processes.

P3 to P6 are the sequence of loess sediments between the two prominent pedocomplexes (Fig. 4-2a). Evidence for eolian sedimentation, redeposition and weak pedogenesis can be found in all units. P3 is the least differentiated, most loess-like substrate, with very weak signs of soil formation and colluvial processes. P4 is more heterogeneous, striat-

### 1) Upper Pedocomplex (P2)



### 2) Loess-sediment (P3-P6)



### 3) Lower Pedocomplex (P7)

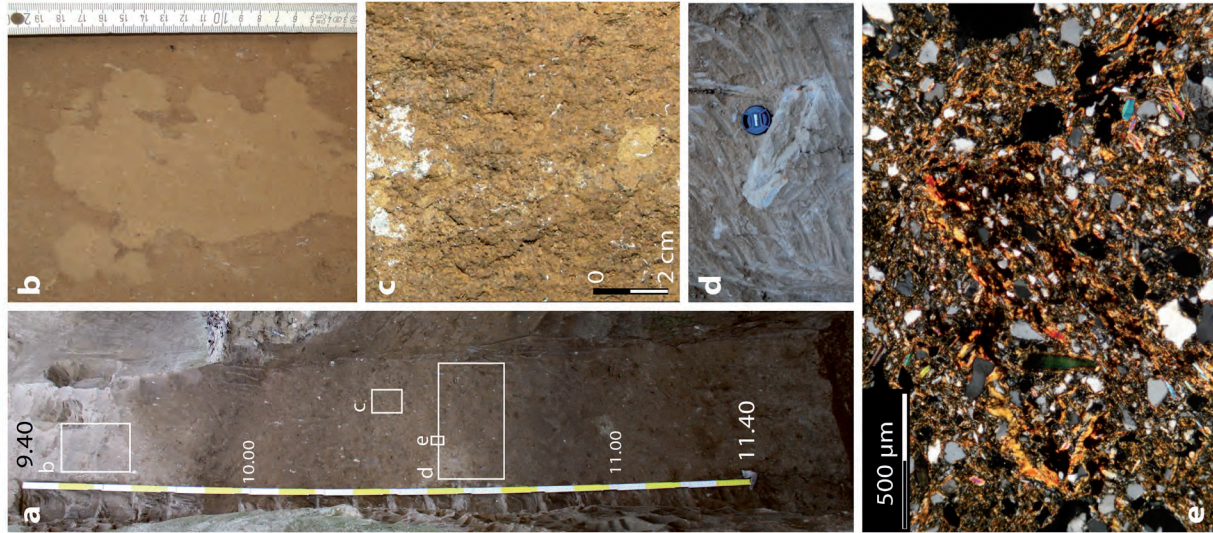


Fig. 4: Panel of macroscopic and microscopic images of characteristic parts of the Paudorf sequence. Rectangles in images 1a, 2a, 3a indicate stratigraphic position of detailed images b-d.

Abb. 4: Zusammenstellung makro- und mikroskopischer Bilder charakteristischer Abschnitte der Sequenz Paudorf. Rechtecke in den Überblicksfotos 1a, 2a, 3a zeigen die stratigraphische Position der Detailbilder b-e.

ed and contains more gravel, partly in distinct bands (Fig. 4-2b). Redeposition features can be seen clearly. A considerable amount of carbonate and silt, as well as the presence of bleached horizons, show that eolian sedimentation occurred and alternated with weak pedogenesis (Fig. 4-2c). The three Cryosols (Reductaquic) of P5 indicate clearly pedogenesis (Fig. 4-2d); especially the upper (P5a) and the lower Cryosol (P5e) are well pronounced. P6 differs considerably from the loess sediments, as it is significantly enriched in sand, has lower carbonate content and a different hue. P6b is a brownish horizon with irregular boundaries to the adjacent horizons. Striations penetrating P6a in the opposite direction to the inclination of the horizon boundary are visible (Fig. 4-2e). Further, the brownish horizon seems to be disturbed internally. Around the lower boundary and slightly below (P6c), the highest sand contents of the entire sequence are found. Sharply defined areas of darker substrate resemble crotoquina, root channels or a cryogenic structure. The color of these domains is comparable to the color of P6b.

P7 exhibits the 2.8 m thick lower pedocomplex and the transitional horizons above (Fig. 4-3a). In a continuous change of substrate from P6 to P7, a 10–20 cm thick layer rich in granulite gravel up to 2 cm in size has been chosen as the boundary (P7a). In general, P7 is carbonate free, more reddish-brown in the central subunits (P7c–e) and more blackish-brown in the surrounding subunits. In P7b unevenly shaped, sharply defined lighter brown mottles are visible (Fig. 4-3b). The staining areas are bigger, but fewer, and are less regularly formed and arranged than the speckles in P2. The subunits P7c and P7e are more consolidated and the aggregates exhibit weak clay coatings (Fig. 4-3c). The intercalated P7d contains higher amounts of granulite and charcoal fragments. P7f has higher silt and less sand content, and lighter color. Pieces of granulite rock are present in this horizon (Fig. 4-3d). P7g consists of dark humic loam with areas of weak staining and crotoquina, underlain by an inhomogeneous light brown horizon. P7i covers the grayish 10–20 cm thick transition horizon of P7 to P8, characterized by a dense network of pseudomycelia. P8 is the basal layer of the exposed profile, and consists of grayish, carbonate-rich sandy loess.

## 4.2 Laboratory results

### 4.2.1 Texture

The rare rock fragments in P2 and P7 were noticed in the field, but were not taken during sampling. For results of texture analyses see Figure 6 and Table 1. Sieving revealed that without exception, gravel (G, grain size > 2000 µm) is present in every sample with an average value of 1.4%. Values exceeding 2% are mainly associated with gravel bands enclosed in the samples (e.g. P4c–d and P7a–b).

In general, the sand (S) content in fine earth ranges from 20–30%. In P5 and P7a–e sand content is around 30% and in parts of P6 contents even exceed 40%. In P4 and P5, the bleached horizons (P4c–d and P5a,c,e) have significantly lower sand content than the neighboring layers. The overall minima are present in P2 and P3, especially near their boundary. In general, fine sand (fS) is the dominating fraction with bulk contents between 10–25%; its absolute content is responsible for variations in the overall sand con-

tent. Coarse sand (gS) follows this trend only marginally and is to a certain extent also independent of gravel variations. Medium sand (mS) ratios lie between those of coarse and fine sand.

On average, silt (U) makes up 50–60% of the fine earth. In P1–2a, P3 and P4c–d silt contents are higher, in the more sandy units P6 and P7 lower, sometimes as low as 40% (P6b–c). Silt contents of the bleached horizons P5a,c,e are ~5% higher than in P5b and d. Bulk coarse (gU) and medium silt (mS) contents are in general similar in both absolute contents and changes throughout the profile, with the first reaching not more than 30% and the latter not more than 25%. As fine silt (fS) varies only slightly around 14% throughout the profile, the changes in silt content are mainly reflected by the two coarser silt fractions. Fine silt shows an increase in the pedocomplexes (P2 and P7). In the sand rich P6, the three silt fractions have similar contents. It is interesting to note that the transition zone between P2 and P3 (enriched in secondary carbonate) exhibits relatively low bulk silt contents, where fine silt has a relative maximum and clay its absolute maximum.

Bulk clay (T) contents exceed 15% in each fine earth sample. In the loess sediment units P3 to P6, the values vary around 20% with no clear trend. In the pedocomplexes (P2 and P7), clay contents reach more than 25%. In general, relative clay maxima are present at the lower boundaries of the two pedocomplexes. Highest clay contents are present at the lower boundary of P2 and in the central horizons of P7. In the uppermost and lowermost units of loess-like sediment (P1 and P8) the clay contents differ strongly (16.6 and 25.2%, respectively).

### 4.2.2 Carbonate content and color

The results of carbonate and color analyses are shown in Figure 6 and Table 1. Carbonate content is significantly lower in the pedocomplexes (P2 and P7) than in all other units. In contrast, the underlying units (P3 and P8) both exhibit strongly enhanced carbonate contents. The upper four units are in general richer in carbonate than the lowest three (more sandy) units. P5 exhibits intermediate values. The lower pedocomplex is almost completely decalcified whereas the upper pedocomplex has 6–7% carbonate. P6b, the brownish colored horizon above the lower pedocomplex has >6% less carbonate content than the over- and underlying units.

Color values support the field observations in a more sensitive and quantitative way. P2 and P7 show much more “warmer” coloring, with P7c–d exhibiting slightly higher intensities than P2c. P6 is also more intensely colored than the more loess-like units. The minima are present in the bleached horizons (P5a,c,e and P4d) and in the carbonate illuvial horizons P3a and P8.

Some interesting relation is visible by direct comparison of the relative maxima of carbonate and color value. Clearly, the pedocomplexes exhibit strong coloring and low carbonate contents, in connection with high carbonate values below. But this pattern is also traceable on a smaller scale: Around P3b and the boundary of P3d–4 a, in P4b and very clear in P6b and below, we detect higher color values connected with lower carbonate values and carbonate enrichment below.



Unit Horizon No. (subunit)	Pedogenic horizon after WRB	Pedogenic horizon after KA 5*	Lower limit depth (cm)	Lower limit shape character after KA 5*	Munsell color(s)	Texture (after KA 5*)	Carbonate content	Description and further details
<b>P 1</b>	C	C	65	w-pl	2.5-10 YR5/4	Uls	****	Loess (redeposited?)
<b>P 2</b>	Ah	fAh	102	pl	7.5-10 YR4/4	Lu	*****	from top to bottom increasing intensity in color (reddish to blackish, depending on light source), some pseudomycelia
	AhBw	fbv-Ah	118	pl	-	-	-	transition horizon, pseudomycelia
	AhBw	fbv-A(x)h +fbv+IC	150	dr-mot	7.5 YR4/6 7.5 YR3/3-4	Lu	***	reddish-brown with brownish speckles, partly lighter colored interior (10 YR 4/4), color partly connected to surrounding of numerous pseudomycelia. Matrix decalcified. Slope parallel granulite rock, charcoal, crotovina
	AhCk	fA(x)h+ICc	163	dr-mot	10 YR4/6	Lu	****	inhomogenous, forming pockets, penetrated by crotovina, abundant pseudomycelia, brick colored concretions of max. 1-2 cm diameter
<b>P 3</b>	Ck	Cc	240	pl	2.5 Y 5/4	Uls	*****	most loess-like, carbonate rich substrate, partly secondary (pseudomycelia)
	C	C(v)	254	pl	2.5 Y 5/4	Uls	****(*)	slightly more pedogenic modification as surrounding layers (?)
	C	C	310	pl	2.5 Y 5/4	Uls	****(*)	inhomogenous, some weak striations, slight bleaching possible
	C	C(v)	340	pl-w	2.5 Y 5/4	Uls	****(*)	comparable to 3b, slightly more modified
<b>P 4</b>	C(k)	(Ck-)Cv	390	pl	2.5 Y 5/4	Uls	****(*)	inhomogenous with darker and lighter striations, especially in lower parts (dry colors 2.5 Y 7/4 vs. 2.5 Y 8/3), gravel band (290), carbonate concretions (up to 2 cm)
	C(k)	(Ck-)Cv	430	pl-w	2.5 Y 5/4	Uls	****	inhomogenous, darker (greyish), striated (dry colors: 2.5 Y 7/4 vs. 2.5 Y 8/3), gravel fragments (granulite, carbonate concretions)
	C(g?)	C(g?)	480	pl	2.5 Y 5/4	Uls	****	inhomogenous, striated, lighter (bleached?), gravel band + gravel lens (460)
	C(g?)	C(g?)	520	pl	2.5 Y 5/4	Uls	****	comparable to 4 c, gravel band (495)
	C	Cv	577	pl-w	2.5 Y 5/4 10 YR6/6	Uls	****	inhomogenous, darker, more brownish than surrounding units, less traces for redeposition
<b>P 5</b>	Cg	C(g)	593	pl	2.5 Y 5/4	Uls	****	strongly bleached, homogenous, almost white with pale red lenses of same type of substrate (dry colors: 2.5 Y 8/2 vs. 2.5 Y 7/4)
	C	C(v)	632	pl-w	2.5 Y 6/4	Uls	****	darker, more sandy
	Cg	C(g)	646	pl	2.5 Y 5/6	Uls	****	weakly bleached, not continuous
	C	C(v)	660	pl	2.5 Y 5/3	Slu	****	darker, gravel bands, especially boundary to 5 e
	Cg	C(g)	690	pl	2.5 Y 5/3	Uls	****	light blueish-grey, finer substrate, relatively homogenous
<b>P 6</b>	C	C(v)	722	w	10 YR4/3	Slu	****	transition horizon, more sandy, brownish stripes coming from unit 6 b
	Bw	f(Ah-)Bv	760	dr	10 YR4/3	Slu	****(*)	inhomogenous brownish horizon (colluvial vs. in-situ ?)
	C	fAh+ICc/Cv	860	w	10 YR5/4	Slu	****	very sandy, crotovina/root burrows, gravel band (818)
	Ah	fAh	880	pl	10 YR4/3	Slu	****(*)	fragments of granulite, up to 2 cm in diameter
	AhBw	fbv-Axh	990	pl	10 YR3/4 10 YR4/4	Lu	****(*)	from top to bottom: decreasing carbonate, increasing color intensity (dark brown humic) and pedogenic structure; mottles in lower parts, singular pseudomycelia, very weak clay coatings in lower parts
	AhBt	fb(tv-)Axh	1010	pl	10 YR3/6	Lu	*	slightly reddish color, some clay coatings
	Ah	fAh	1030	pl	10 YR3/4	Lu	*	less reddish (more brownish), inhomogenous, gravel component, iron- and manganese concretions, charcoal
	AhBt	fbvt-Axh	1060	pl	10 YR3/6	Lt2	*	slightly reddish, inhomogenous, clay coatings
	AC	fAh-Cv	1074	pl	10 YR3/4	Uls	*	lighter colored (loess component?), granulite rock fragments
	Ah	fAxh+ICc	1103	pl	10 YR3/4	Lu	*	dark humic, weak mottles, crotovina, charcoal
	AhBw	fbv-Ah	1131	pl	10 YR4/4	Lu	*	brownish(-reddish?), inhomogenous, to the bottom increasing pseudomycelia
	AhCk	fbv-Cv-ICc	1144	mot	10 YR5/4	Uls	****	greyish transition horizon, pseudomycelia abundant
<b>P 8</b>	Ck	Cv-Cc	1180	-	2.5 Y 5/4	Lu	****	(sandy) loess, pseudomycelia

Table legend (cf. KA 5*):		fossil loose substrate		humic		dr		drop-like		sh		sharp		*	
f	i	h	x	h	bioturbated	dr	mot	dr	drop-like	sh	cl	cl	clear	no	
-	+	v	v	v	weathered	w	w	w	mottled	di	di	di	diffuse	weak	
-	+	c	c	c	containing sec. carbonate	pl	pl	pl	plain	di	di	di	diffuse	moderate	
-	+	g	g	g	affected by retained water	pl	pl	pl	plain	di	di	di	diffuse	strong	
-	+	g	g	g	affected by retained water	pl	pl	pl	plain	di	di	di	diffuse	very strong	

\*KA 5 = German mapping key for soils, 5th ed. (Ad-Hoc Arbeitsgruppe Boden 2005).  
S = sand, U = silt, T = clay, L = loam (S+U+T).  
For grain size classes see Fig. 6

Fig. 5: Field description of the Paudorf sequence (stratigraphy sketch in Fig. 6)

Abb. 5: Geländedaten der Löss-Paläoboden-Sequenz Paudorf (Profilskizze in Abb. 6)

### 4.3 Microscopic studies

A regular pattern of the petrographical composition of the different fractions from the sieving (for classes cf. Fig. 6) was identified under binoculars. Gravel consists mainly of small

angular to subangular rock particles bearing quartz, feldspar, garnet and other unidentified components. The composition and structure is identical to the local granulite. The surfaces of the fragments are partly oxidized and thus have yellowish to orange pigmentation. Coarse sand has a comparable com-

position. In the medium sand fraction, granulitic material is successively replaced by aggregates of feldspar, quartz and pale garnet without darker accessories. The major portion of the 40–200  $\mu\text{m}$  fractions consists of unoxidized particles of disintegrated granulite. To a certain degree in the fine sand, and most obvious in the 40–63  $\mu\text{m}$  fraction, the petrographic composition shows greatest variety.

As no pretreatment with hydrochloric acid has been carried out before grain size analysis, secondary carbonates typical for loess-paleosol sequences (cf. BARTA 2011) contribute to the sand and gravel fractions. Carbonate concretions of 1–2 cm in size (“loess dolls”) were found in P4a–b (cf. Fig. 5). Ochre colored, rough, branch shaped fragments of up to 3 mm in length were found in the sieves that are result of carbonate cementing (hypocoating) of pore wall substrates (mostly silt) in former root channels. Plain white calcified root cells (CRC) of 0.2–1 mm size, were also present.

The above mentioned patterns are also visible in the thin sections studied under petrographic microscopes, for example in the section of P2b (upper pedocomplex). Here, (sub-) angular coarse mineral grains of > 100  $\mu\text{m}$  diameter are predominantly local clastic material (granulite rock fragments or single granulite minerals). Furthermore, secondary carbonates (CRC, mollusk shell fragments, crystallized earthworm cast) and manganese nodules can have sand size. However, the dominant fractions in the sample are silt to clay sized. It consists of a broad spectrum of minerals typical for loess (quartz, feldspar, pyroxene, amphibole, mica, primary carbonates, clay minerals), secondary (pedogenic) minerals (clay minerals, iron- and manganese oxides), including secondary carbonates (lubinite, calcitic hypocoatings). These observations are of importance with respect to the results of texture analysis.

In the upper pedocomplex, minerals are only slightly altered. Micas for example are weakly oxidized. Clay coatings are not present, and only some domains of oriented clay are traceable; in many parts the structure is granular. Fragments of more intensely colored and more clay rich substrate containing less or no carbonate are present (Fig. 4-1e). Overall the substrate is quite heterogeneous.

Apart from a higher sand content, the composition of the groundmass of the lower pedocomplex is comparable. The thin section of P7e is darker pigmented than the sections of the upper pedocomplex and contains only very few secondary carbonate (micrite). The sample contains more clay in the groundmass as well as reddish-brown colored deformed clay coatings (Fig. 4-3e). Some areas of granular structure are also present in this horizon. CRC are only present in the sample of P7b, which exhibits more granular structure.

## 5 Discussion

In the morphological setting at the foot of the Waxberg hill, slope processes contribute to the genesis of the sequence. This is best reflected in the ubiquitous presence of gravel that does not fulfill expectations regarding an eolian sediment. Furthermore, none of the renowned paleosols described earlier is monogenetic (if we understand monogenetic as development over time under relatively homogenous climatic conditions). In the following the loess sediments between the two pedocomplexes will be discussed. First correlations

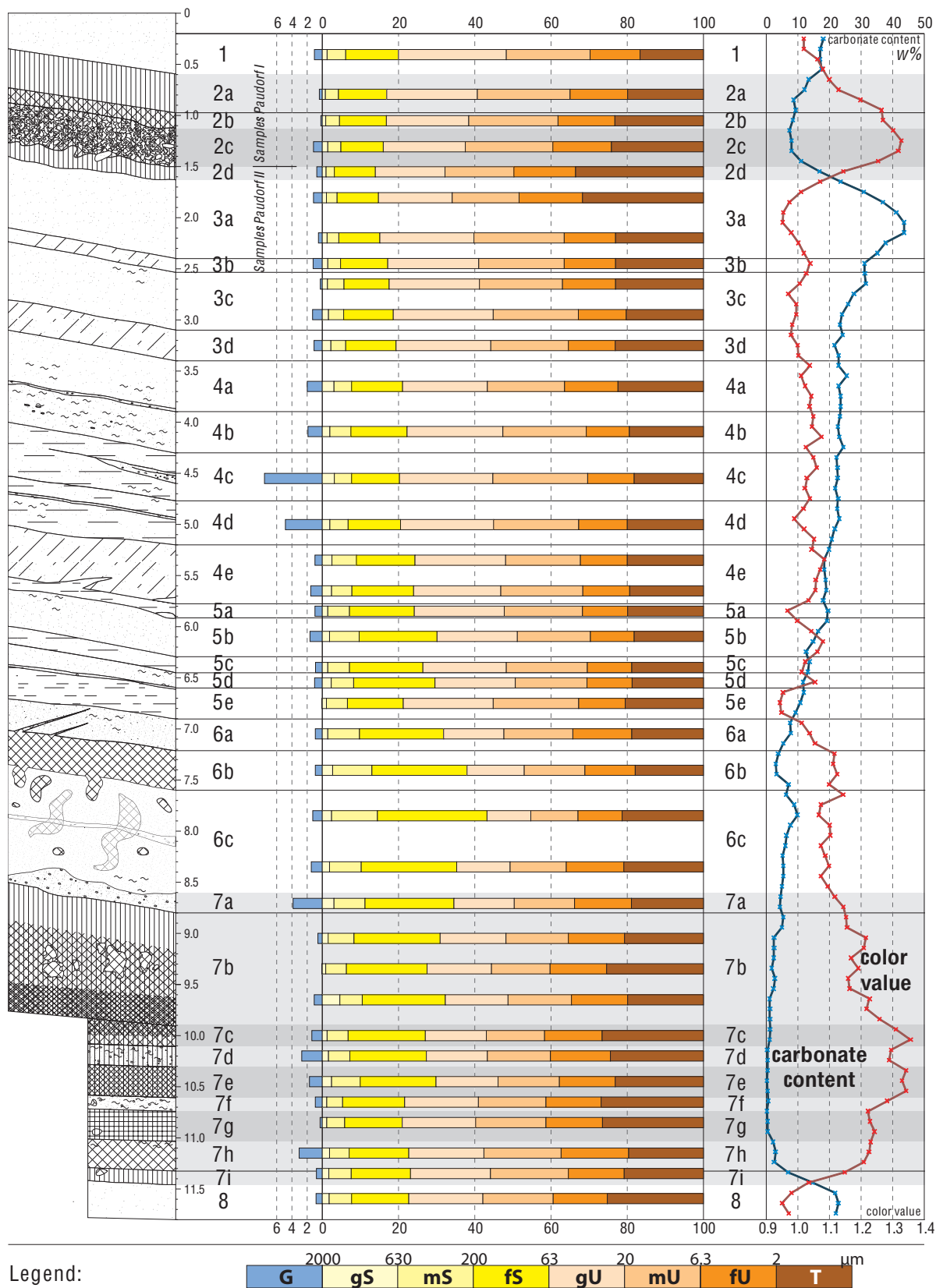
with other records are attempted where chronological data are available. Afterwards the upper pedocomplex will be reviewed on the basis of existing and new data and assumptions on its genesis and its significance presented. The lower pedocomplex lacks a reliable chronological framework; therefore only preliminary concepts in respect to its genesis can be discussed.

### 5.1 Loess sediment units [P1, P3–6, P8]

P1, P3–6 and P8 are made up mostly of silt rich yellow-gray substrate that in some parts contains considerable amounts of coarser material of local origin. Next to eolian sedimentation, redepositional processes were common, creating an admixture of local materials. The substrate has blocky to subpolyhedral structure. Several paleosols that evidence phases of morphological stability were identified in the sequence between the two pedocomplexes.

A considerable part of the primary carbonate and silt rich material in P1, P3–6 and P8 can be attributed to eolian input. The finer sand fractions with a broad petrographic/mineralogical spectrum are most likely also wind-blown deposits. Saltation and reptation in environments of sparse vegetation are possible processes for the admixture of the coarser clastic particles of local origin. The partly layered appearance of the substrate and the presence of gravel bands indicates colluvial processes. The pale yellow color and the character of aggregates – most likely due to cementation by carbonate and clay in cold steppe environments (PÉCSI & RICHTER 1996) resemble typical loess of temperate latitudes. These near-surface diagenetic processes and redeposition may have occurred with seasonal alternation and are thus to be classified as syngenetic on the studied time scales.

If loess is defined genetically as wind-blown dust (e.g. PYE & SHERWIN 1999), there is clearly no loess (sub)unit present in the Paudorf sequence. Following SMALLEY et al. (2011) input of eolian dust is the most crucial element for loess genesis; the eolian dust provides physiochemical conditions, which are of great importance for the subsequent specific processes that alter the deposit in situ. In Paudorf, mineral dust is only one component of a substrate that has structural similarity to loess. Structural development, equally important for the genesis, is supposed to have taken place in situ, in a cold steppe environment. All twelve descriptive criteria for an adequate description of loess attributed to Márton Pécsi, listed in SMALLEY et al. (2011) are fulfilled to a certain grade within P1, P3–6 and P8. An exception is the very sandy P6c in which the silt content is less than 40% and therefore too low for loess after this definition. Referring to section 3 of this list, which says that loess horizons are usually unstratified, the striated appearance, especially in P4, should be noted as unusual. In PÉCSI (1990) and PÉCSI & RICHTER (1996) criteria are slightly different but the conceptual idea is identical. Acknowledging the fundamental ideas of L.S. Berg (cf. SMALLEY et al. 2011), adjusted by PÉCSI (1990; PÉCSI & RICHTER 1996), it is possible to classify this yellowish substrate with loess structure, which is dominated by silt particles as “loess”. According to KOCH & NEUMEISTER (2005), most parts of the loess sediment in Paudorf correspond to be named solifluction sand-loess (“Solifluk-tions-Sandlöss”).



Legend:

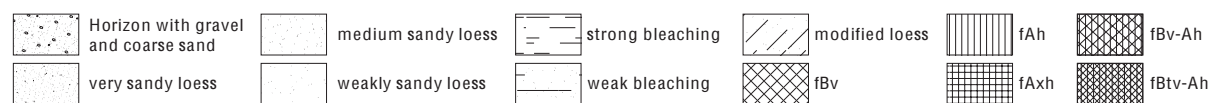


Fig. 6: Stratigraphy of the standard Paudorf sequence with texture analyses and high resolution carbonate contents and color values. The results are listed in Table 1. Grain size classes and names after KA 5 (AD-HOC-ARBEITSGRUPPE BODEN 2005).

Abb. 6: Stratigraphie der Standardsequenz Paudorf mit Ergebnissen der Korngrößenanalysen und Daten zu Carbonatgehalten und Farbwerten in hoher Auflösung. Die Ergebnisse sind in Tabelle 1 aufgelistet. Die Korngrößenklassen und -bezeichnungen folgen der KA 5 (AD-HOC-ARBEITSGRUPPE BODEN 2005).

Tab. 1: Results of laboratory analyses of texture, carbonate and color. For texture classes and names see Fig. 6. Color values and carbonate contents are calculated for the horizons of the texture analyses.

Tab. 1: Resultate der Laboranalysen von Korngrößen, Karbonatgehalten und Farbwerten. Die Korngrößenklassen und -bezeichnungen sind Abb. 6 zu entnehmen. Farbwerte und Karbonatgehalte sind auf die Horizonte der Korngrößenanalyse umgerechnet.

Sub-unit	<b>G [w%]</b>	gS [w%]	mS [w%]	fS [w%]	<b>Σ S [w%]</b>	gU [w%]	mU [w%]	fU [w%]	<b>Σ U [w%]</b>	<b>T [w%]</b>	Color value	Carbonate [w%]
P1	<b>1.0</b>	1.3	4.8	13.8	<b>20.0</b>	28.3	22.0	13.1	<b>63.4</b>	<b>16.6</b>	1.04	17.2
P2a	<b>0.4</b>	0.8	3.5	12.6	<b>16.9</b>	23.8	24.3	15.1	<b>63.2</b>	<b>19.9</b>	1.16	10.4
P2b	<b>0.2</b>	1.0	3.6	12.3	<b>16.8</b>	21.6	23.4	14.9	<b>59.9</b>	<b>23.3</b>	1.27	8.4
P2c	<b>1.2</b>	1.5	3.4	11.1	<b>16.0</b>	21.5	22.9	15.3	<b>59.8</b>	<b>24.2</b>	1.32	8.0
P2d	<b>0.7</b>	1.0	2.2	10.8	<b>14.0</b>	18.2	18.1	16.2	<b>52.5</b>	<b>33.5</b>	1.14	16.9
P3a1	<b>1.2</b>	1.2	2.7	10.9	<b>14.8</b>	19.3	17.6	16.6	<b>53.5</b>	<b>31.7</b>	0.99	34.0
P3a2	<b>0.5</b>	1.3	3.0	10.7	<b>15.1</b>	24.7	23.7	13.4	<b>61.8</b>	<b>23.1</b>	0.99	40.8
P3b	<b>1.2</b>	1.4	3.4	12.3	<b>17.2</b>	23.9	22.4	13.5	<b>59.8</b>	<b>23.1</b>	1.04	31.1
P3c1	<b>0.3</b>	1.4	4.3	11.8	<b>17.6</b>	23.7	21.8	13.9	<b>59.3</b>	<b>23.1</b>	1.01	31.6
P3c2	<b>1.3</b>	1.7	3.9	13.0	<b>18.7</b>	26.2	22.4	12.6	<b>61.1</b>	<b>20.2</b>	1.00	24.0
P3d	<b>1.1</b>	2.3	3.9	13.2	<b>19.3</b>	24.8	20.4	12.3	<b>57.5</b>	<b>23.1</b>	1.00	21.7
P4a	<b>2.0</b>	3.1	4.6	13.4	<b>21.1</b>	22.2	20.2	14.0	<b>56.4</b>	<b>22.5</b>	1.02	23.0
P4b	<b>1.9</b>	2.1	5.5	14.8	<b>22.3</b>	25.1	21.9	11.3	<b>58.2</b>	<b>19.5</b>	1.06	22.9
P4c	<b>7.6</b>	3.2	4.5	12.5	<b>20.3</b>	24.5	24.8	12.3	<b>61.6</b>	<b>18.2</b>	1.03	22.6
P4d	<b>4.8</b>	2.1	4.7	13.7	<b>20.6</b>	24.4	22.4	12.7	<b>59.5</b>	<b>20.0</b>	1.01	22.5
P4e1	<b>1.0</b>	2.6	6.4	15.4	<b>24.4</b>	23.7	19.7	12.3	<b>55.6</b>	<b>20.0</b>	1.08	18.4
P4e2	<b>1.5</b>	2.4	5.3	16.2	<b>24.0</b>	22.8	21.5	12.3	<b>56.7</b>	<b>19.4</b>	1.06	19.0
P5a	<b>1.0</b>	1.5	5.6	17.0	<b>24.1</b>	23.6	20.6	11.8	<b>56.0</b>	<b>19.9</b>	0.97	19.6
P5b	<b>1.6</b>	1.9	7.8	20.5	<b>30.1</b>	21.1	19.1	11.5	<b>51.7</b>	<b>18.2</b>	1.06	15.7
P5c	<b>0.9</b>	1.5	5.7	19.2	<b>26.4</b>	21.8	21.2	11.7	<b>54.7</b>	<b>18.8</b>	1.02	13.5
P5d	<b>1.0</b>	2.4	5.9	21.3	<b>29.6</b>	21.0	18.9	11.8	<b>51.7</b>	<b>18.7</b>	1.05	11.8
P5e	<b>0.1</b>	1.2	5.5	14.6	<b>21.2</b>	23.6	22.4	12.3	<b>58.2</b>	<b>20.5</b>	0.94	10.8
P6a	<b>0.9</b>	1.5	8.2	22.1	<b>31.8</b>	15.8	18.1	15.5	<b>49.4</b>	<b>18.8</b>	1.04	7.8
P6b	<b>0.9</b>	2.8	10.3	24.9	<b>38.0</b>	15.0	15.9	13.2	<b>44.1</b>	<b>17.9</b>	1.12	3.2
P6c1	<b>1.3</b>	2.5	12.0	28.7	<b>43.2</b>	11.5	12.4	11.6	<b>35.4</b>	<b>21.3</b>	1.07	9.9
P6c2	<b>1.5</b>	1.9	8.4	25.0	<b>35.3</b>	14.0	14.8	15.1	<b>43.8</b>	<b>20.9</b>	1.10	5.5
P7a	<b>3.8</b>	3.1	8.1	23.3	<b>34.5</b>	15.8	15.8	14.9	<b>46.5</b>	<b>19.0</b>	1.13	4.5
P7b1	<b>0.6</b>	1.6	6.8	22.6	<b>31.0</b>	17.2	16.4	14.7	<b>48.3</b>	<b>20.7</b>	1.22	2.5
P7b2	<b>0.1</b>	0.9	5.5	21.1	<b>27.5</b>	16.8	15.4	14.8	<b>47.0</b>	<b>25.4</b>	1.19	1.8
P7b3	<b>1.1</b>	4.7	5.9	21.7	<b>32.3</b>	16.5	16.7	14.7	<b>47.8</b>	<b>20.0</b>	1.23	1.1
P7c	<b>1.4</b>	1.3	5.5	20.2	<b>27.0</b>	16.0	15.2	15.1	<b>46.4</b>	<b>26.6</b>	1.33	1.2
P7d	<b>2.7</b>	1.6	5.7	20.1	<b>27.4</b>	16.0	16.5	15.8	<b>48.2</b>	<b>24.4</b>	1.29	0.4
P7e	<b>1.7</b>	2.5	7.5	19.8	<b>29.8</b>	16.3	16.1	14.6	<b>47.0</b>	<b>23.2</b>	1.33	0.3
P7f	<b>0.9</b>	1.2	4.2	16.3	<b>21.7</b>	19.3	17.7	14.5	<b>51.5</b>	<b>26.9</b>	1.28	0.6
P7g	<b>0.3</b>	1.2	4.7	15.1	<b>21.0</b>	19.3	18.4	14.8	<b>52.5</b>	<b>26.5</b>	1.23	0.5
P7h	<b>3.0</b>	1.9	5.2	15.6	<b>22.7</b>	19.7	20.3	17.7	<b>57.7</b>	<b>19.7</b>	1.23	3.0
P7i	<b>0.8</b>	1.7	5.9	15.5	<b>23.1</b>	21.0	20.4	14.7	<b>56.1</b>	<b>20.8</b>	1.15	7.0
P8	<b>0.8</b>	1.8	5.9	15.0	<b>22.7</b>	19.4	18.4	14.2	<b>52.1</b>	<b>25.2</b>	0.97	22.4

Morphologically stable phases of slightly moister conditions with reduced sedimentation rates (cf. BIBUS 1974, ANTOINE et al. 2009) are documented by Cryosols. The higher silt and lower sand contents in the bleached horizons of P5 compared to the surrounding (sub)units appear significant. They can be attributed to cryogenic weathering, but also to ongoing dust input. Post IR-IRSL ages (THIEL et al. 2011b) indicate that the major part of the sequence between the two pedocomplexes (P3–5) can be attributed to MIS 6. The formation of Cryosols in the penultimate glacial is reported for several locations from Western to Central Europe. For Belgium/Netherlands “tundrasols” are reported by MEIJIS (2002) in the “B-Loess”. This correlates to the findings of BIBUS (1974, 2002), who names a sequence of six “Naßböden” (wet soils) in the “Jungriss” loess as “Bruchköbeler Böden” (BIBUS 2002). These paleosols are labeled from base to top

B1 to B6. B1 and B3 are more pronounced (BIBUS 1974, 2002) and potentially correlate to the Cryosols P5a and P5e in Paudorf.

The genesis of P6b, a brownish horizon with reduced carbonate content in the upper part of the sandy P6 remains unclear. It appears rather inhomogeneous and could be interpreted as soil sediment. A crotovina-like pattern below can be seen as evidence for morphologic stability in a warmer phase.

The relation between carbonate content and color described in 4.2.2 could be interpreted by slight sedimentological changes. On the other hand it is possible that further phases of morphological stability were detected several times in the profile: Slight decalcification with an increase in coloring, in combination with calcification a few decimeters below, could be attributed to slight pedogenesis.

## 5.2 Upper pedocomplex (P2)

The upper pedocomplex (P2) formed during MIS 5 (ZÖLLER et al. 1994; THIEL et al. 2011b). The combination of its features points strongly to polygenesis: It is characterized by a reddish-brown color, partly in a speckled pattern, enhanced clay content and less carbonate compared to the loess sediments. However, primary carbonates can be detected. Clay coatings were not identified. Rock fragments that are oriented parallel to the inclination of the horizons and frost structures at the lower boundary are characteristic.

P2, the “Paudorfer Bodenbildung”, is contrary to previous assumptions not a Bw or Bt horizon (BRANDTNER 1954; PETICZKA et al. 2009); neither it can be classified as an Ah horizon (SEMMELE 1968; BRONGER 1976). V. Ložek (in FINK 1976) stated that the “Paudorfer Bodenbildung” is a complicated pedocomplex. However, no attempt was made to investigate the sequence in greater detail to reconstruct the processes leading to its formation. We assume the formation of a Chernozem in a redeposited Cambisol. During redeposition of the Cambisol, admixture of primary carbonate took place that was not leached during the development of the Chernozem.

### Stages of development

The first step in the genesis of the pedocomplex after deposition of the parent material (loess sediment) is the development of a Cambisol. Passive and active reddish-brown pigmentation due to carbonate leaching and formation of iron oxides respectively lead to the formation of a Bw horizon. In general, rubified soils develop in ecosystems with a pronounced dry season. However, the reddish component can also be explained by reddening after burial (RESTALLACK 2001). The carbonate maximum is found directly below P2, in P3a. This could evidence relatively shallow decalcification if the Cambisol was not too much eroded. It has to be noted in this context that the role of P3a as uppermost loess-unit in the subdivision of the polygenetic loess-paleosol sequence is ambiguous. Being the Ck-horizon of the Cambisol it could be assigned also as lowermost subunit of P2. Reddish pigmentation is as shallow decalcification in general common in environments with a seasonal precipitation.

Clay contents in P2 are enhanced; therefore the question of humidity resulting in hydrolytical weathering with clay neoformation is to be discussed. It is firsthand likely that the higher fine silt and clay content at the boundary of P2 and P3 is partly due to precipitation of microcrystalline calcite that is visible in thin sections. However, some chemical weathering resulting in clay neoformation likely took place, as decalcified clay-rich material was detected in the thin section (Fig. 4-1e). V. Ložek found mollusks (*Aegopis verticillus* [LAM], *Pagodulina pagodula* [DESM.], *Cepaea vindobonensis* [FÉR.]) at the lower boundary of the “Paudorfer Bodenbildung”, which indicate humid interglacial conditions; even more humid than indicated by the Holocene fauna (in FINK 1976). Our results, however, give no evidence for pronounced humidity, which would most likely result in clay illuviation. To summarize the first phase in the genesis of the pedocomplex, the formation of a Cambisol, potentially in a temperate climate with seasonal precipitation is assumed.

Rock fragments as well as primary carbonate grains imply that the (probably decalcified) Cambisol was later trun-

cated and reworked. During redeposition wind-blown dust with carbonate component was admixed. At least one interval of periglacial conditions during this phase can be assumed; the rock fragments in P2 are oriented parallel to the inclination of the horizons, which is typical for solifluction, and the boundary to P3 is characterized by deformed wedges and pockets, which are most likely caused by cryogenic processes.

A dark pigmentation of P2, primary carbonate and a granular structure led SEMMELE (1968) and BRONGER (1976) to the conclusion that the “Paudorfer Bodenbildung” at the locus typicus is a fossilized Chernozem. In its upper part V. Ložek detected a *Tridens*-fauna that corresponds to early last glacial interstadial Chernozems (FINK 1976). This typical steppe soil is characterized by humification and bioturbation. Calcified root cells may be relics of the steppe environment. The speckles/mottles in humic horizons are likely caused by post-pedogenetical humus degradation and are a common feature in early last glacial humic horizons (ROHDENBURG 1964).

Pedogenesis in the steppe environment was successively replaced by accumulation of eolian dust and fine sand, documented in the diffuse upper boundary of P2 and increasing silt contents in P1.

### Chronology and correlation

The post-IR IRSL date of  $106 \pm 12$  ka above the “Paudorfer Bodenbildung” published by THIEL et al. (2011b) indicates that pedogenesis during MIS 5a is not documented in Paudorf. On the contrary, ZÖLLER et al. (1994) revealed TL-ages of  $54 \pm 6$  ka from a comparable stratigraphical position.

On the base of these datings and our results, the Cambisol formation is attributed to MIS 5e (Eemian), the redeposition to the cold 5d stage, and the early glacial interstadial Chernozem to MIS 5c[-a]. The processes found in P2 during Eemian and early last glacial can be correlated with standard MIS 5 pedocomplexes in Central Europe: The paleosol stack of Stillfried A, also named “Stillfrieder Komplex” exhibits a basal “Verlehmungszone” (Bw horizon) of a truncated Cambisol and three Ah horizons on top, separated by thin layers of loess sediment (FINK 1954; BRONGER 1976). In the German loess stratigraphy a comparable scheme exists. The fossil Ah horizons above the remaining Bt horizons of the truncated Eemian Luvisols are named Mosbacher Humuszonen (SCHÖNHALS et al. 1964; SEMMELE 1968). A comparable pattern of MIS 5 paleosols is found in the Czech Republic, e.g. in Dolní Věstonice (BRONGER 1976; FRECHEN et al. 1999; BÁBEK et al. 2011). In P2 of the Paudorf sequence, the paleosols of the classical MIS 5-complexes are found not superimposed but interlaced.

### Geographical significance

Pedocomplexes comparable to P2 in Paudorf are present in the region, for example in the outcrops of Stiefen, Buchberg or Göttweig-Aigen (cf. FINK 1956, 1976). At the present state of research it is likely that P2 is a typical MIS 5-pedocomplex in the transition area of the Austrian loess landscapes after FINK (1956). P2, the classical “Paudorfer Bodenbildung”, can be regarded as the missing link between the MIS 5-complexes of the “humid loess landscape” (Linzer Komplex: an Eemian fossil Bt horizon and early glacial soil sediment) and “dry loess landscape” (Stillfrieder Komplex), defined by FINK

(1956). Although FINK (1961) presented time ranges and correlations that were revised at a later stage, we agree with his early assumption that the transition area is characterized both by pedogenesis of the “dry loess landscape” as well as intensive early glacial colluvial events, which are clearly documented in the profiles of the “humid loess landscape”. In this context it should be mentioned that more recent studies have revealed that colluvial layers above Eemian paleosols in the “humid loess landscape” do not necessarily represent only the early glacial (MIS 5d–5a), but also younger stages up to MIS 3 (TERHORST et al. 2002).

In future, it has to be discussed, whether Paudorf will be again accepted as type locality, but now for the MIS 5-complex of the “transition area” in Austrian loess landscapes, and if yes, whether the term “Paudorfer Bodenbildung” should still be used or finally dismissed. It is not a monogenetic paleosol and it carries some historical burden in context of Quaternary stratigraphy.

From a broader paleoclimatological perspective, the presented findings appear promising, related to the question of Eemian (MIS 5e) paleoclimate in Central Europe. In Paudorf (and possibly in the comparable pedocomplexes of the “transition area”) as well as in Stillfried (as type profile for the “dry loess landscape”) the Eemian paleosol is a Bw horizon of a Cambisol. The climate during the Eemian in the neighboring loess regions to the west (“humid loess landscape”) and the north (Moravia) must have been significantly moister, as evidenced by Eemian Bt horizons of former Luvisols (FINK 1956; BRONGER 1976; TERHORST et al. 2002; BÁBEK et al. 2011). In contrast, in the Pannonian Basin, the last interglacial is frequently represented by Chernozems or Phaeozems (BRONGER 2003; MARKOVIĆ et al. 2011).

### 5.3 Lower pedocomplex (P7)

The overall brownish lower pedocomplex (P7) has a reddish-brown middle part. Clay content is enhanced and carbonates are almost completely leached. In parts pedogenic structure is well expressed and horizons with clay coatings reflect longer phases of morphological stability. In the upper half of the pedocomplex (P7a–e), sand contents are higher. Several horizons bear considerable amounts of granulate and even rock fragments were found in P7f, providing evidence for phases of geomorphological activity. The genesis of this over 2 m thick pedocomplex seems thus even more complex than the genesis of the upper pedocomplex (P2). The lack of chronological data however, does not allow paleoecological implications and broader correlations.

Micromorphological assessment by L. Smolíková (in KOVANDA et al. 1995) was the first approach to decipher aspects of the pedogenesis: Two superimposed paleosols were identified in the field. Microscopic analyses of both paleosols revealed clay illuviation processes and humification/bioturbation, representing forest and steppe environments respectively. Besides these findings, previous studies (e.g. FINK, 1954, 1976; VERGINIS 1993; KOVANDA et al. 1995) have not described the remarkable complexity of the lower pedocomplex.

Below, preliminary assumptions of the main formation stages are discussed. In general, several phases of pedogenesis in steppe- to forest-steppe environments, disturbed by redeposition were found:

In the lower parts of P7, a dark Ah horizon (P7g) above a weak Bw horizon (P7h) are developed. The boundary area to P8 is enriched in secondary carbonate. This succession can be interpreted as monogenetic degraded Chernozem (Phaeozem) under (forest-)steppe conditions. Intensive redeposition and input of dust and local material is attributed to a degradation phase (P7f). Both field survey and microscopical analyses indicate phases of interglacial pedogenesis and redeposition in subunits P7c–e. Deformed clay coatings and areas with clear granular structure, as observed in thin sections, reveal clay mobilization under more humid conditions (Forest ecosystem) and bioturbation/humification in a drier climate (Steppe ecosystem). Gravel and charcoal especially in P7d, can be seen as macroscopic signs of colluvial processes. P7b–a represent a complex of steppe to forest steppe soils, which were successively buried by enhanced accumulation of eolian dust and admixture of coarser sediments of local origin.

In the context of paleoclimate, the overall more intense weathering as well as the presence of clay coatings can indicate significantly moister conditions compared to the Holocene climate in the study area or the Eemian paleoclimate. However, BRONGER et al. (1998) interpret more intensive weathering in MIS13–15 soils with prolonged pedogenesis. The thickness and complexity of P7 points to a formation during more than one interglacial, but the lack of datings allows no final conclusion whether climate or time is the main factor in respect to the grade of development.

Following the assumptions of KOVANDA et al. (1995), the lower pedocomplex is at least one interglacial older than the upper pedocomplex, which was formed during the Eemian (MIS 5e) and early last glacial. ZÖLLER et al. (1994) assume as well its formation during the penultimate interglacial. As the oldest age according to THIEL et al. (2011b) in the loess sediment above the lower pedocomplex is  $189 \pm 16$  ka, the genesis of P7 in MIS 7 and/or some older interglacial(s) is most probable. In discussion of the chronology, however, possible discontinuities in P6 and P7 have to be taken into account.

It should be noted that the lower pedocomplex has been previously correlated with the “Göttweiger Verlehmungszone” (GÖTZINGER 1936; FINK 1976). However, this paleosol was dated to  $\geq 350$  ka (THIEL et al., 2011b) in Göttweig-Furth. Further, the “Göttweiger Verlehmungszone” has been correlated to the KR 4 paleosol of Krems-Schießstätte (shooting range) (GÖTZINGER 1936; FINK 1976); absolute ages are not available for that site, but in the loess below KR 4 the Brunhes/Matuyama reversal was found (FINK 1976). Therefore, it can be hypothesised that a major portion of the Middle Pleistocene interglacials could be recorded in Paudorf P7.

The two pedocomplexes in Paudorf thus cannot reflect two interstadials of the last glacial as postulated by LAIS (1951) and BRANDTNER (1954, 1956) but may be two interglacial pedocomplexes, the upper one representing the Late Pleistocene and the lower one an as yet unknown time span of Middle Pleistocene.

## 6 Conclusions

Detailed paleopedological, sedimentological and micromorphological analyses reflect the complex genesis of Paudorf locus typicus. Although this archive was seen as not favora-

ble during its scientific history (cf. FINK 1954, 1976, BRONGER 1976), it records a well resolved and thus a considerable archive to reconstruct landscape development during Middle Pleistocene and MIS 5 in the interplay of local influences and climatically driven environmental changes. This study demonstrates that process oriented research can provide valuable insights into paleoenvironmental archives even in complex morphological settings. The scientific potential of the sequence is not only focussed on the prominent pedocomplexes, but further in the sequence of loess sediments in between. In complex loess-paleosol sequences it is next to reliable datings crucial, to trace phases of pedogenesis in order to detect phases of stability, which allow for correlation with loess-paleosol sequences of the same age.

P2, the “Paudorfer Bodenbildung”, is a pedocomplex attributed to MIS 5. It is an early glacial Chernozem that developed in the redeposited (MIS 5d) Eemian Cambisol. From a paleoclimatological perspective it is important to note that P2 in Paudorf includes, like the basal soil of Stillfried a further example for Eemian Cambisols in Lower Austria.

P3–5 probably represent MIS 6. Under periglacial environmental conditions eolian dust sedimentation and (solifluidal) redeposition took place as well as pedogenesis (Cryosols) during more stable geomorphodynamic phases. To establish the genesis and age of P6, which includes a brownish horizon, micromorphological and geochronological investigations must be carried out. The thick lower pedocomplex (P7) provides research potential for the future, as it seems to represent local effects of few 100 ka of climatically driven sedimentation, pedogenesis and colluvial activity during the Middle Pleistocene.

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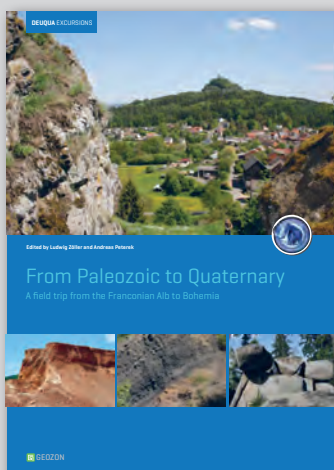
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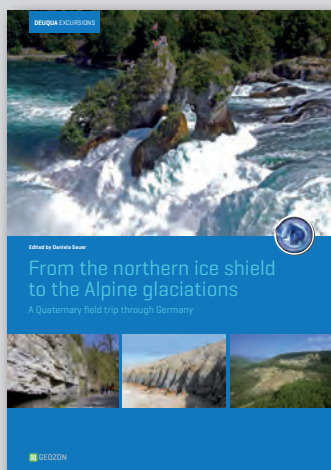
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Furthermore, the association has set itself the task of operating the contacts between the Quaternary Scientists and related organizations at home and abroad.

The DEUQUA published annually several editions of "E&G – Quaternary Science Journal". In that journal research results from the field of Quaternary Science are published. In addition, developments in the DEUQUA are announced in the "Geoscience messages" (GMIT). GMIT is published quarterly.

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Eiszeitalter und Gegenwart  
Quaternary Science Journal



Vol. 62  
No 2  
2013

## ENVIRONMENT – MAN – GEOHAZARDS IN THE QUATERNARY

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Extent of the August 2005-debris flow deposits in Antholz-Mittertal village (Fig. p. 141)

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## Foreword

*special issue*

Ludwig Zöller

The present volume 62/2 of “E&G Quaternary Science Journal” contains contributions submitted to the 36<sup>th</sup> assembly of the “Deutsche Quartärvereinigung” (DEUQUA, German Quaternary Association), held at the University of Bayreuth from 16 to 20 September, 2012. Thanks to generous support from the Deutsche Forschungsgemeinschaft (DFG) substantial contributions by external scientists were made possible. The program included 45 oral and 53 poster presentations, a minor part of which are being published for the first time in the present volume. Nevertheless, the publications presented here reflect the broad scope of the scientific program under the principal theme “Umwelt – Mensch – Georisiken im Quartär” (Environment – Man – Geohazards in the Quaternary). This included geoarchaeologic research as well as studies in the sedimentary heritage of Quaternary processes influencing present-day and future geohazards.

The first article by *Solis-Castillo et al.* investigates the provenance of Holocene sediments in the Mayan Lowlands in Central America, their degree of weathering, palaeosols and pedostratigraphy, and their chronology.

The second article by *Tillmann et al.* deals with the historical processes and landscape evolution of the western coast of the island of Amrum in the German tidelands of the North Sea. The results are obtained by careful multidisciplinary work including historical maps and documents as well as geophysical and sedimentological methods.

The third and the fourth articles report on the state of the art in deciphering the Palaeolithic record at the Lower Gravettian open air site of Grub-Kranawetberg, Lower Austria. The contribution by *Antl* highlights recent and ongoing archaeological research, thereby focussing on Archaeological Horizons 4 and 3, which point to different groups of humans occupying the site within a short time under changing environmental conditions. The article by *Zöller et al.* (edited by

Holger Freund) reports on a diachronic chronology of the site and discusses some unexpected problems with luminescence dating techniques (IRSL and OSL) in comparison with calibrated radiocarbon ages.

The relevance of Quaternary researchers dealing with present and near future geohazards in an Alpine environment is demonstrated in the article by *Damm & Felderer*. Based on intensive studies in the spatial distribution of debris flows since the Little Ice Age, the authors attempt to project their results into the middle of the 21st century for a scenario of 1 to 2 K atmospheric warming. The retrospective study by *Jaeger et al.* investigates a landslide from 1957 AD in the average mountain area built up from slightly dipping Mesozoic sedimentary rocks in Northern Bavaria, Germany. The area is very prone to landslides. The contribution is able to further differentiate some elements and processes of the landslide compared to earlier studies. With respect to ongoing climate change and possible change of precipitation patterns an endangered susceptible slope area is indicated in the future.

The last article by *Diedrich* deals with a famous cave in the northern Franconian Alb, Germany. The intermittent use of the cave by Pleistocene animals and humans is discussed with respect to the geomorphologic evolution of river terraces. The contribution boldly raises the question of a limited valley glaciation in the area during the Last Glacial Maximum at altitudes of around 500 m a.s.l., sedimentary remnants of which may only have been protected in caves from later erosion.

The editors are much obliged to the authors for submitting their manuscripts and hope for a comparatively wide range of contributions in future.

Bayreuth, 1<sup>st</sup> December 2013

# Holocene sequences in the Mayan Lowlands – A provenance study using heavy mineral distributions

Berenice Solís-Castillo, Christine Thiel, Héctor Cabadas Baez, Elizabeth Solleiro Rebolledo, Sergey Sedov, Birgit Terhorst, Bodo Damm, Manfred Frechen, Sumiko Tsukamoto

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## Abstract:

Heavy mineral analysis of alluvial sediments and paleosols on Holocene terraces of the Usumacinta River provided an effective tool to reconstruct sediment provenance in the Mayan Lowlands. Furthermore, the mineralogical data are useful for pedostratigraphic correlations in the region. Based on our observations from the Tierra Blanca profile, the ultrastable detrital heavy minerals assemblage (mostly zircon, tourmaline, and rutile) are the most promising mineral proxies to recognize the provenance of the sediments. Those minerals are accompanied by an intriguing variety of high density authigenic minerals (including titanite). Using the specific characteristics and the ages obtained for some layers, it may now be possible to develop a regional chronostratigraphy for the paleosols and alluvial sequences. Our data suggest that sediments were transported westward in river channels originating from the highlands of Guatemala. The studied materials also contain high amounts of volcanic minerals, most of them fresh and with angular shapes, thus indicating a proximal source, mostly likely from Tacana Volcano, Mexico/Guatemala.

## Holozäne Sequenzen in den Tieflandgebieten der Mayas – Eine Untersuchung der Liefergebiete auf Basis von Schwermineralgesellschaften

## Kurzfassung:

Schwermineralanalysen an alluvialen Sedimenten und Paläoböden des Usumacinta-Flusses sind ein sehr effektives Werkzeug für die Rekonstruktion der relevanten Liefergebiete in den Tieflandgebieten der Mayas. Die mineralogischen Daten können für die pedostratigraphische Korrelation in der Region nützlich sein. Auf der Basis unserer Beobachtungen für Tierra Blanca ist die ultrastabile Schwermineral-Komponente (zumeist Zirkon, Turmalin und Rutil) besonders erfolgversprechend für den Nachweis der Liefergebiete der Sedimente. Diese Minerale treten gemeinsam mit einer großen Vielfalt von sehr dichten authigenen Mineralen auf (u.a. Titanit). Unter Berücksichtigung der spezifischen Eigenschaften und der Alter von ausgewählten Horizonten, kann nun versucht werden, eine regionale Chronostratigraphie für Paläoböden und alluviale Sequenzen zu entwickeln. Unsere Daten deuten auf Sedimenttransport in Flussrinnen von den Hochländern Guatemalas gen Westen hin. Die untersuchten Schichten enthalten auch große Anteile an vulkanischen Mineralen mit frischen und eckigen Formen, was auf ein proximales Liefergebiet (wahrscheinlich Tacana Volcano, Mexico/Guatemala) hinweist.

## Keywords:

*heavy mineral assemblages, pedostratigraphy, micromorphology, alluvial sediments, Mayan Lowlands, Mexico*

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## 1 Introduction

The reconstruction of the environmental and landscape changes in Mexico for the Late Quaternary is based on numerous data. The majority of data is derived from lacustrine sediments, glacial and paleopedological records; however, due to the fragmentary character of glacial records, hiatuses in lake cores, and low temporal resolution of te-

phra-paleosols sequences these records are still contradictory (SEDOV et al. 2007). With respect to this problem, new insights might be obtained by studying alluvial sequences. Until now the alluvial and fluvial archives in Mexico have not received sufficient attention as a source to understand short-term changes in the landscape; only very few detailed studies deal specifically with flu-

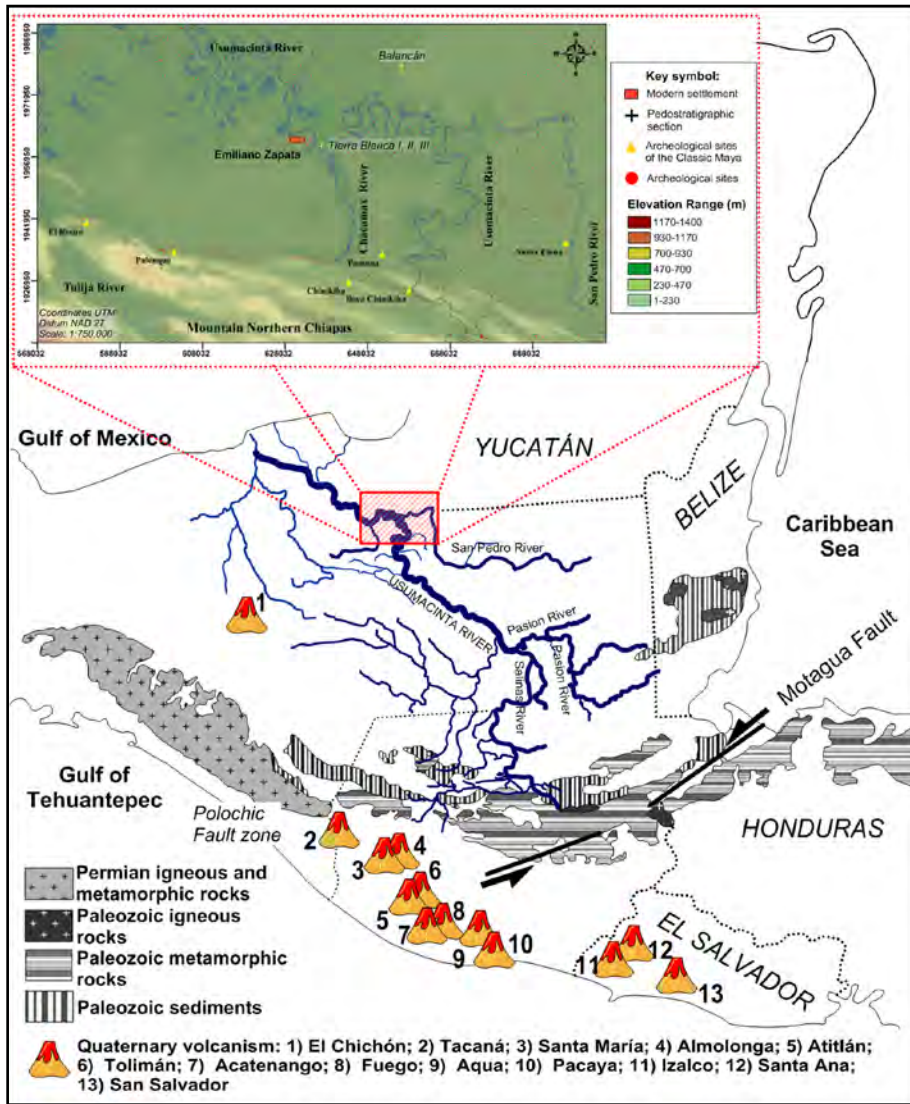


Fig. 1: Location of the study area and simplified geological map showing the boundary between the Maya and the Chortis blocks. Modified after ORTEGA-GUTIÉRREZ et al. (1992), and FRENCH & SCHENK (1997).

Abb. 1: Lage des Untersuchungsgebiets und vereinfachte Darstellung der regionalen Geologie mit der Grenze zwischen der Maya- und der Cortis-Platte. Verändert nach ORTEGA-GUTIÉRREZ et al. (1992) und FRENCH & SCHENK (1997).

vial Late Pleistocene-Holocene sequences (cf. SOLLEIRO-REBOLLEDO et al. 2011).

Particularly, the alluvial-paleosol sedimentary sequences in the Southeast of Mexico are highly sensitive to the environmental changes and are rich and detailed Holocene archives (SOLÍS-CASTILLO et al. 2013). However, river systems are very complex due to the dynamics involved in their formation, thus controlling the different phases of aggradation, degradation, and pedogenesis. Aggradation phases can be recognized by selective sorting of particles due to differential entrainment and transport. This is especially seen in the characteristics of heavy minerals, such as density, grain size and shape. In general, the sorting may be a result of ancient depositional environments and provenance of sediments (KOMAR 2007). In rivers with high sedimentation rates, heavy minerals are affected by considerable sorting before they reach the alluvial plain (PIRKLE et al. 2007).

Alluvial landscapes are very complex due to the presence of discontinuities in vertical successions, being controlled by several forces (climatic, geologic, geomorphic, anthropogenic), and interpretation of the successions can be problematic. Similar sediments can be produced in different and distant areas or periods; this makes paleoenvironmental reconstruction difficult. In consequence, paleosols are important records for the environmental history

and can help to unravel the alluvial stratigraphy (HUGHES 2010). Alluvial soils exhibit characteristics of both sedimentary and soil formation processes. Such soils are characterized by the textural and mineralogical composition of the material transported across the drainage basins (KRAUS 2002). Traditionally, analyses of mineral composition have been applied to paleosols on surfaces of different ages to establish morphostratigraphical units (SCHATZL & ANDERSON 2005). In particular, heavy minerals in soils are used as indicators for sediment provenance. Furthermore, heavy minerals have been used to determine weathering degrees, thus being useful as relative dating tools (MIKESELL et al. 2004). In sequences where paleosols show contrasting differences, weathering indices are commonly used to evaluate the intensity of the process and their relation to environmental conditions.

The Usumacinta River in Tabasco, Mexico (Fig. 1) has carried vast quantities of sediments from Chiapas and Guatemala since the Plio-Pleistocene and therefore exhibits a great potential for the application of heavy mineral analyses. At present, detailed information on the heavy mineral assemblages in soils are very scarce for the Mayan Lowlands. This area is known for its high cultural diversity since the Mid-Holocene, which is probably related to the river-channel stability and soil formation on the Holocene terraces, thus

providing space for settlements in the alluvial plain (SOLÍS-CASTILLO et al. 2013).

Altogether, the analysis of mineral provenance is of great importance to better understand the river dynamics and climatic changes in this region, which is reflected in different weathering intensities. Unraveling the changes in the landscape and in the soils will contribute to a better understanding of human migration along the Mayan Lowlands.

It is surprising that most of the information about the climate history of the Maya region comes from studies in Yucatan (ROSENMEIER et al. 2002; HODELL et al. 2005; SEDOV et al. 2007; FREDICK et al. 2008; CABADAS et al. 2010), as well as from Guatemala and Belize (DAHLIN et al. 1980; BEACH et al. 2003; FERNANDEZ et al. 2005; DUNNING et al. 2002, 2006). Until now, for the Maya Lowlands, only SOLÍS-CASTILLO et al. (2013) provided data on the reconstruction of Holocene climatic changes in the Usumacinta River region on the base of paleopedology.

The authors found that the Late Pleistocene-Early Holocene paleosols show strong weathering and gleyzation related to more humid climatic conditions. These conditions were only observed in the paleosols with gleyic features, while in the alluvial sediments only few oxide-reduction characteristic were found. Mid-Holocene paleosols are characterized by the presence of carbonate concretions combined with vertic features, which indicate a major drying trend as well as fluctuations between wet and dry conditions. Late Holocene paleosols show vertic properties and lack gleyic features; this corresponds to dryer conditions. Further, Middle and Late Holocene paleosols are observed at the base of other sequences located on the younger terraces; these paleosols do not show any signs of oxide-reduction conditions.

In this study we use heavy minerals to detect the weathering degree of alluvial sediments and paleosols; these data serve as proxy for the relative stability during the Late Pleistocene and the Holocene. Some samples were selected for micromorphology in order to investigate the bulk mineral composition.

Furthermore, the composition of the heavy minerals from the alluvial sediments of the Usumacinta River helped to reconstruct the provenance. In addition, radiocarbon and luminescence ages provide a numerical chronology for the sediments under study.

## 2 Regional setting and the Usumacinta River

The Mayan Lowlands are characterized by active volcanoes, rugged terrain of the Sierra de Chiapas and the Central Cordillera, and faulting as well as extensive karst systems in the northern lowlands (Figure 1). Volcanism and tectonics are a result of the highly variable spatial and temporal evolution of plate boundaries between Cocos, North American, and Caribbean Plates (DONNELLY et al. 1990).

The Sierra de Chiapas is mainly formed by Tertiary folded, northwest-southeast oriented limestones. The limestones show extensive karstic features with abundant subterranean drainage and ephemeral surface streams. The highlands of the Chiapas region are composed of extru-

sive igneous rocks (andesites, dacites, and pyroclastic products) and sedimentary rocks (shales, sandstones, and limestones). Their ages range from Late Cretaceous to Tertiary, and Quaternary (HERNANDEZ et al. 2012). The Paleozoic geology in Chiapas is characterized by metamorphic rocks, which consist of granitic gneiss and gneiss of biotite and orthoclase, sedimentary rocks, and biotite schists. In the western part of Guatemala, gneiss and schists of quartz are present. The Paleozoic Granitic Massif of Chiapas is composed of pink granite of biotite with gradation to granodiorite (MONOGRAFIA GEOLOGICA-MINERA DEL ESTADO DE TABASCO 1999).

The Usumacinta River is one of the larger fluvial systems in Mexico with a drainage area of 63,804 km<sup>2</sup> (WEST et al. 1969), which flows through both the highlands of Chiapas and Guatemala and the coastal plains of the southern Gulf of Mexico.

The latter are formed by Late Tertiary (Pliocene) and Quaternary deposits, which are mainly composed of alluvium, lacustrine and marsh sediments, as well as coastal bars and residual soils. During the Holocene, three tributary channels (San Pedro in the east, Chakamax in the center, and Tulijá in the west) flew across the terraces of the Central Usumacinta and discharged into the Lower Usumacinta (SOLÍS-CASTILLO et al. 2013).

Starting in Guatemala, the Usumacinta River runs north-east to the Bay of Campeche in the Gulf of Mexico (Figure 1). This vast region has been subdivided into two main areas: the Upper Usumacinta (from the rivers Salinas and Passion in Guatemala to Boca del Cerro, Chiapas), and the Lower Usumacinta (from Boca del Cerro to the Gulf of Mexico). In pre-Hispanic times, several population centers were located in these areas. Until now, there is evidence of 2,300 archaeological sites with different characteristics; they are preferably located in the plains of Tabasco. The number of occupation sites in the numerous side valleys along the Usumacinta River in the mountainous region is still unknown (LIENDO et al. 2012). The archeological research has revealed a long sequence of occupation for the region ranging from the Middle Preclassic period (800–300 B.C.) to the Terminal Classic period (850 A.D.) (LIENDO et al. 2012, and references therein).

In the north, the Usumacinta River flows through the State of Tabasco in an alluvial valley composed of Plio-Pleistocene to Holocene terraces (ORTIZ-PEREZ et al. 2005); these are affected by Neogene tectonic activity with a set of normal faults causing a horst-graben system (PADILLA & SANCHEZ 2007). The main tributaries follow normal fault planes. The oldest terraces are located in the areas more distant from the sea. In contrast, Holocene terraces are formed by incisions and floodplain deposits along the main channel. Three levels of Holocene terraces have been recognized by SOLÍS-CASTILLO et al. (2013), referred to as HT2 (at 15–10 m asl), HT1 (at 10–5 m asl) and HT0 (at <5 m asl), from the oldest to the youngest.

## 3 Materials

In Tierra Blanca (Figure 1), the studied sequence is composed of alluvial sediments intercalated with paleosols. Three profiles were studied (Figure 2): Tierra Blanca (TB) I, TB II and TB III. SOLÍS-CASTILLO et al. (2013) have

presented soil morphologic data, archeological and cultural evidence, as well as radiocarbon ages for TB I and TB II (amongst other sections from the same region). Here, we present new data for TB III and compare them to TB I and II in order to get a more complete understanding of the landscape formation and paleoenvironmental changes. A composite profile is shown in Figure 3. The pedological and sedimentological survey is complemented by micromorphology. Furthermore, three alluvial layers were dated using optically stimulated luminescence (OSL); these ages can be compared with the radiocarbon ages presented in SOLÍS-CASTILLO et al. (2013). Heavy mineral analysis is used to trace the provenance. In addition to the Tierra Blanca site, data from a sequence in Balancán (Figure 1) are presented; the section in Balancán is also located on a Pleistocene terrace of the Usumacinta River.

The TB I, II and III sequences are located on the oldest Holocene terrace (HT2), and represent the most complete stratigraphic section of the area. In TB I and TB II two types of paleosols are present, gleyic at the base and vertic at the top. They are clearly separated by alluvial sediments (TB III) (Figure 3).

On top of a sandy Pleistocene alluvium the lower part of the composite profile (TB I) contains four paleosols (numbered 6, 7, 8, 9). This section with a total thickness of 267 cm, is characterized by the following horizons: 6G, 7Bg, 8G, 8Gk, 9G, 9Bkg, 9BCgk (Figure 3). A horizons are not preserved due to erosional processes; there is no separation of the individual paleosols by C horizons. All horizons show strong gleyic features expressed as grayish brown colors with reddish-yellowish-greenish mottles, coarse subangular blocky structure, Fe concretions and/or spots, and dendritic Mn. In the paleosols of unit 9 the gleyic features decrease with depth, and at the base of horizon 9BCgk the oxide-reduction characteristics are less. The strongest gleyic features are observed in paleosol 6, where also slickensides are present. One of the most remarkable features in the gleyic

unit in horizon 9Bkg is the presence of hard carbonate concretions, around 5 to 10 cm in diameter. These concretions were radiocarbon dated to 5450–5380 cal. B.P. (3240–3110 B.C.) (SOLÍS-CASTILLO et al. 2013).

The upper paleosols (section TB II; Figure 3) formed on top of silty sediment (i.e. on top of section TB III; Figure 3). The paleosols in section TB II (labeled as paleosols 2, 3, 4, 5) show evidence of human occupation in three different periods (according to the Mesoamerican chronology): Formative (800 B.C. – A.D. 150), Classic (A.D. 150–1000) and Post-classic (A.D. 1000–1500). Paleosols 2 and 3 are weakly developed. Horizon 4Bk is a pedosediment, composed of soil fragments and broken carbonate concretions. Paleosol 5 bears the strongest pedogenic features of all paleosols, and can be subdivided in 5Ass, 5Bss and 5BC, with a total thickness >1 m. This paleosol shows strong vertic features: slickensides, hard angular blocky structure, and vertical cracks. Carbonates in the form of white spots and filling fractures and pores are found throughout the entire profile. The modern surface is made up of ~100 cm of alluvial sediment little affected by pedogenesis (C horizon).

The chronology of these paleosols has been previously established by radiocarbon dating (Table 1) and archeological evidence (SOLÍS-CASTILLO et al. 2013). The soil organic matter (SOM) in 5Ass was dated to 2340–2300 cal. B.P. (390–350 B.C.) The carbonates disseminated in 5Bss horizon gave a much younger age, 720–660 cal. B.P. (1230–1290 A.D.). However, artifacts found in this paleosol belong to the Formative Period (1800 B.C.–150 A.D.), which confirms its stratigraphic position. Clearly, the carbonates must have formed later. Charcoal from the 3A horizon yielded an age of 1140–970 cal. B.P. (810–980 A.D.). Classic ceramic fragments (A.D. 150–1000) recovered from this paleosol support this date. The upper paleosol 2 has not been dated with any numerical technique; however it contains Post-classic artifacts (A.D. 1000–1500) and thus allows for assigning an age to it.

The alluvial sediments were best expressed in profile TB III located only 10 m to the east of TB I. At the contact to

Tab. 1: Summary of the dating results. Blue OSL data are from quartz, pIRIR<sub>290</sub> are from K-feldspar. For sample 2464, the quartz is in saturation, i.e. a minimum dose and correspondingly age are given. The radiocarbon ages are taken from SOLÍS-CASTILLO et al. (2013).

Tab. 1: Zusammenfassung der Datierungsergebnisse. Blaue OSL-Daten stammen von Quarz, pIRIR<sub>290</sub> von K-Feldspat. Der Quarz für Probe 2464 ist in Sättigung, d.h. es kann nur eine Mindestdosis und ein -alter angegeben werden. Die Radiokarbonalter stammen aus SOLÍS-CASTILLO et al. (2013).

Lab code	Horizon	Dose rate [Gy/ka]		Equivalent dose [Gy]		Age [ka]		Radiocarbon age cal. BP [2σ]
		quartz	feldspar	blue OSL	pIRIR <sub>290</sub>	blue OSL	pIRIR <sub>290</sub>	
BETA-300447	3A	-	-	-	-	-	-	1140-970 <sup>(3)</sup>
2462	3C	2.90 ± 0.11	-	6.1 ± 1.4 <sup>(1)</sup>	-	2.1 ± 0.5	-	
BETA-300448	5Ass	-	-	-	-	-	-	2340-2300 <sup>(4)</sup>
BETA-300449	5Bss	-	-	-	-	-	-	720-660 <sup>(5)</sup>
2463	TB3_07	3.27 ± 0.11	-	29 ± 6 <sup>(1)</sup>	-	9 ± 2	-	-
BETA-277572	9Bkg	-	-	-	-	-	-	5450-5380 <sup>(5)</sup>
2464	9BCgk	1.98 ± 0.10	2.61 ± 0.11	> 130 <sup>(2)</sup>	322 ± 10 <sup>(2)</sup>	> 65	123 ± 6	-

<sup>1</sup> >24 aliquots measured.

<sup>2</sup> 9 aliquots measured.

<sup>3</sup> Charcoal dated.

<sup>4</sup> Soil organic matter dated.

<sup>5</sup> CaCO<sub>3</sub> nodules dated.

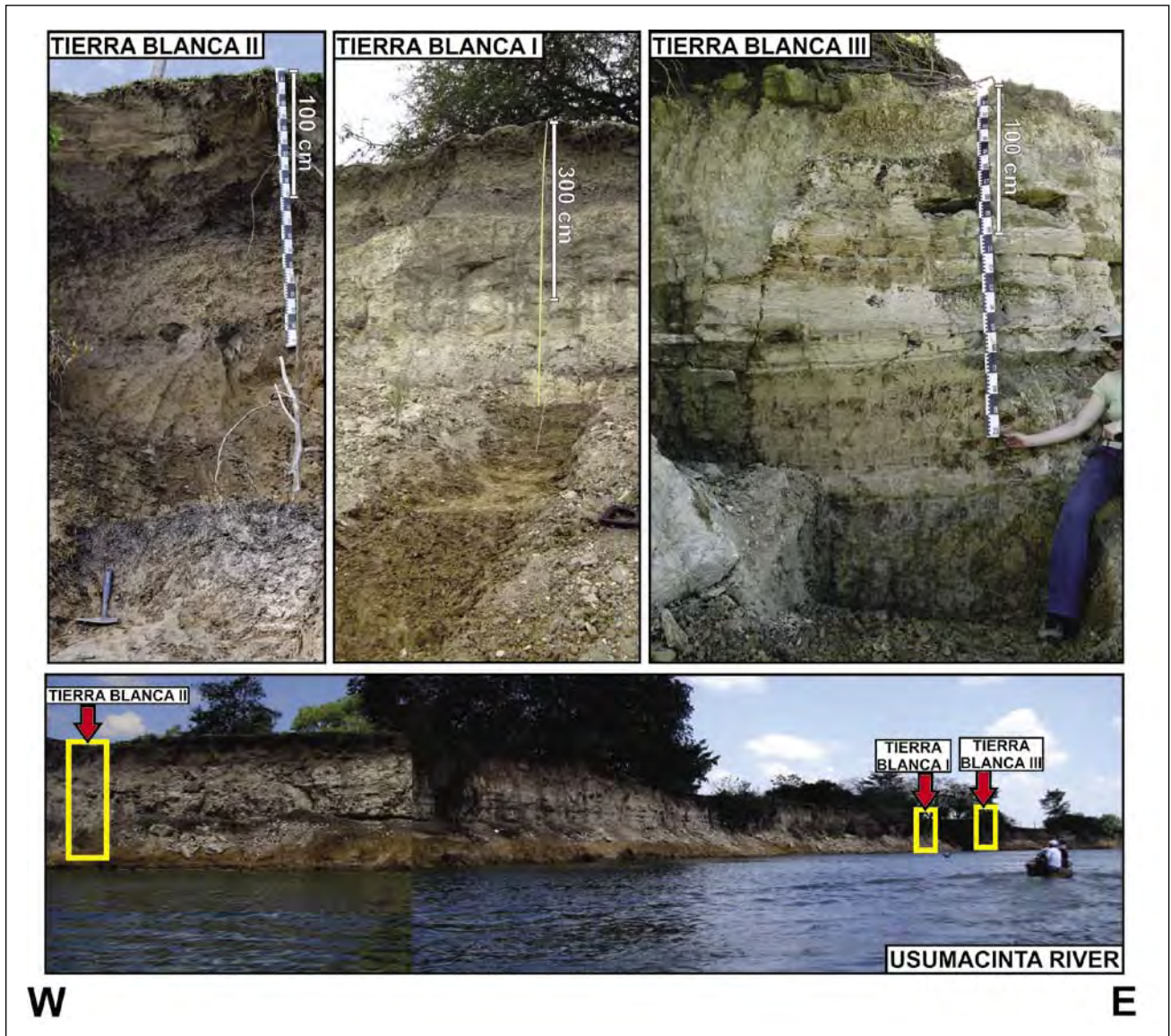


Fig. 2: Photograph showing the positions of the individual profiles in Tierra Blanca.  
 Abb. 2: Übersichtsaufnahme von Tierra Blanca und einzelne Profile.

paleosols with evidence of human occupation, the overlying sediment is laminated and shows a wavy boundary and has a total thickness of 10 cm. The middle part alternates from lamination to crossed stratification and is characterized by a sand texture, reductomorphic features and carbonate concretions. In this section, two phases representing slow water flow or even standing water can be found (as shown by clayey material with some reductomorphic features and root traces).

The upper part is also laminated, with vertical cracks up to 40 cm in length cutting through the sediment layer (cf. Figure 3). The material of all laminated strata reacts intensively with HCl. We refer to this material as silt sediment due to the dominance of this fraction (62%); however, it also has a high proportion of clay (32%) (SOLÍS-CASTILLO et al. 2013).

Another profile situated on top of a Pleistocene terrace is found in the Balancán section (Figure 4). It is characterized by a monogenetic soil (A/Bg/G<sub>1</sub>/G<sub>2</sub>) with strong gleyic features (grayish brown colors with reddish-yellowish-greenish mottles, Fe concretions and dendritic Mn). The soils at Balancán are morphologically similar to the ones

at the base of the Tierra Blanca sequence. However, the Balancán soils are more intensely weathered as shown by the larger amount of clay, and the more compact and pronounced oxide-reduction features (Figure 4).

## 4 Methods

### 4.1 Mineralogy and micromorphology

Thirty-two samples were taken for heavy mineral analyses (cf. Figure 5). The fine sand (63–125 μm) was separated by sieving and pretreated with 10% hydrochloric acid (HCl) to dissolve carbonates, and subsequently with 30% hydrogen peroxide (H<sub>2</sub>O<sub>2</sub>) to remove organic matter. The sands were dried and then floated in sodium polytungstate liquid with a density of 2.82 ± 0.02 g/cm<sup>3</sup>. Heavy minerals were collected in filters and after drying dispersed in Mountex resin with a refraction index of N=1.67. The identification of the minerals was conducted under a polarization microscope after MANGE & MAURER (1992). Two hundred mineral grains were counted for each sample.

We furthermore took three samples representative for the



three types of materials found: the gleyic paleosols at the base (9Bgk horizon, TB I); the paleosols with signs of human occupation (3A horizon, TB II), and the top of the silty sediment (TB3\_00, TB III; cf. Figure 3). Using micromorphology the bulk mineral composition was applied, focusing on the light minerals. The thin-sections (30 µm thick) were prepared from undisturbed soil samples, impregnated at room temperature with resin Cristal MC-40. They were studied under a petrographic microscope, and described following the terminology of BULLOCK et al. (1985). Additionally, the thin section of the silty sediment was scanned with high resolution, 4800 and 9600 dots per inch (DPI) in order to investigate its overall structure.

#### 4.2 Luminescence dating

Luminescence samples were taken by hammering metal tubes into the freshly cleaned profile (cf. Figures 2 and 3); the tubes were sealed to prevent any light intrusion. Additional samples for dose rate determination (dosimetry) were taken from immediately around the tube samples. The luminescence samples were treated under subdued orange light. The material from the outer ends of the tubes was discarded, and the samples were then dried prior to sieving. The fraction 100–150 µm was treated with 30% HCl, 10% H<sub>2</sub>O<sub>2</sub>, and sodium oxalate prior to density separation using sodium polytungstate (quartz:  $\rho < 2.7 \text{ g/cm}^3$  but  $> 2.62 \text{ g/cm}^3$ ; potassium (K)-rich feldspar  $\rho < 2.58 \text{ g/cm}^3$ ). The quartz grains were subsequently etched at least once in 30% hydrofluoric acid (HF) for 1 hour, while the feldspar grains only got a short (20 minutes) HF etch (10%) to remove the outer layer of the grains. Finally all fractions were treated with 30% HCl to destroy any fluorides that might have build up during HF etching.

The equivalent doses ( $D_e$ ) were measured with automated Risø TL/OSL readers as small (samples 2463 and 2464) or medium (sample 2462) aliquots mounted on stainless steel cups. Both infrared (IR) light emitting diodes (LED) and blue LEDs were used. For the quartz extracts the luminescence was detected through a Hoya U340 filter, whilst feldspar detection was through a Schott BG39/Corning 7–59 filter (IR stimulation). All measurements procedures are single aliquot regenerative (SAR) protocols (MURRAY & WINTLE 2000).

Quartz purity checks showed that some aliquots of samples 2462 and 2463 exhibited contamination from IR sensitive material (most likely feldspar), even though the average IR depletion ratio (DULLER 2003) was within 10% of unity. To ensure a quartz signal as pure as instrumentally possible from all aliquots, a double SAR (BANERJEE et al. 2001) was applied using a preheat of 200°C (10 s) and a cutheat of 180°C. IR and blue stimulation were for 100 s at 125°C. At the end of each cycle, a high temperature blue clean-out (280°C for 40 s) was inserted. The quartz of 2464 was measured without the IR step as it was found to be clean (IR depletion within 2% of unity); the preheat temperature was 260°C and the cutheat temperature 220°C, respectively. Early background subtraction (CUNNINGHAM & WALLINGA, 2010) was used to calculate the equivalent doses.

Because the quartz for sample 2464 was found to be in saturation (cf. results section and Table 1), equivalent

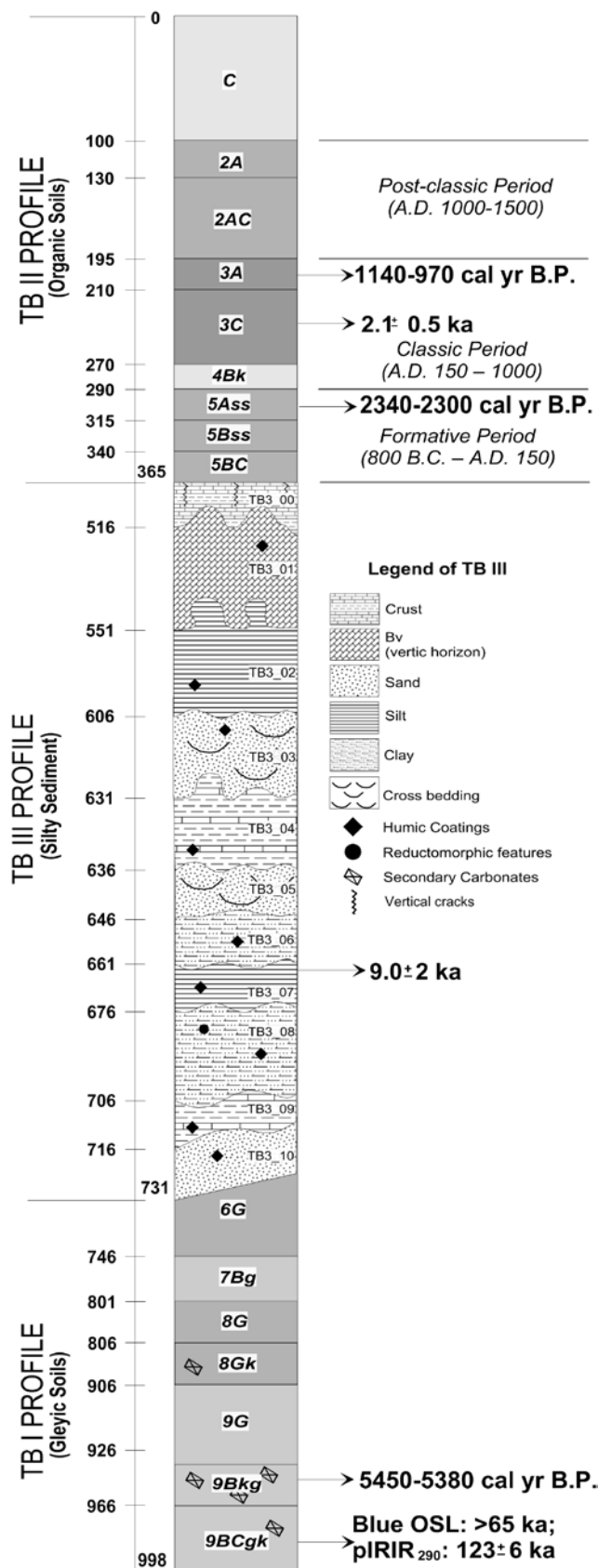


Fig. 3: Tierra Blanca: composite profile and ages.

Abb. 3: Gesamtprofil und Alter für Tierra Blanca.

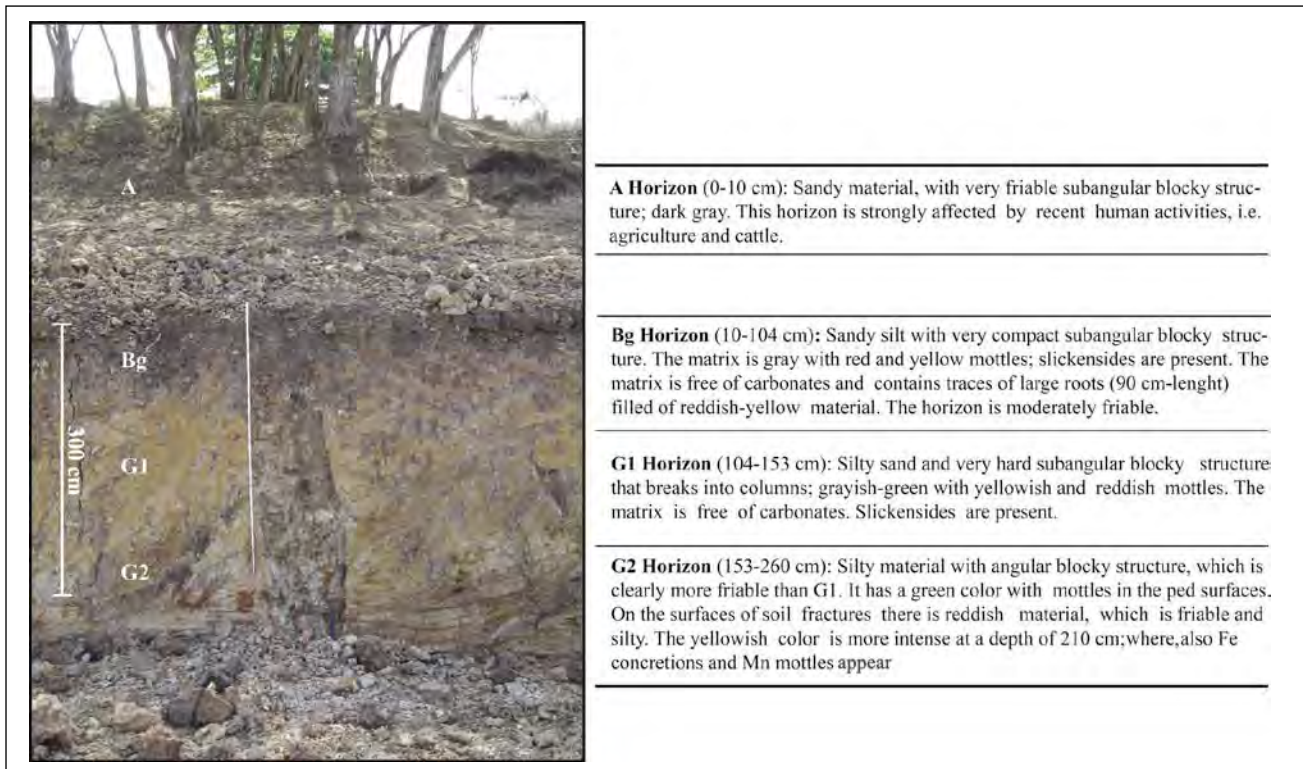


Fig. 4: Balancán: composite profile and field description.

Abb. 4: Gesamtprofil und Geländeaufnahme für Balancán.

doses were also derived from the K-feldspar fraction. A post-IR IRSL protocol was applied using a preheat and cutheat temperature of 320°C, IR stimulation at 50°C (200 s) and post-IR IR stimulation at 290°C (200 s; referred to as pIRIR<sub>290</sub>) (THIEL et al. 2011; BUYLAERT et al. 2012). An IR clean-out at 325°C (100 s) complemented the measurement cycle. The initial 2 s minus a background of the last 40 s were used for equivalent dose calculation.

Dosimetry samples were dried and homogenized prior to packing ~50 g into N-type beakers. High-resolution gamma-counting was conducted at the Leibniz Institute for Applied Geophysics (Hannover, Germany). Conversion to dose rates is based on the values given in GUÉRIN et al. (2011). The cosmic ray contribution was calculated using the data given in PRESCOTT & HUTTON (1994), and for all samples a water content of 25 ± 5% was used. Equivalent doses, dose rates and ages are listed in Table 1.

## 4 Results

### 4.1 Heavy minerals

An overview of the heavy minerals found is shown in Figures 6 and 7. The heavy mineral composition of the paleosols and river sediments of the Usumacinta shows a clear pattern for all studied samples (Figure 5): There is a dominance of opaque minerals. These show signs of weathering (Table 2), indicated by leucoxene whitish rings. These observations are specifically pronounced in the alluvial sediments with concentrations ranging from 14.5% to 59.6%. The samples from the TB II section (the youngest paleosols with strong evidence of human occupation; see Table 2) have the lowest concentration of fresh opaque minerals for all studied sam-

ples (including Balancán). The smallest amount is found in horizons 3A and 3C, where it is about 4.5%. The underlying horizons show values between 22% and 33%. Epidote is the second most abundant mineral (1% to 21%), with a notable increase in the lowest horizons (5Bss, 5BC, 5C). Interestingly, there are high concentrations of pyroxene in the three uppermost horizons (AC-2A-2AC) with values between 12% and 22%; these are the highest concentration of all studied profiles (Table 2). Furthermore, these horizons show the highest concentrations of green (11%) and brown amphibole (5% to 9%). In general green amphibole has greater abundance than brown hornblende. The presence of calcite is more or less constant throughout the profile, with the exception of the uppermost horizons (AC-2A-2AC-3A), where the content ranges from 18% to 55%. Chlorite and zoisite were also found; they do not exceed 4.5%, again with the highest concentrations in the uppermost horizons. The lower part of the profile (4Bk-5As-5Bss-5BC-5C) shows remarkable heavy mineral patterns with relatively high concentrations of rutile (1.3% to 2.5%), chromspinel (3.0% to 5.8%), turmaline (1.1% to 4.0%), kyanite (1.4% to 2.0%) and monazite (2.0% to 2.8%). It is interesting to note that the 4Bk horizon is the only layer with well-rounded minerals, including coarse silt size, and the presence of volcanic glass on the surface of amphiboles.

For the alluvial units, the most striking aspect is the high concentration of volcanic material represented by amphibole, which is coated by volcanic glass and pyroxenes with dissolution surfaces. However, is not possible to identify a general pattern throughout the sequence except for the presence of calcite in percentages between 1% and 59% (the major part located in the middle of the sedimentary sequence). The accumulation of minerals is less homogene-

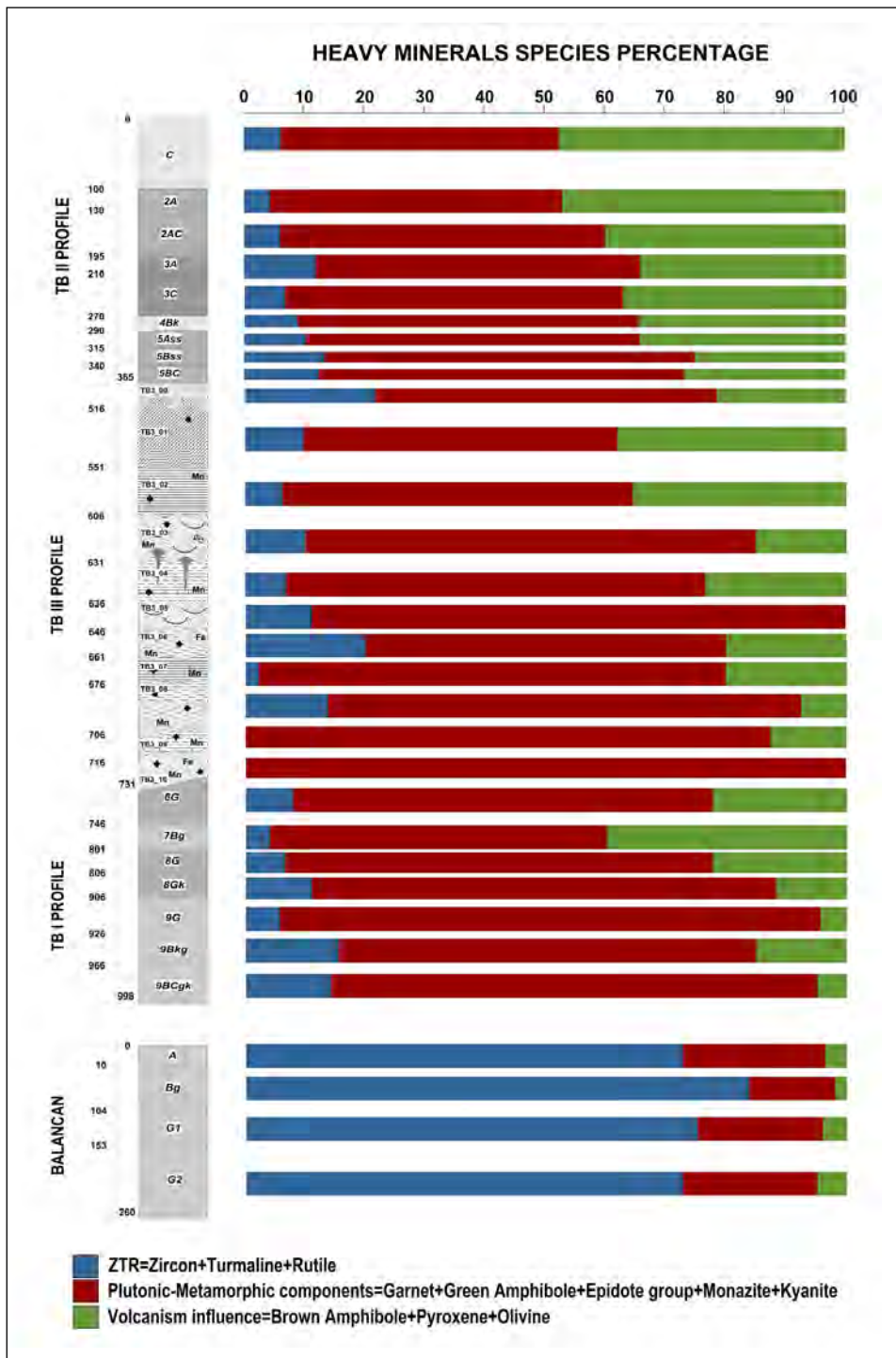


Fig. 5: Tierra Blanca and Balancán profiles and relative abundance of heavy minerals.

Abb. 5: Zusammenstellung der Profile und relative Vorkommen der Schwerminerale für Tierra Blanca und Balancán.

ous compared to the soil profiles. There seems to be a weak tendency of enhanced concentrations for green amphibole, epidote, and pyroxene in the bottom of the profile. Garnet is only of importance in the middle part of the profile (3.4% to 6.0%).

The lower part of the TB profile (TB I; Figure 3) shows a more regular pattern in the mineral concentration. The most important characteristic is the highest presence of epidote ranging from 20.8% to 28.3%, and the concentration of stable minerals (garnet, titanite, zircon, zoisite, rutile, chromspinel, monazite, and kyanite) with a decrease of less stable minerals (like amphibole). Many of these minerals have a metamorphic source (from moderate to high grade metamorphic phases). The influence of the volcanic source is less evident,

however, pyroxene and olivine show large concentrations (mainly in the upper part of the profile, i.e. 6G-7Bg-8G-8Gk). Horizons 6G and 9G show the largest abundance of opaque minerals (Table 2). There is clearly a change in the mineral composition at the boundary between 6G and the lowermost part of the silty sediment (Figure 3), where the metamorphic components dominate. The amount of calcite is much smaller than for TB II and TB III (0.2% to 3.5%).

The Balancán profile shows clear differences in comparison to the Tierra Blanca sections; there ultrastable minerals are present (Figure 5). The sequence shows very high contents of zircon (17% to 25.5%), rutile (3.4% to 8.4%), and titanite (2.0% to 5.3%). There is only very little epidote and hardly any calcite present (Table 2).

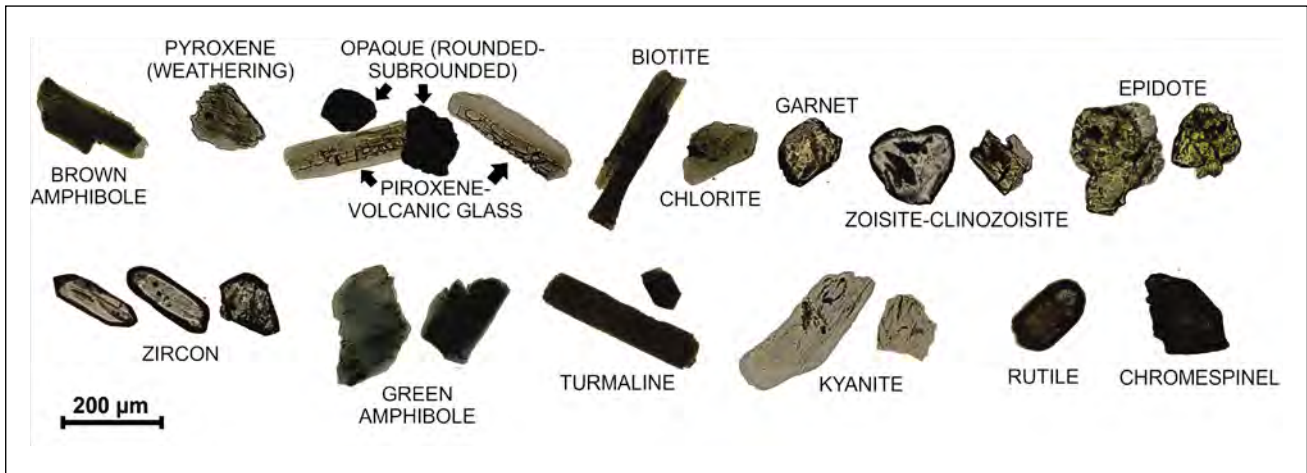


Fig. 6: Heavy minerals under the petrographic microscope using plane polarized light.

Abb. 6: Ausgewählte Schwerminerale unter dem petrographischen Mikroskop unter der Verwendung von eben-polarisiertem Licht.

## 4.2 Micromorphology of selected horizons

Horizon 3A (TB I) is dominated by a clayey-silty texture; there are only a few sand grains. The structure evidences vertic features (angular blocky structure with porostriated b-fabric; Figure 7a). Weathered volcanic glass is also present (Figure 7b). 9Bkg horizon is very clayey but contains some sand grains, which are mainly composed of quartz (Figures 7c and 7d). A few micas are also present; these are all highly weathered. Clay coatings are frequent. In TB III the most remarkable feature in the silty sediment is the strong lamination that is clearly shown not only in the profile (Figure 2c) but also in the scanned section (Figure 7e). This laminated sediment is dominated by volcanic glass which is angular (Figure 7f).

## 5 Discussion

### 5.1 Ages of the paleosol units and pedostratigraphy

The gleyic paleosols (horizons 6 to 9) found in TB I are clearly much older than the overlying succession (TB III and TB II). Horizon 9BCgk was luminescence dated to >65 ka (quartz OSL; sample 2464); the feldspar ( $\text{PIRIR}_{290}$ ) yielded an age of  $123 \pm 6$  ka (Table 1). This age represents the age of the parent material of paleosol 9, i.e. the paleosols are younger. One might argue that alluvial sediments are not well-bleached prior to deposition, i.e. the luminescence signal is not reset, which would result in an age overestimate. It is very unlikely that the alluvial sediments at this site carry a large residual dose, which is shown by the youngest sample (horizon 3C; sample 2462): Its luminescence age is in very good agreement with the radiocarbon ages of the over- and underlying layers (Table 1). The radiocarbon age of 5450–5380 cal. B.P. for the lowermost paleosol (horizon 9Bkg; Table 1) was derived from neoformed carbonates; they postdate the age of the sediments. The gleyic paleosols show strong redoximorphic features, leaching (carbonates are absent in the groundmass) and intense clay accumulation (including well preserved clay coatings). All these characteristics point to pedogenesis in a moist environment, sometimes accompanied by water-logging that is hardly compatible with the formation

of calcite. Thus, carbonates precipitated much later, after environmental conditions changed. As a consequence, the age of the carbonate concretions (5450–5380 cal. B.P.) cannot be interpreted as the age of the first intensive pedogenetic phase. According to the luminescence ages of the parent material and of the overlying alluvial sequence (sample 2463; cf. Table 1) the soils formed after  $123 \pm 6$  ka but prior to  $9 \pm 2$  ka.

The latter age is the only age estimate for the silt sediment between the two pedocomplexes (TB I and TB II) (cf. Figure 3). The deposition thus took place during the Holocene. The underlying pedocomplex had a very long time for its development (i.e. from  $123 \pm 6$  ka until the Holocene). The presence of several phases of sedimentation and soil formation (to constitute each paleosol in the pedocomplex) take a considerable time span, especially if one takes into account the weathering degree of each paleosol, which is higher than for the soils of the overlying sequence.

However, one has to consider that the migration of the river system might have caused erosion of parts of the Late Pleistocene sequence. Evidence of such erosional phases are as follows: None of the Gleysols have A horizons, and there are changes in the heavy mineral distribution, which clearly reflects discontinuities (Figure 5). Such discontinuities are obvious between 8G/7Bg; 7Bg/6G, and 6G/silty sediment (the latest is the most contrasting).

SOLÍS-CASTILLO et al. (2013) have presented a chronological framework based on radiocarbon and archeological evidence for the younger paleosols (TB II). They presented an age of 2340–2300 cal B.P. for soil horizon 5Ass (Figure 3), and they found ceramic belonging to the Formative culture, which agrees with the dating result. The presence of Vertisols, which formed during the same period, has been documented in other areas of the Mayan Lowlands (DAHLIN et al. 1980; POPE & DAHLIN 1989, 1993; DUNNING & BEACH 2004; BEACH et al. 2006; DUNNING et al. 2006).

Horizon 3A was radiocarbon dated to 1140–970 cal. B.P. (SOLÍS-CASTILLO et al. 2013). The luminescence age of  $2.1 \pm 0.5$  ka for horizon 3C fits well into this chronological framework, and can give some insights about the period of relative landscape stability. The age from the 3C horizon represents the time when the sediment was deposited, that is the maxi-

Tab. 2: Summary of the heavy mineral analysis results (in percentage).

Tab. 2: Zusammenfassung der Ergebnisse der Schwermineralanalyse (angegeben in Prozent).

Horizon	Volcanic		Plutonic-Metamorphic					ZTR [ultra stable]			Opaque Minerals	
	Brown amphibole	Pyroxene	Garnet	Green amphibole	Epidote group	Monazite	Kyanite	Zircon	Turmaline	Rutile	Weathered	Fresh

Tierra Blanca II

C	8.95	17.02	2.28	11.05	8.07	0.00	0.00	2.28	0.00	1.05	25.26	0.00
2A	5.32	22.20	1.28	11.19	13.03	0.00	0.00	0.73	0.73	1.10	22.02	0.00
2AC	6.52	11.86	7.31	2.17	12.85	0.00	0.59	1.38	0.20	1.19	26.68	0.99
3A	0.00	0.00	0.25	0.75	1.01	0.00	0.00	0.25	0.00	0.00	4.52	5.03
3C	1.67	0.83	0.83	3.33	2.92	0.00	0.00	2.08	0.42	0.00	4.58	0.83
4Ck	1.42	3.30	6.60	5.19	10.85	0.00	0.47	2.36	0.94	0.00	33.02	3.30
5Ass	7.17	7.89	2.15	2.15	10.04	2.51	1.43	6.45	3.94	0.72	27.60	6.09
5Bss	6.31	4.32	1.99	1.33	20.60	1.99	1.99	3.99	1.66	1.33	24.25	6.31
5BC	6.20	5.43	4.65	1.94	17.83	1.94	0.00	3.49	1.16	1.16	29.07	4.26
5C	4.13	6.06	3.31	2.20	20.94	2.75	0.83	6.61	3.31	2.48	27.00	2.75

Tierra Blanca III

TB3_0	11.01	12.50	0.89	7.74	20.83	1.79	1.19	2.68	2.38	1.19	24.40	1.49
TB3_1	1.74	6.67	2.32	2.32	7.83	0.58	0.00	1.16	0.00	0.29	19.13	7.83
TB3_2	2.54	0.00	1.69	4.24	5.93	0.00	0.00	1.69	0.00	0.00	50.00	16.10
TB3_3	0.78	7.03	1.56	4.69	14.06	0.78	0.00	2.34	0.00	0.00	30.47	17.19
TB3_4	0.00	0.00	0.81	2.42	3.23	0.00	0.00	0.81	0.00	0.00	14.52	16.13
TB3_5	6.78	3.39	3.39	13.56	11.86	0.00	0.00	8.47	1.69	0.00	18.64	3.39
TB3_6	7.69	3.85	8.97	19.23	15.38	0.00	0.00	1.28	0.00	0.00	23.08	2.56
TB3_7	1.82	1.82	5.45	12.73	18.18	0.00	0.00	3.64	3.64	0.00	14.55	1.82
TB3_8	1.75	1.75	0.00	3.51	21.05	0.00	0.00	0.00	0.00	0.00	59.65	7.02
TB3_9	0.00	0.00	3.09	0.00	6.79	0.00	0.00	0.00	0.00	0.00	61.11	27.16
TB3_10	3.55	8.28	2.37	18.93	14.20	1.79	0.00	1.78	2.37	0.00	37.87	4.14

Tierra Blanca I

6G	2.49	7.21	2.62	0.98	22.62	0.00	0.00	2.30	1.49	0.50	25.87	0.50
7Bg	0.66	2.53	2.11	0.84	28.27	0.00	0.33	2.11	0.33	0.33	37.70	4.92
8G	0.84	2.70	1.96	0.25	26.72	0.98	0.00	2.45	0.00	0.42	38.40	9.28
8Gk	1.96	1.22	6.40	1.22	28.96	1.69	0.00	0.91	0.25	1.96	44.61	0.98
9G	0.00	5.38	6.01	0.95	20.89	1.72	0.00	5.70	0.30	1.22	36.89	10.37
9Bkg	1.90	0.70	6.32	1.75	23.51	0.61	2.85	5.26	0.32	1.58	25.63	13.29
9BCgk	1.05	7.21	2.62	0.98	22.62	0.95	0.35	2.30	0.00	1.05	32.28	4.56

Balancan

A	1.55	0.00	0.00	7.75	1.55	0.78	0.00	24.42	0.39	7.36	44.96	3.88
Bg	0.62	0.00	0.00	0.62	2.80	0.93	0.62	20.81	0.00	8.39	25.16	30.12
G <sub>1</sub>	0.76	0.38	0.00	1.14	2.28	1.14	1.14	19.77	0.00	3.42	46.77	10.27
G <sub>2</sub>	1.06	0.53	0.00	3.19	2.66	1.06	0.53	17.02	1.60	6.38	48.40	6.38

imum age of the soil. In contrast, the date obtained from the bulk organic matter of 3A represents the minimum age of the paleosol, just prior to its burial. These ages show that the soil formed within 1000 years (the difference between the age of 3C and 3A horizons: beginning and end of pedogenesis).

## 5.2 Provenance and landscape evolution: Mineralogical evidences

Mineralogical assemblages are similar among all studied profiles (Figure 5). However, in the upper units, higher amounts of volcanic minerals can be detected. It is clear

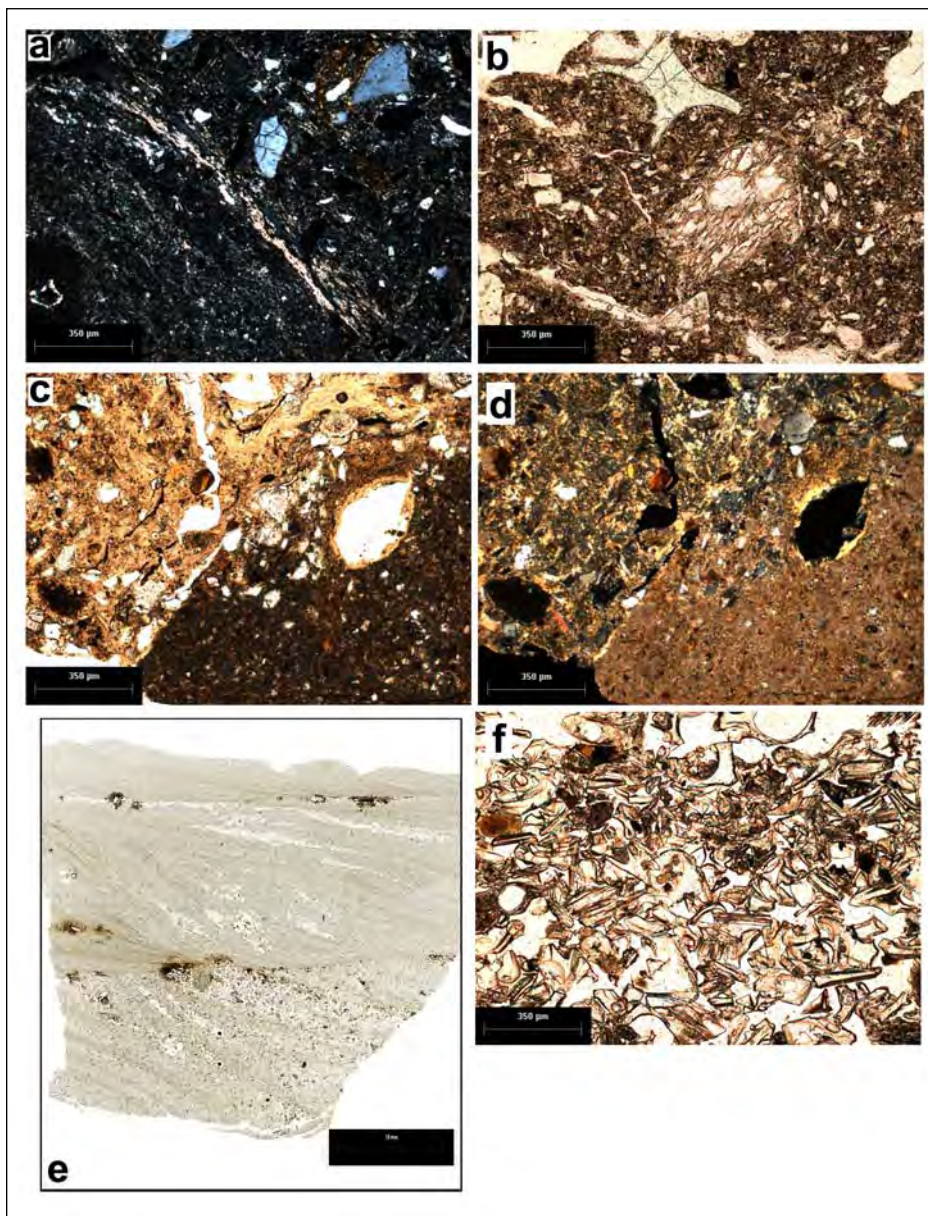


Fig. 7. Micromorphology of selected horizons: (a) porostriated b-fabric in horizon 3A TB I, cross polarized light (XPL); (b) weathered pumice in horizon 3A TB I, plane-polarized light (PPL); (c) Clay coatings and pedogenetic carbonates in 8Bkg horizons (PPL); (d) Clay coatings and pedogenetic carbonates in 8Bkg horizons (XPL); (e) silty sediment from TB III; (f) volcanic glass in the silty sediment of TB III (PPL).

Abb. 7: Mikromorphologie von ausgewählten Horizonten: (a) poro-streifiges b-Fabric aus Horizont 3A TB I, gekreuz-polarisiertes Licht (XPL); (b) verwitterter Bims aus Horizont 3A TB I, eben-polarisiertes Licht (PPL); (c) Toncutane und pedogenes Karbonat aus Horizont 8Bkg (PPL); (d) Toncutane und pedogenes Karbonat aus Horizont 8Bkg (XPL); (e) siltiges Sediment aus TB III; (f) vulkanisches Glas im siltigen Sediment aus TB III (PPL).

that the active volcanism in the Central America Volcanic Arc can provide such minerals to the alluvial plain of Tabasco (Figure 1). It is interesting to note that these minerals have angular and subangular shapes suggesting transport by slow river flow. It is likely that they originate from ash falls coming directly from the primary source.

One of the most intriguing materials is the silty sediment in the upper part of TB III (TB3\_00). Its sand fraction contains a mixture of volcanic, metamorphic, and plutonic minerals, but in the thin sections, we clearly observe the dominance of laminated volcanic materials, particularly, volcanic glass (Figure 7). Based on the mineralogical data and an age of  $9 \pm 2$  ka of the sediment, in which the minerals were found, we attempt to correlate the mineral assemblage to a possible eruption in the surrounding area. According to ARCE (personal communication), it is likely that the Tacana volcano (Figure 1) provided the volcanic material found in the sediment. This volcano had large voluminous eruptions during the Late Pleistocene (ARCE et al. 2012). This material can subsequently be reworked over short distances by the river. However, a detailed analysis is needed to draw any

final conclusions on the provenance of this sample.

The heavy mineral associations (Figure 8) of the investigated sediments imply a broad range of metamorphic source rocks, suggesting the influence of different tectonic environments. Zircon-Turmaline-Rutile (ZTR) suit minerals (HUBERT 1962) are common in acidic to intermediate granitoid rocks as well as in mature siliciclastic sediments and some metamorphic rocks (e.g. VON EYNATTEN & GAUPP 1999). The highest concentrations of ZTR in the Balancán profile can be interpreted as originating from an ancient sedimentary recycling phase, where the dissolution and pedogenesis reduces the less stable minerals (GARZANTI & ANDO 2007). It is interesting to draw attention to the origin of the rutile phases: The majority of detrital rutile comes from medium to high grade metamorphic rocks (FORCE 1980; 1991) and recycled sediments. The same mineralogical evidence is found in the metamorphic complex of Southeast Guatemala (garnet, kyanite, rutile schists, and abundant ortho-gneisses, which range from mafic to granitic); this is where the Usumacinta River starts. In consequence, our data suggest that the river transported material from this region.

Additionally, the dominance of euhedral tourmaline indicates proximity to the source rocks and short transport from a predominantly low-grade metamorphic source (SINGH et al. 2004). Chlorite and epidote were also derived from low-grade metamorphic series, whereas gneisses, granitoid rocks and recycled sedimentary rocks are possible sources of (ultra) stable minerals like zircon, tourmaline and rutile. Kyanite indicates the presence of high-pressure metamorphic rocks in the catchment; it has to be noted that it is not a dominant mineral in the assemblage. SOLARI et al. (2011) have documented areas in Central Guatemala with high-pressure metamorphic rocks. As this area is cut northwards by the Chixoy River until it joins with the Passion River (a tributary of the Usumacinta; Figure 1), it is very likely that this is the sediment source for the alluvial plain.

The presence of chromspinel and epidote group minerals are indicative for a source area with ophiolite complexes associated with suture zones. The most important suture zone near the study area is in Guatemala where the Usumacinta River springs (Figure 1). At present, this suture zone is part of a major left lateral strike slip boundary between the North American plate to the north (locally known as the Maya block) and the Caribbean plate (locally the Chortís block) to the south (BRUECKNER et al. 2009).

The units in the Mayan block include the Chuacús metamorphic complex, recently shown to include eclogitic lenses that record a Late Cretaceous event (ORTEGA-GUTIERREZ et al. 2007; MARTENS et al. 2007, 2010), Paleozoic sedimentary rocks of the Santa Rosa Group, low-grade meta-sediments (white mica-chlorite schists, quartzite, and minor marble) associated with antigorite schist/mélange and deformed granites. The Chortís block contains the green schist facies, San Diego phyllite, the amphibolites facies Las Ovejas complex, and relatively undeformed granitoids. Numerous U-Pb zircon ages from granites and their host gneisses north of the fault resulted in Late Proterozoic (Grenville), Carboniferous, and Triassic ages as well as a metamorphic event at c. 70 Ma (MARTENS et al. 2007, 2010).

### 5.3 Implication for human land use

Our data can provide some insight into the land use of the region, especially with respect to fluctuations of human occupation and to the ceramic production. It is obvious that the presence of well-drained soils in Tierra Blanca (TB II) favored human occupation. These paleosurfaces were not inundated and represented stable areas for human activities, which were more or less continuous throughout the last 2–3 millennia. In contrast, other archeological sites such as Palenque, even though bigger in size, had less continuous occupation (for details see LIENDO et al. 2013).

Further, it is interesting to note that the suite of heavy minerals in paleosols and sediments found in Tierra Blanca has also been documented in petrographic studies about the pottery of the region (OBANDO et al. 2011). It can therefore be deduced that the material at Tierra Blanca is very suitable for the manufacturing of the pottery. The gleyic paleosols at the base provided clay for the clay matrix of the ceramic, while the volcanic glass found in the silty sediment (TB III) could be used as temper. Thus, this area has been of great importance to the humans at this time.

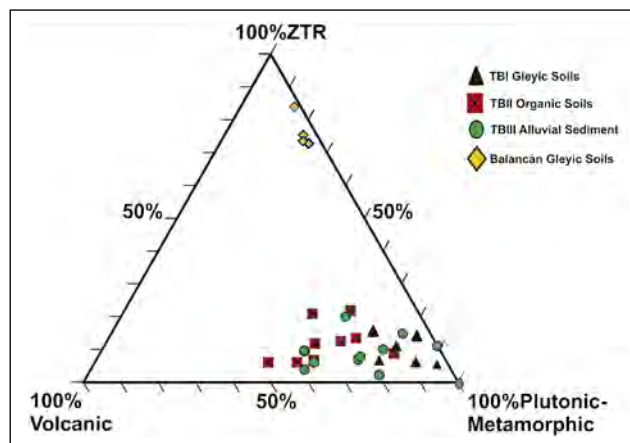


Fig. 8: Ternary diagram with the relative mineralogical suites of heavy minerals. ZTR = zircon + tourmaline + rutile; Plutonic-Metamorphic = garnet + green amphibole + epidote group + monazite + kyanite; Volcanic = brown amphibole + pyroxene + olivine.

Abb. 8: Phasendiagramm für die mineralogischen Charakter der Schwerminerale. ZTR = Zirkon + Turmalin + Rutil; plutonisch-metamorph = Granat + grüner Amphibol + Epidotgruppe + Monazit + Kyanit; Vulkanisch = brauner Amphibol + Pyroxen + Olivin.

## 6 Conclusions

The alluvial terrace system of the Usumacinta River reflects the Late Quaternary pedostratigraphy of the region. On the alluvial Pleistocene terrace, soils with gleyic features are present. On the oldest Holocene terrace (HT2), we found the remnants of the gleyic paleosols in sediment dated to  $123 \pm 6$  ka (pIRIR<sub>290</sub>), while on the younger terraces (HT1, HT0) only less developed paleosols are present.

Metamorphic and plutonic terrains of the south of Mexico and Guatemala were identified as the main source of sediments for the Tierra Blanca area. However, the heavy mineral assemblages reflect some changes in provenance throughout time. Soils located on the older terraces, as in Balancán, show the presence of ultra-stable minerals; these reflect prolonged landscape stability and soil formation. In consequence, this mineralogical set can be used for pedostratigraphic correlation and tracing the oldest terraces in the area.

Larger amounts of volcanic minerals in the younger paleosols may allow for pointing out areas which were affected by eruptions. Further, the mineral compositions, if compared with those of ceramics might make it possible to trace the source regions for ceramic production.

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# Holozäne Landschaftsentwicklung an der Westküste der Nordseeinsel Amrum

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**Kurzfassung:** Barriereinseln und Außensände, geformt durch eine Kombination aus Wind, Wellen, Strömung und Küstenlängstransport gelten als morphologisch hoch aktive Küstenbereiche und variieren häufig in Ursprung, Genese und Entwicklung. Sie besitzen durch ihre dissipative Wirkungsweise eine bedeutende Schutzfunktion für rückwärtige Inseln, Halligen und Festlandbereiche und bilden vor der Küste Nordfrieslands die westliche Außengrenze des Wattenmeeres.

Ziel der nachfolgenden Studie ist es, anhand hochauflösender Georadarmessungen und sedimentologischer Daten aus Bohrungen die Landschaftsgeschichte an der Westküste Amrums sowie die Entstehung und interne sedimentäre Architektur des der Insel westlich angelagerten Kniepsandes zu untersuchen und die Prozesse, die zur Genese geführt haben, durch Datierungen zeitlich einzuordnen. Auf Grundlage der gewonnenen Daten wurden zwei stratigraphische Modelle entwickelt, welche die geologisch-geomorphologischen Prozesse und Sedimentationsbedingungen im Westküstenvorfeld erklären.

Ein Modell zeigt die Landschaftsentwicklung an der Westküste Amrums und beschreibt die Sedimentationsbedingungen, die im Vorfeld der Westküste herrschten, bevor der Kniepsand an die Insel heranwanderte. Auf der Landoberfläche des ertrinkenden saaleiszeitlichen Geestkerns wurden zu Beginn der Flandrischen Transgression feinkörnige Sedimente eines Misch- und Schlickwatts abgelagert. Es ist davon auszugehen, dass der damals noch weit vor der Küste Amrums liegende Kniepsand eine Barriere bildete und so an der heute hochenergetischen Westküste für strömungsberuhigte Sedimentationsbedingungen sorgte. Durch Erosion am Geestkern bildeten sich in unterschiedlichen Höhenpositionen fossile Kliffs, die dem damaligen Meeresspiegelstand entsprechen.

Ein weiteres Modell beschreibt den Andockmechanismus des Kniepsandes an die Insel Amrum. Durch die Anlagerung des ehemaligen Außensandes und den damit einhergehenden Sedimentinput wurden die Bedingungen für eine großflächige Dünenbildung geschaffen.

## Holocene landscape development in front of the west coast of Amrum (North Sea/ German Bight) – Insights from ground-penetrating radar surveys and sedimentological data

**Abstract:** Barrier sandbars and barrier islands are geologically young, highly dynamic and represent a complex coastal system that includes a number of different but closely related sedimentary depositional environments with geomorphologic elements of varying origin, genesis and evolution. Barrier sandbars are exposed ridges of sand that are built offshore by wave action.

The so-called Kniepsand on the island of Amrum represents one of the widest beaches in Europe. The Kniepsand originally belongs to a system of sandbars lining the North-Frisian coast. Some of these sandbars are already attached to the mainland (e.g. St. Peter-Ording-Sand and Westerhever-Sand). The offshore sandbars of Japsand, Norderoogsand and Süderoogsand are located in front of the western coastline of the North-Frisian Island and the Halligen and fulfil the function of natural coastal defence dissipating the energy of the incoming deep-water waves of the North Sea.

Barrier sandbars are usually investigated through the use of aerial photos and borehole data. Therefore, the processes of evolution, migration and the internal structure of sandbars are often unknown. That is why this study chooses an integrated approach using high-resolution ground-penetrating radar (GPR) and sedimentological analyses of shallow sediment cores drilled at selected sites along the radar profiles. A geophysical Survey Systems Inc. radar system, SIR-2000 coupled with a 200 MHz antenna, was used.

Based on these data two sedimentary models were generated. One model describes the process of barrier sandbar migration and the attachment to the Pleistocene island core. According to historical maps and nautical charts of the sixteenth and seventeenth century, the Kniepsand used to be a solitary barrier sandbar located well in front of the island's west coast without any connection to its Pleistocene core consisting of Saalian moraine deposits. The presented model shows how the Kniepsand has welded onto the island core of Amrum.

Before the barrier sandbar was connected to the island, conditions for sedimentation had been quite different. Tidal flat deposits had been accumulated in a low energy environment. Tidal flat deposits show a general coarsening upward trend and turn into overlying coarser grained beach deposits. Old cliffs formed through several storm surges are also preserved in GPR data. The landscape evolution in front of the westcoast of Amrum was summarized in the second model.

**Keywords:** *Holocene landscape development, barrier sandbar migration, barrier island stratigraphy, ground-penetrating radar (GPR), Amrum, North Sea/German Bight*

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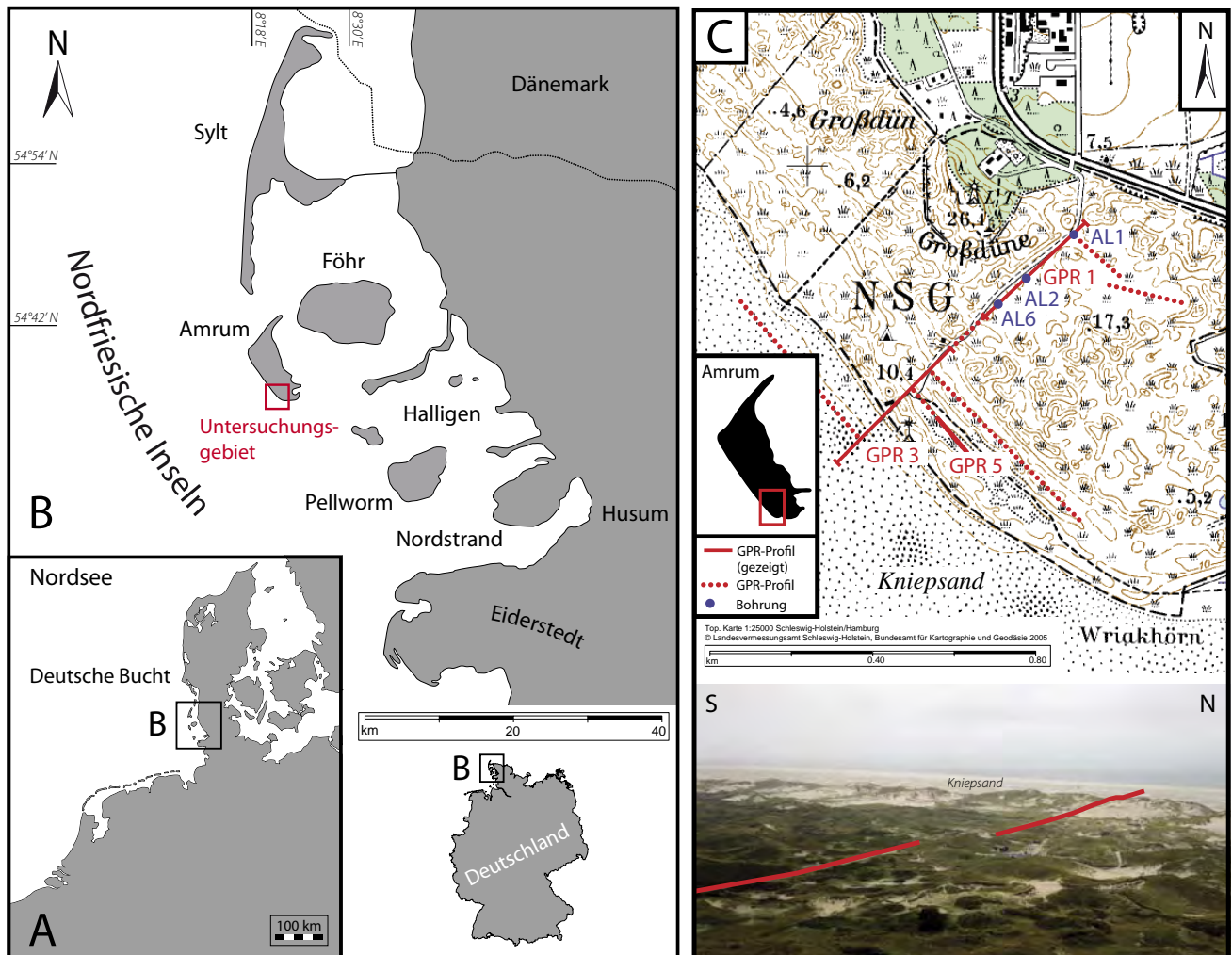


Abb. 1: Überblick über das Untersuchungsgebiet. A: Lage der Nordfriesischen Inseln innerhalb der Deutschen Bucht. B: Lage des Untersuchungsgebietes nahe des Leuchtturms der Insel Amrum. C: Position der GPR-Profile und Bohrungen im Untersuchungsgebiet.

Fig. 1: Overview of the study area. A: German Bight with the position of the Northfrisian Islands. B: Position of the study area next to the lighthouse of Amrum Island. C: Position of GPR-profiles and cores in the study area.

## 1 Einleitung

Barriereinseln und Außensände, geformt durch eine Kombination aus Wind, Wellen, Strömung und Küstenlängstransport gelten als morphologisch hoch aktive Küstenbereiche und variieren häufig in Ursprung, Genese und Entwicklung. Durch ihre dissipative Wirkungsweise besitzen sie eine bedeutende Schutzfunktion für die rückwärtigen Inseln, Halligen und Festlandbereiche und bilden so vor der Küste Nordfrieslands die Außengrenze des Wattenmeeres.

Bisherige Untersuchungen zur Stratigraphie dieser Sedimentationsräume basieren häufig allein auf Bohrungen. Daher sind die interne sedimentologische Struktur und die Prozesse, die zur Entwicklung und Wanderung von Barriereinseln und -sänden geführt haben, oftmals unzureichend untersucht.

Ziel der folgenden Studie ist es daher, anhand hochauflösender geophysikalischer und sedimentologischer Daten die Landschaftsgeschichte an der Westküste Amrums zu rekonstruieren sowie die interne sedimentäre Architektur des der Insel westlich angelagerten Kniepsandes zu untersuchen. Neben der Rekonstruktion der Landschaftsentwicklung sind vor allem die Sedimentationsbedingungen, die an

der Westküste der Insel Amrum geherrscht haben, bevor sich der Kniepsand an die Insel anlagerte, von besonderem Interesse. Dabei stehen folgende Fragen im Vordergrund:

- Wie hat sich der Kniepsand als ehemaliger, weit vor der Westküste Amrums liegender Außensand, an die Insel anlagert?
- Welche Landschafts- und Sedimentationsbedingungen herrschten an der Westküste Amrums bevor der Kniepsand sich an die Insel anlagerte?
- Welche morphologischen Prozesse und Bedingungen haben zum Heranwandern und Andocken des ehemaligen Außensandes an den Inselgeestkern geführt?
- Inwieweit ist es möglich mit Hilfe hochauflösender Georadardaten die ursprüngliche Andockstelle des Kniepsandes an den Inselgeestkern zu definieren und rekonstruieren?

## 2 Untersuchungsgebiet

Die Insel Amrum in der südlichen Nordsee (Deutsche Bucht) bildet zusammen mit der nördlichen Nachbarinsel Sylt, den dänischen Inseln Fanø, Mandø, Rømø und den nordfriesischen Außensänden Japsand, Norderoog- und

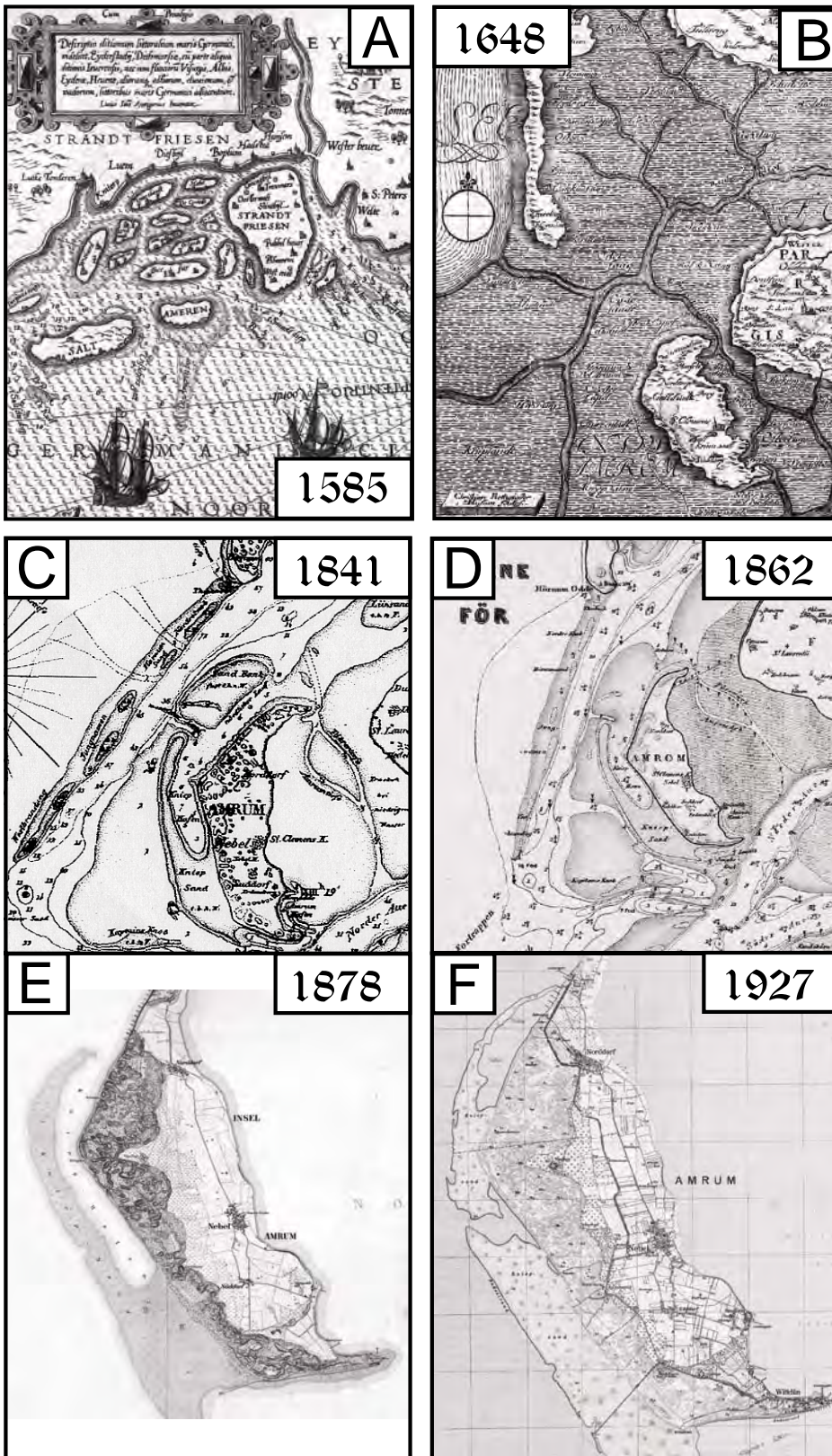


Abb. 2 A: Der Ausschnitt aus der Seekarte der Deutschen Bucht von W. J. Waghenaer aus dem Jahr 1585 mit Blickrichtung nach Osten zeigt die Insel Amrum mit vorgelagerter Sandbank „Ameren Bor“. B: Ausschnitt aus der Landkarte des Amt Tondern aus dem Jahre 1648 nach Johannes Mejer verändert nach Danckwerth (1652). Westlich vor der Küste Amrum ist der „Kniepsandt“ erstmals namentlich erwähnt. C: Ausschnitt aus einer Seekarte der Helgoländer Bucht von Zahrtmann aus dem Jahre 1841 (aus Lang 1973). D: Ausschnitt aus einer Seekarte der Helgoländer Bucht von Zahrtmann aus dem Jahre 1862 (aus Lang 1973). E: Ausschnitt aus einer topographischen Karte der Königl. Preuss. Landes-Aufnahme (Maßstab 1:25000) aus dem Jahre 1878. F: Ausschnitt aus einer topographischen Karte (Meßtischblatt 199 im Maßstab 1:25000) aus dem Jahre 1927.

Fig. 2 A: The map section of the German Bight of 1585 from W. J. Waghenaer shows the island of Amrum with the sandbar „Ameren Bor“. B: Map section of Tondern of 1648 from Johannes Mejer modified by Danckwerth (1652). The „Kniepsandt“ is named to the west of Amrum’s coast for the first time. C: Map section of the Helgoland Bight from Zahrtmann (1841) in Lang (1973). D: Map section of the Helgoland Bight from Zahrtmann (1862) in Lang (1973). E: Map section of a topographic map of the year 1878. F: Map section of a topographic map of the year 1927.

Süderoogsand im Süden eine Kette aus Barriereinseln, die als westliche Außengrenzen des Wattenmeeres der dänischen und schleswig-holsteinischen Festlandsküste vorgelagert sind. Von der Wortbedeutung leitet sich der Name Amrum von „Am Rem“ ab, das soviel bedeutet wie „sandiger Rand“ (QUEDENS 2008) und sich morphologisch auf den der Insel Amrum im Westen vorgelagerten Kniepsand bezieht.

Geologisch gehört Amrum zu den Nordfriesischen Geestk-

erninseln und weist als kleinste dieser Inseln eine Nord-Süd-Erstreckung von knapp 10 km und eine Fläche von rund 20 km<sup>2</sup> auf. Wird der Kniepsand, ein heute bis zu 1,5 km breiter Sandstrand, der sich als ehemaliger Außensand entlang der Westküste erstreckt, mit zu der Inselfläche gerechnet, misst die Gesamtfläche 30,5 km<sup>2</sup>.

Das Untersuchungsgebiet befindet sich im Südwesten der Insel und umfasst das Dünengebiet südlich der 26,4 m ho-

hen „Großdüne“ auf der 1875 der Leuchtturm (54°37'52,2" N/ 8°21'16,5" E) errichtet wurde (Abb. 1). An die Dünen schließt sich im Südwesten der Kniepsand an.

Der pleistozäne Geestkern der Insel Amrum nimmt heute mit 13,7 km<sup>2</sup> etwa 2/3 der Inselfläche ein (REMDE 1972). Im Zuge der Flandrischen Transgression zu Beginn des Holozäns griff die Brandung des ansteigenden Meeresspiegels den Geestkern Amrums von allen Seiten an und schuf durch Abrasion an der gesamten Westküste und an vorspringenden Teilen der Ostküste ein Kliff. Auf diese Weise entstand an der Westküste Amrums auf einer Strecke zwischen Wriakhörn und Norddorf das Littorina Kliff. Im Gegensatz zum Steenodder Kliff an der Wattseite ist das Littorina Kliff an der Westküste rezent durch den Kniepsand und die vorgelagerten Dünenketten vor der Brandung geschützt und so zum fossilen Kliff geworden.

Eine Besonderheit hinsichtlich der Morphologie der Insel Amrum stellt der Kniepsand dar, der auf 16,5 km Länge entlang der Westküste von Wittdün im Süden bis nach Norddorf im Norden reicht. Mit bis zu 1,5 km Breite zählt er heute zu den breitesten Stränden Europas. Die rund 10 km<sup>2</sup> große Sandfläche weist im Mittel eine Höhe von NN +1,75 m auf und wird nur noch während winterlicher Sturmfluten vom Meer überflutet.

Aus geologischer Sicht ist der Kniepsand kein echter Bestandteil der Insel Amrum. Im weitesten Sinne handelt es sich um eine ehemalige der Westküste Amrums vorgelagerte Sandbank, die sich als einstiger Außensand der Inselküste im Westen angeschlossen hat. Der Kniepsand gehört damit zu einer Reihe von Außensänden und Sandbänken, die den Inseln und Halligen der Nordfriesischen Küsten im Westen vorgelagert sind. Einige dieser ehemaligen Außensände wie beispielsweise der Sand vor St. Peter-Ording und der weiter nördlich gelegene Westerhever-Sand sind bereits komplett an die Küste heran gewandert und stellen die heutigen Strände der dortigen Festlandsküste dar. Andere Außensände wie zum Beispiel der Norderoog- und Süderoogsand oder der deutlich kleinere Japsand westlich der Hallig Hooge liegen noch heute im *offshore* Bereich und dienen so den rückwärtigen Inseln und Halligen während Sturmfluten als natürlicher Wellenbrecher.

Einzelheiten hinsichtlich der Entstehung sowie der Herkunft des Kniepsandes sind bis dato nicht bekannt. Als erster verzeichnete ihn Mejer in seiner Karte aus dem Jahre 1648 (Abb. 2 Bild B), wo einige Kilometer westlich der Insel Amrum ein „Kniepsandt“ eingetragen ist. Dieser stellte vermutlich zu dieser Zeit für die Seefahrt eine Untiefe dar.

In historischen Seekarten aus dem 16. und 17. Jahrhundert, wie z. B. der Waghenaer-Karte aus dem Jahr 1585, ist der Kniepsand als solitäre Sandbank mit der Bezeichnung „Ameren bor“ (Amrumer Barriere) eingetragen (vgl. Abb. 2 A), die der Westküste von Amrum vorgelagert ist. Nach HASSENPFUG (1985) lag der Kniepsand auch im 17. Jahrhundert noch als Außensand weit westlich der Insel und ist erst im Laufe der letzten Jahrhunderte von Westen an die Küstenlinie Amrums herangewandert. ZAUSIG (1939) stellte anhand alter Seekarten fest, dass der Kniepsand bei seiner Wanderung nach Osten einen alten Verlauf des Hörnum-Tiefs zuschüttete, das damals zwischen dem Kniepsand und Amrum verlief. Der mittlere und südliche Teil der Sandbank lagerte sich zuerst an Amrums Westküste an.

So wird der Kniepsand in den Karten in den Abbildungen 2 C bis 2 F als eine zu Amrum zählende Sandbank eingezeichnet, die auf der Höhe des Untersuchungsgebietes am Leuchtturm von Amrum bereits eine direkte Verbindung zum Inselkern aufweist. Deutlich erkennbar ist in den zwei Seekarten (Abb. 2 C und D) und in der topographischen Karte von 1878 (Abb. 2 E) der Kniephafen, der zu dieser Zeit einen beschiffbaren (REMDE 1972; QUEDENS 2008) Hafen darstellte.

Das Andocken des Kniepsandes an die Insel erfolgte morphologisch als Sandbank. Nach der Anlagerung des einstigen Außensandes erfuhr der Kniepsand allmählich eine Modifikation, die primär durch eine nach Norden und Süden gerichtete Strömungskomponente des Küstenlängstransportes gesteuert wurde. Die Art und Weise seiner Weiterentwicklung charakterisieren den heutigen Kniepsand dementsprechend als Ablagerung einer Meeresströmung und somit eindeutig als Nehrungsbildung. Bereits JOHANNSEN (1867) zitiert nach MÜLLER & FISCHER (1937: 93) wies auf diese morphologische Veränderung hin und schrieb über den Kniepsand „[...] diese zwei Stunden lange Sandbank verlängert sich nach Süden wie nach Norden und wird nur bei Springflut und außerordentlichen Sturmfluten überschwemmt“.

Die Wurzel der Nehrung bzw. ihr Aufhängerpunkt an der Küste ist nach VOIGT (1969) westlich des Amrumer Leuchtturms zu suchen. Von seiner ursprünglichen Ansatzzone aus schob sich der Kniepsand dementsprechend nach Nordwesten vor. Durch eine nach Norden gerichtete Meeresströmung wuchs der Nehrungsstreifen allmählich von seinem Wurzelgebiet am Wriakhörn südlich des Leuchtturms zunächst nach Nordwesten und später nach Norden (VOIGT 1969). Der nördliche Teil bewegte sich langsamer, sodass eine Bucht mit Nehrungscharakter und Lagune entstand in der sich bis 1890 (QUEDENS 2008) der Kniephafen befand. Dieser stellte in historischer Vergangenheit einen wichtigen Ankerplatz für Fischer- und Segelboote dar (JESSEN 1932) und reichte in seiner äußersten Ausdehnung bis westlich der Satteldüne. Reste der ehemaligen Bucht waren noch 1913 zu erkennen (KRAUSE 1913) und sind in der Karte von 1927 (Abb. 2 E) nördlich des Quermarkenfeuers abgebildet. Der Kniephafen verlandete allmählich von Süden her und wurde im Norden von dem langsam nachrückenden Teil des Kniepsandes geschlossen (REMDE 1972).

Gemäß JOHANNSEN (1861) war der Kniepsandhafen bereits um das Jahr 1860 nicht mehr für Lastensegler schiffbar. Nach KRAUSE (1913) lagerte sich der Kniepsand zuerst in seiner Mitte an den Inselkörper an, sodass infolge seiner West-Ost-Wanderung die Strömung in dem parallel zur Küstenlinie verlaufenden Priel „Troode diep“ (VOIGT 1969) unterbrochen und Stillwasserbereiche für die Akkumulation feinkörniger Sedimente geschaffen wurden. ZAUSIG (1939) vermutete ferner, dass der nördliche Teil von einem anstehenden festeren Untergrund in seiner Wanderungsbewegung verlangsamt wurde. Dieser festere Untergrund ist in den historischen Karten als „Sylter Riff“ eingetragen. Heute hat der Kniepsand sich vollständig an die Westküste von Amrum angeschlossen und den ehemaligen Kniephafen mitsamt der Bucht zugeschüttet. Demzufolge setzt sich der heutige Kniepsand morphologisch aus zwei ihrer Entstehung nach verschiedenen Tei-

len zusammen, nämlich der eigentlichen Nehrung und dem Kniephafen, den man als das zum Nehrungsstreifen gehörende Hafl bezeichnen kann (VOIGT 1969).

### 3 Methoden

Im Untersuchungsgebiet Amrumer Leuchtturm wurde eine Methodenkombination aus Georadarmessungen, Bohrungen, sedimentologischen Laboranalysen und AAR-Datierungen angewandt.

#### 3.1 Georadar (GPR)

Das Georadar, engl. Ground-penetrating radar (GPR), ist ein hochauflösendes elektromagnetisches Impulsreflexionsverfahren, das eine hochauflösende und zerstörungsfreie Prospektion des oberflächennahen Untergrundes bietet (vgl. u. a. ANNAN 2001, 2009; BLINDOW et al. 2005). Elektromagnetische Wellen breiten sich im Untergrund aus und werden bei Änderungen elektrischer Materialeigenschaften reflektiert, gebrochen, gestreut und zum Teil absorbiert. Das reflektierte elektrische Feld der elektromagnetischen Welle wird aufgezeichnet und in einem Weg-Zeit-Diagramm, dem Radargramm, dargestellt. Durch die Reflexion an Schichtgrenzen oder Störkörpern erhält man ein quasi kontinuierliches Profil des Untergrundes in Abhängigkeit von dessen dielektrischen Eigenschaften (ANNAN 2009).

Die nachfolgenden Georadarmessungen wurden mit einem Georadarsystem SIR 2000 der Firma Geophysical Survey Systems Inc. (GSSI SIR-2000) in Kombination mit einer 200 MHz Antenne durchgeführt. Insgesamt wurden im Untersuchungsgebiet am Leuchtturm von Amrum 8 GPR-Profile auf natürlichem Untergrund gemessen (Abb. 1). Für die Georadarmessungen wurde ein 2-dimensionales und 2,5-dimensionales Survey Design gewählt. Eine 2,5 dimensionale Darstellung wird durch Kreuzungs- oder Berührungspunkte einzelner zweidimensionaler GPR-Profile gewährleistet und ermöglicht so eine pseudo-dreidimensionale Betrachtungsweise der Untergrundstrukturen im Arbeitsgebiet. Die Messungen erfolgten parallel und senkrecht zur heutigen Küstenlinie.

Die Transekte der GPR-Profile wurden bevorzugt auf ebenem Gelände und in den flachen Bereichen der Dünentäler gemessen. Dennoch führen auch schon geringe topographische Höhenunterschiede entlang der Georadar-Transekte zu Verzerrungen im Radargramm (z. B. FISHER et al. 1996; JOL & BRISTOW 2003; TILLMANN & WUNDERLICH 2012) und wurden daher im unmittelbaren Anschluss an die Georadarmessungen mit Hilfe eines differentiellen GPS Systems (Ashtech ProMark 2) erfasst. Die Ausbreitungsgeschwindigkeiten der elektromagnetischen Welle im Untergrund wurden mit Hilfe von Diffraktionshyperbeln und durch die Korrelation mit Bohrungen ermittelt.

Das Processing der GPR-Rohdaten erfolgte mit der Software „Reflex-Win“ (Version 5.6) der Firma „Sandmeier Scientific Software, Karlsruhe“. So wurde ein Bearbeitungsverfahren aus *offset*-Korrektur, *dewow*-Filter, Bandpassfilter und *background removal*-Filter standardmäßig angewendet.

Eine mit zunehmender Tiefe verstärkte Signalabschwächung konnte mit Hilfe der *gain*-Funktion (*energy decay*)

kompensiert werden. Die Migration der Radargramme erfolgte unter Verwendung des Migrations-Algorithmus nach STOLT (1978). Jedem einzelnen GPR-Profil wurde dabei eine einheitliche Durchschnittsgeschwindigkeit zugrunde gelegt. Durch die Migration wurden die Achsen von Diffraktionshyperbeln entfernt und die Reflexionen der Objekte auf einen punktförmigen Reflektor reduziert, der der tatsächlichen Objektlage und -größe entspricht. Wo Diffraktionshyperbeln im starken Maße zur Interpretation beitragen, wurde auf das Entfernen der Hyperbeln im Radargramm verzichtet.

#### 3.2 Bohrungen und sedimentologische Laboranalysen

Im Untersuchungsgebiet am Leuchtturm von Amrum wurden Rammkern- und Pürkhauersondierungen durchgeführt. Die Rammkernsondierungen AL2 und AL6 sowie die Pürkhauersondierung AL1 wurden direkt in das Transekt von GPR-Profil 1 eingebracht. Die Rammkernsondierungen AL2 und AL6 wurden mit einer Kombination aus Benzinschlagkopfborner (Wackerborner), Schlitzsonden und hydraulischem Ziehgerät durchgeführt und bis zu einer Tiefe von 8 m abgeteuft. In Abhängigkeit von der Eindringtiefe kamen unterschiedliche, sich mit zunehmender Tiefe verjüngende Schlitzsonden ( $\varnothing = 80$  mm,  $\varnothing = 60$  mm,  $\varnothing = 30$  mm), zum Einsatz.

Aus den Bohrkernen wurden ca. 90 Sedimentproben für granulometrische Laboranalysen (Nass- und Trockensiebung) entnommen. Zur Berechnung korngößenstatistischer Parameter wurde das Programm Gradistat nach BLOTT & PYE (2001) verwendet. Die Auswertung nach korngößenstatistischen Parametern erfolgte mit der Methode nach FOLK & WARD (1957).

#### 3.3 Datierung

Für die Datierung der Probe aus dem Untersuchungsgebiet Amrum Leuchtturm wurde die Aminosäure-Racemisierungsmethode (AAR) herangezogen.

Die AAR-Datierung wurde im Jahre 2012 von Herrn Dr. Daniel Ziehe im Johann Heinrich von Thünen-Institut in Braunschweig durch die in ZIEHE (2009) entwickelte Methode durchgeführt.

Die AAR-Datierung beruht auf der Bestimmung von Aminosäure-D/L-Verhältnissen in den organischen Matrices biogener Carbonate (vgl. DEMARCHI et al. 2011; GOODFRIEND 1987; MURRAY-WALLACE & BELPERIO 1994; ZIEHE 2009). Die Anwendbarkeit der AAR-Methode für geologisch sehr junge Ablagerungsräume wurde von ZIEHE (2009) am Beispiel unterschiedlicher Standorte im Wattenmeer nachgewiesen.

Zur Datierung wurde eine Herzmuschel (*Cerastoderma edule*) herangezogen, die als intakte Schalenklappe aus einer Bohrtiefe von 728 cm gewonnen wurde. Aufgrund ihrer Einbettung in schillfreie, feinkörnige Wattedimente ist davon auszugehen, dass es sich um den ursprünglichen Lebensraum der Muschel handelt, bzw. diese nicht in Form von Bruchschill verlagert wurde. Um etwaige Messungenauigkeiten zu minimieren, wurde die Probe 3-fach datiert. Aus abweichenden Messergebnissen wurde anschließend ein Mittelwert gebildet. Die zeitliche Einordnung mittels der AAR-Methode ist aufgrund der geringen Datendichte an datierbarem Mate-

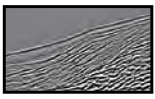
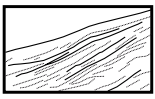

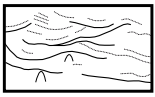
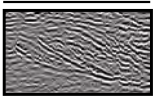
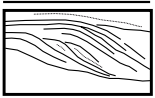
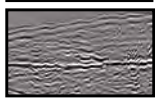
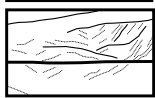

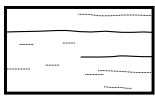

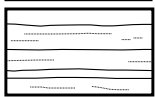
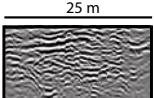
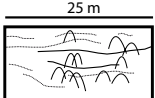
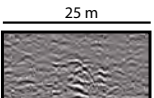
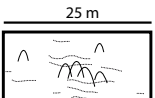
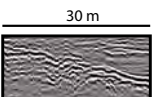
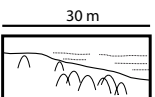
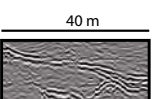
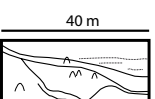
Radargramm	Interpretation	Radar-fazies	Amplitude	Form, Winkel, Kontinuität, Besonderheiten	Position, Tiefe	Sedimentologische Interpretation
<b>A. Einfallende Reflexionen</b>						
		Rf-AL-A1	stark - mittel	kontinuierlich, subparallel - parallel	vorwiegend oberhalb von Rf-AL-B1	Dünen-Fazies, äolische Kreuz- und Schrägschichtung
		Rf-AL-A2	mittel	wellen- bis rinnenförmige konkave + konvexe Reflexionen, einzelne Diffraktionshyperbeln, mäßig kontinuierlich	im Bereich von Rf-AL-B1 bis unter Rf-AL-B1	intertidale Sedimentkörper/ swash bars und ridge-and-runnel-Topographie auf dem Strand
		Rf-AL-A3	stark - mittel	sigmoidal - oblique tangential, kontinuierlich - mäßig kontinuierlich	ausschließlich unter Rf-AL-B1	Progradations-Fazies/ Kniepsand
<b>B. Horizontale Reflexionen</b>						
		Rf-AL-B1	stark	planar kontinuierlich, horizontal - subhorizontal, linear		Grundwasserspiegel
		Rf-AL-B2	mittel-schwach	horizontal - planar, parallel - subparallel, mäßig kontinuierlich - diskontinuierlich	ausschließlich unter Rf-AL-B1	Watt-Fazies
		Rf-AL-B3	mittel - schwach	horizontal, planar, teilweise wellenförmig, parallel - subparallel, kontinuierlich - mäßig kontinuierlich	im Bereich von Rf-AL-B1 bis unter Rf-AL-B1	Strand-Fazies
<b>C. Irreguläre Reflexionen</b>						
		Rf-AL-C1	stark - mittel	mäßig kontinuierlich, wellenförmig, teilweise chaotisch, viele Diffraktionshyperbeln	unter Rf-AL-B1	Geschiebenester im Geestkern
		Rf-AL-C2	schwach	diskontinuierlich, chaotisch, Diffraktionshyperbeln	vorwiegend unter Rf-AL-B1	Geestkern-Fazies
<b>D. Radar-Schichtflächen</b>						
		Rf-AL-D1	stark	nach Westen einfallende Schichtfläche mit Diffraktionshyperbeln, kontinuierlich	größtenteils unter Rf-AL-B1	Geestkernoberfläche, Grenze zwischen Watt/ Strand und Geestkern coastal onlap
		Rf-AL-D2	stark	steil einfallende Schichtflächenkonfiguration, Diffraktionshyperbeln, kontinuierlich	unter Rf-AL-B1	fossile Kliffs entstanden durch sturmflutbedingte Erosion am Geestkern, Brandungshohlkehlen

Abb. 3: Radarfazies-Definition aus dem Untersuchungsgebiet Amrum Leuchtturm.

Fig. 3: Radarfacies-definition of the study area next to the lighthouse of Amrum.

rial sowie dem Fehlen von Vergleichsaltern lediglich als grober zeitlicher Anhaltspunkt zu sehen.

## 4 Ergebnisse

### 4.1 Radarfazies Analyse

Reflexionsmuster in GPR-Daten lassen auf Sedimentstrukturen im Untergrund schließen. Die Interpretation der gezeigten GPR-Profile basiert auf der Methode der Radarstratigraphie. Die Radarstratigraphie wurde erstmals durch JOL & SMITH (1991) als eine Interpretationstechnik für Georadardaten eingeführt und basiert in ihren Grundlagen auf den Prinzipien der in den 1970er Jahren von der Erdölindustrie entwickelten Seismischen Stratigraphie nach MITCHUM et al. (1977). Diese geht von sedimentären Sequenzen (speziell siliziklastischen Sequenzen) aus, die in eine Hierarchie der Sedimentationseinheiten von einzelnen Laminae bis hin zu sedimentären Beckenfüllungen unterteilt werden (ALLEN 1982; CAMPBELL 1967; MIALL 1991) und stellt damit die Grundlage der Sequenzstratigraphie dar (PAYTON 1977). Durch die Definition einer Radarfazies werden die Reflexionsmuster des Untersuchungsgebietes klassifiziert und entsprechend einer von NEAL (2004) entwickelten Standardterminologie zusammengefasst. Auf diese Weise ist es möglich, die Radargramme in ihre stratigraphischen Einheiten zu unterteilen, und so in eine relative chronologische Reihenfolge einzuordnen. Im Untersuchungsgebiet am Leuchtturm von Amrum ließen sich insgesamt 10 unterschiedliche Radarfazies definieren. Die Benennung der Radarfazies erfolgte nach folgendem Schema. Eine erste Einteilung der Reflexionen erfolgte in Bezug auf das Untersuchungsgebiet und wird mit „AL“ für „Amrum Leuchtturm“ abgekürzt. Des Weiteren wird zwischen einfallende Reflexionen (Rf-AL-A), horizontale Reflexionen (Rf-AL-B), irreguläre Reflexionen (Rf-AL-C) und Radarschichtflächen (Rf-AL-D) unterschieden (Abb. 3).

#### Rf-AL-A1 Dünen-Fazies

##### Radarfazies

Die Radarfazies Rf-AL-A1 umfasst einfallende kontinuierliche bis moderat kontinuierliche Reflektoren von mittlerer bis starker Amplitude (vgl. auch TILLMANN & WUNDERLICH 2011a, 2011b, 2013). Die einzelnen Reflektoren sind zueinander parallel bis subparallel orientiert und besitzen ein steiles bis mittleres Einfallen ( $10^{\circ}$ – $30^{\circ}$ ) in unterschiedliche Richtungen. Die Reflexionen sind in einzelne Radarschichtpakete unterteilt.

##### Sedimentfazies

In den Bohrkernen AL1, AL2 und AL6 besteht Rf-AL-A1 ausschließlich aus hellen Fein- und Mittelsanden. Die mittlere Korngröße bewegt sich im feinen Mittelsandbereich zwischen  $2,11 \phi$  und  $1,73 \phi$ . Die Sedimente von Rf-AL-A1 weisen eine mäßig gute ( $0,60 \phi$ ) bis mäßige Sortierung ( $0,81 \phi$ ) auf und zeigen dadurch nur geringe Abweichung innerhalb des Korngrößenspektrums.

Die Korngrößenverteilung von Rf-AL-A1 ist in der Regel unimodal bis bimodal mit dem größten Anteil an der Korngröße Mittelsand. Die Schiefe ist dementsprechend vorwiegend symmetrisch ( $0,067 \phi$  bis

$-0,002 \phi$ ). Geringfügige Abweichungen zu einer positiven und negativen Schiefe sind dennoch festzustellen und abhängig davon, ob innerhalb der Sedimentprobe von Rf-AL-A1 gröbere oder feinere Mittelsande dominieren.

Die Sedimente von Rf-AL-A1 enthalten im Bohrkern keine makroskopisch erkennbaren Muscheln oder Schillanteile. Grobe Komponenten wie Kiese sind ebenfalls nicht nachzuweisen. Demgegenüber sind Wurzeln in Verbindung mit organischen Humushorizonten häufig.

##### Interpretation

Die Radarfazies Rf-AL-A1 wird als Dünen-Fazies interpretiert. Eine vergleichbare Radarfazies-Definition ist bereits aus früheren Arbeiten (z. B. BRISTOW et al. 1996, 2000a, 2000b, 2010a, 2010b; BOTHA et al. 2003; GIRARDI & DAVIS 2010; HARARI 1996; TILLMANN & WUNDERLICH 2012, 2013) bekannt. Im Untersuchungsgebiet bildet die Dünen-Fazies die jüngste stratigraphische Einheit und zeigt im Radargramm (Abb. 4) die für Küstendünen typischen äolischen Sedimentstrukturen der Schräg- und Kreuzschichtung, die durch kleinräumig wechselnde Windrichtungen bedingt sind (MCKEE 1966; TUCKER 1985; VAN OVERMEEREN 1998).

Die interne Dünenschichtung setzt sich aus einzelnen sog. „foresets“ zusammen und ist von als „bounding surfaces“ (vgl. BRISTOW 2009; BRISTOW & PUCILLO 2006) bezeichneten Schichtflächen unterbrochen. Generell resultiert die interne Struktur und Schichtung von Dünen aus Änderungen des vorherrschenden Windregimes bzw. einer dadurch bedingten Wanderungsbewegung (DAVIS 1992).

Ein Beispiel für die komplexe interne Geometrie der Dünen im Untersuchungsgebiet am Leuchtturm von Amrum zeigt die Vordüne aus GPR-Profil 5 in Abbildung 4. Das 130 m lange GPR-Transekt wurde im Weißdüngürtel parallel zur Küste mittels einer 200 MHz-Antenne gemessen und verläuft von Nordwesten nach Südosten. Der als Sedimentlieferant fungierende Strand (Kniepsand) befindet sich südwestlich der Vordünenkette. Die Düne zeigt zahlreiche einzelne foresets-Schichten, die in Sedimentpakete von bis zu 2 m Mächtigkeit zusammengefasst und durch zahlreiche bounding surfaces begrenzt werden. Unterschiedliche Einfallswinkel und Richtungen der bounding surfaces lassen kleinräumig wechselnde Windbedingungen vermuten. Diffraktionshyperbeln innerhalb des Dünenkomplexes und an den Oberflächen der bounding surface sind häufig und werden in Übereinstimmung mit Beiträgen von GIRARDI & DAVIS (2010) und VAN DAM (2012) als durch Dünensand überdeckte Vegetation (evtl. Strandhaferhorste) interpretiert.

Abbildung 4 (Bild D) zeigt einen Ausschnitt der internen Dünenschichtung im Kern einer in das GPR-Transekt eingebrachten Pürkhauersondierung. Die mittleren Korngrößen bewegen sich zwischen  $1,092 \phi$  und  $1,719 \phi$ . Demzufolge entsprechen alle Schichten Sanden der Mittelsandfraktion und zeigen somit nur geringe Variationen hinsichtlich der Korngröße.

Dennoch können schon minimale Änderungen von Korngröße und Kornzusammensetzung der Dünensande Änderungen im Wasserhaushalt der einzelnen Schichten hervorrufen und dadurch im Radargramm zu Reflexionen führen. Auch organische Boden- und Wurzelhorizon-



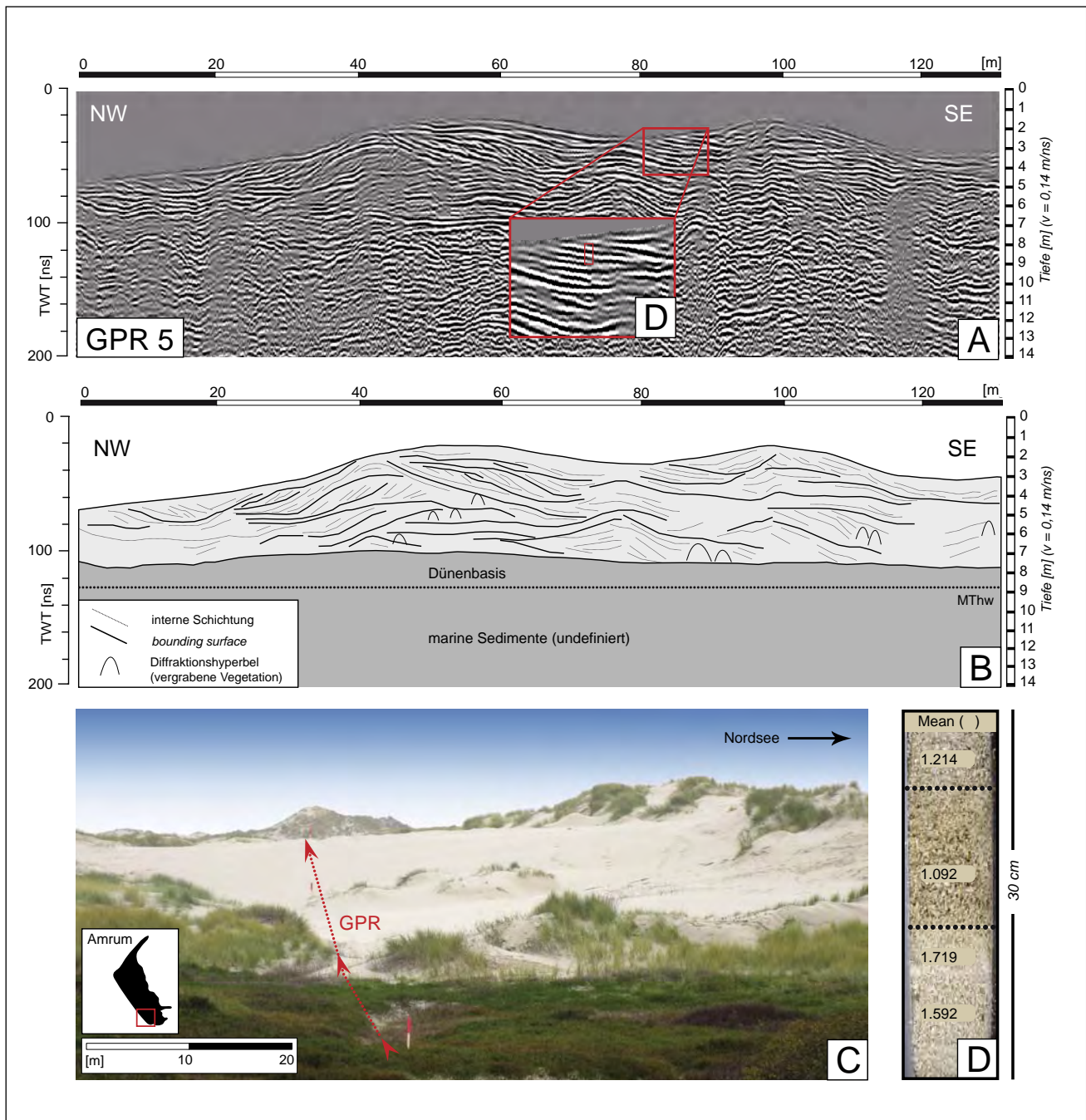


Abb. 4 A: Das Radargramm (200 MHz) zeigt einen Ausschnitt aus GPR-Profil 5. Die Lage von GPR-Profil 5 im Untersuchungsgebiet ist aus Abbildung 1 ersichtlich. Für die GPR-Daten wurde ein Basis-Processing aus (static correction, subtract-mean dewow, Bandpassfilter, background removal, gain) angewandt und die Topographie mit Hilfe von dGPS Koordinaten ausgeglichen. Für die trockenen Dünen sande wurde eine Durchschnittsgeschwindigkeit von 0,14 m/ns ermittelt. B: Interpretation und Legende. Bounding surfaces sind durch starke Linien gekennzeichnet. Die interne Schichtung ist anhand feiner Linien zu erkennen. C: Foto der Düne mit Lage von GPR-Profil 5. D: Pürkhauer-Bohrung und Korngrößenverteilung innerhalb der Dünen-Fazies.

Fig. 4 A: The radargramm (200 MHz) shows a section of GPR-profile 5. Position of GPR-profile 5 in the study area is shown in figure 1. Processing (static correction, subtract-mean dewow, bandpassfilter, background removal, gain) and topographic correction of the radargramm were already applied. A velocity of 0.14 m/ns was determined for dry dune sand. B: Interpretation and legend. Bold lines represent bounding surfaces, fine lines display internal structures. C: Picture of the dune with position of GPR-profile 5. D: Core and grain size characteristic of the dune facies.

te konnten innerhalb der Dünen-schichtung nachgewiesen werden. Sie verursachen im Radargramm ebenfalls verstärkte Reflexionen.

#### **Rf-AL-A2 intertidale Sedimentkörper (swash bars) Radarfazies**

Die Reflexionen der Radarfazies Rf-AL-A2 kommen im Radargramm im direkten Verbund mit den Reflexionen der Radarfazies Rf-AL-B3 vor und umfassen den Tiefenbereich

von 0 m NN bis -4 m NN. Dementsprechend ist Rf-AL-A2 ausschließlich unterhalb des Grundwasserspiegels nachzuweisen.

Im Radargramm bilden die Reflexionen von Rf-AL-A2 sowohl konkave, rinnenförmige Hohlformen als auch konvexe Vollformen aus. Hohlformen und Vollformen kommen in unmittelbarer Nachbarschaft zueinander vor und lassen auf einen morphologischen Zusammenhang schließen. Gelegentlich sind innerhalb der Radarfazies von Rf-AL-A2 ein-

zelle Diffraktionshyperbeln festzustellen, die allerdings kein eindeutiges Verteilungsmuster aufweisen.

### Sedimentfazies

Hinsichtlich der Sedimentfazies bestehen keine oder nur geringfügige Unterschiede von Rf-AL-A2 und Rf-AL-B3. Dies deutet auf einen morphologischen Zusammenhang oder einen einheitlichen Sedimentationsraum hin. So dominieren schillführende Fein- und Mittelsande ( $1,00\phi - 2,13\phi$ ), die eine mäßige bis mäßig gute ( $0,60\phi - 0,97\phi$ ) Sortierung aufweisen. Die Korngrößen sind vorwiegend unimodal und bimodal verteilt, besitzen eine beinahe symmetrische Schiefe und zeigen daher nur geringfügige Abweichungen.

### Interpretation

Der direkte Verbund der Reflexionen von Rf-AL-A2 und Rf-AL-B3 im Radargramm deutet auf einen einheitlichen Sedimentationsraum sowie eine ähnliche Genese der durch Rf-AL-A2 und Rf-AL-B3 abgebildeten Sedimentstrukturen hin. Geomorphologisch handelt es sich bei den Radarfazien Rf-AL-A2 und Rf-AL-B3 um einen Strand, der den Tiefenbereich von 0 - 4 m unter NN umfasst.

Im Vergleich zu horizontal lagernden und in der Regel gut geschichteten Wattsedimenten bilden die als Strand-sedimente interpretierten Reflexionen von Rf-AL-A2 und Rf-AL-B3 eine typisch unruhige Strandmorphologie aus. Primär wird diese durch die Radarfazies Rf-AL-A2 hervorgerufen, die aufgrund ihrer Reflexionsgeometrien sedimentologisch auf Voll- und Hohlformen hindeutet.

Der interne Aufbau der Vollformen ähnelt dabei der internen Schichtung intertidaler Sedimentkörper (*tidal bars*, *swash bars*) wie sie heute im Strandbereich vieler Nordseeinseln zu finden sind. Intertidale Sedimentkörper (*intertidal bars* und *swash bars*) zählen nach FITZGERALD (2005) zu den typischen Elementen der Strandmorphologie im foreshore-Bereich. Die sog. *swash bars* zeigen im Allgemeinen eine durch Wellenenergie induzierte, auf die Küste gerichtete Migrationsrichtung (HAYES 1980; JACKSON et al. 2007), die sich im Radargramm durch landwärts einfallende Reflexionen abbildet (PEDERSEN & CLEMMENSEN 2005). Stratigraphisch kommen diese überwiegend als Vollformen ausgebildeten Sedimentkörper im Verbund mit parallel zur Küstenlinie verlaufenden linearen rinnen- oder wannenförmigen Geländedepressionen vor und gehören zu den temporären Sequenzen innerhalb des Strandprofils (FITZGERALD 2005; GREENWOOD 2005).

Aus dem deutschsprachigen Raum sind diese intertidalen Vollformen mit den in SCHÄFER (2005: 228 ff.) angesprochenen „Strandriffen“ vergleichbar, die als mehrfach aufeinander folgende, strandparallele Sandbänke in der Brecherzone von der senkrecht zur Küste auflaufenden Brandung aufgeworfen werden. Die Seeseite der Strandriffe ist flach.

Auf der Landseite bilden die steil einfallenden ( $18^\circ - 30^\circ$ ) Riffstirnschichten mit landwärts orientierten Anlagerungsflächen eine parallele Schrägschichtung aus (SCHÄFER 2005).

Oft sind diese vorwiegend aus Sand aufgebauten intertidalen Sedimentkörper stratigraphisch in das Strandprofil eingebaut und aufgrund des geringen morphologischen und sedimentologischen Unterschiedes zum Strand

(JACKSON et al. 2007) nur schwer anhand von Bohrungen zu identifizieren. Die Entstehung und Anlagerung von *swash bars* an den Strand ist vor allem auf ruhige Phasen zwischen oder nach Sturmfluten zurückzuführen (JACKSON et al. 2007; TANNER 1995). Während den Sturmfluten werden die INTERTIDALEN Sedimentkörper zum Teil ausgeräumt, aufgearbeitet oder in ihrem oberen Bereich gekappt, sodass ein durch Wellen- und Brandungserosion bedingtes ebenes Strandprofil entsteht (FITZGERALD 2005; JACKSON et al. 2007).

Vollformen und Geländedepressionen wechseln sich innerhalb der Radarfazies Rf-AL-A2 ab und sind mit der in GREENWOOD (2005) beschriebenen typischen *ridge-and-runnel*-Topographie einiger Strände vergleichbar. Die konvexen Vollformen der *intertidal bars* oder *swash bars* repräsentieren dabei die Sedimentrücken (*ridges*). Rinneförmige Reflexionen entsprechen den sog. *runnels*, die von SCHÄFER (2005) auch als Strandpriele bezeichnet werden. Die *ridge-and-runnel*-Strukturen kommen in Verbindung mit den annähernd horizontalen Reflektoren von Rf-AL-B3 vor, die als nahezu sählig lagernde, leicht in Richtung Meer einfallende Strandsandschichten gedeutet werden. Diese zeigen die von SCHÄFER (2005: 228) als „Hochenergie-Parallelschichtung“ bezeichnete interne Sedimentstruktur. Die Ablagerung intertidaler Sedimentkörper erfolgt vorzugsweise in den Ruhephasen zwischen einzelnen Sturmfluten. Wiederkehrende Sturmfluten sorgen teilweise für die Erosion der intertidalen Vollformen und für die anschließende Ausbildung eines zeitweilig ebenen Strandprofils (JACKSON et al. 2007).

### Rf-AL-A3 Progradations-Fazies

#### Radarfazies

Die Radarfazies Rf-AL-A3 gehört zu den einfallenden Reflexionen und setzt sich aus sigmoidal bis schräg tangential geformten Reflexionen zusammen, die in einzelne Radarschichtpakete zusammengefasst sind. Rf-AL-A3 ist ausschließlich unterhalb des Grundwasserspiegels (Rf-AL-B1) nachzuweisen. Die Reflektoren von Rf-AL-A3 verlaufen kontinuierlich bis mäßig kontinuierlich und sind von mittlerer bis starker Amplitude. Die gebündelten Reflektoren enden in der Regel mit einem *downlap* und zeigen ein nach Westen gerichtetes Einfallen.

#### Sedimentfazies

Die Radarfazies Rf-AL-A3 wurde an keiner Stelle direkt durch Bohrungen durchteuft. Die Reflektoreigenschaften und Reflexionsgeometrien sowie der Vergleich ähnlicher Reflexionsmuster (TILLMANN & WUNDERLICH 2011a, 2011b, 2013) lassen jedoch eine sandige Sedimentfazies vermuten.

#### Interpretation

Die sigmoidale Form der Reflexionen sowie die schräg tangentielle Beziehung der Reflektoren zueinander deuten auf eine progradierende Schichtung der Sedimente von Rf-AL-A3 hin. Die in *radar packages* gebündelten Reflexionen lassen auf eine strömungsbedingte Anlagerung einzelner sandiger Sedimentpaketen schließen. Das Vorkommen von Rf-AL-A3 unterhalb des Grundwasserspiegels lässt zudem rein marine Sedimente vermuten.

## **Rf-AL-B1 Grundwasserspiegel**

### **Radarfazies und Interpretation**

Rf-AL-B1 markiert einen horizontalen, durchgehenden Hauptreflektor, der der Position des Grundwasserspiegels entspricht und folglich als solcher interpretiert wird. In den gezeigten Radargrammen befindet sich der Grundwasserspiegel Rf-AL-B1 im Liegenden oder innerhalb der Dünenfazies (Rf-AL-A1) und teilt die Radargramme hinsichtlich ihrer elektromagnetischen Eigenschaften in einen wassergesättigten unteren und einen wasserungesättigten oberen Teil. Bedingt durch die hohe Leitfähigkeit und den starken dielektrischen Kontrast von wassergesättigten und wasserungesättigten Sedimenten zeigt der Reflektor des Grundwasserspiegels eine starke Amplitude. Im Bereich des heutigen Kniepsandes geht der Grundwasserspiegel mit Annäherung an die MThw-Linie im Westen in salzhaltiges Brack- und Meerwasser über. Der Kontaktbereich zwischen Meer- und Grundwasser zeigt sich in den Georadardaten durch eine allgemeine Abschwächung der Reflexionen im Untergrund sowie eine nach Westen verringerte Erkundungstiefe.

## **Rf-AL-B2 Watt-Fazies**

### **Radarfazies**

Die Radarfazies Rf-AL-B2 umfasst horizontal planare Reflektoren, die zueinander annähernd parallel orientiert sind. Die einzelnen Reflektoren verlaufen kontinuierlich bis mäßig kontinuierlich und zeigen eine mittlere bis schwache Amplitude. Rf-SS-B2a ist ausschließlich unterhalb des Grundwasserspiegels im Tiefenbereich zwischen -3,5 m NN bis -7 m NN zu finden.

### **Sedimentfazies**

Sedimente der Radarfazies Rf-AL-B2 wurden durch die Bohrung AL6 abgeteuft und durch diese komplett durchdrungen. Rf-AL-B2 besteht aus einer sehr feinschichtigen Wechsellagerung von Feinsand und Silt mit eingeschalteten Tonlagen. Damit umfasst die Radarfazies Rf-AL-B2 die feinkörnigsten Sedimente des Untersuchungsgebietes. Das Korngrößenpektrum der gemessenen Sedimentproben variiert zwischen feinem Mittelsand (1,89  $\phi$ ) und feinem Silt (7,07  $\phi$ ). Der Anteil an reinem Ton beträgt zwischen 4,2 %–40,7 %.

Die Sortierung der feinkörnigen Sedimente aus der Feinsand- und Silt-Fraktion ist mäßig gut (0,59  $\phi$ ) bis sehr schlecht (3,17  $\phi$ ). Eingeschaltete Mittelsand-Lagen sind mit Werten von 0,60  $\phi$ –0,62  $\phi$  insgesamt besser sortiert. Die Korngrößenverteilung ist zumeist bimodal und zeigt neben dem Maximum im Feinsandbereich auch einen deutlichen Anteil an Ton. Die Sedimentfarbe ist grau bis schwarz. Innerhalb der Sedimentfazies von Rf-AL-B2 lassen sich zahlreiche gut erhaltene ca. 2–4 mm große Gehäuse der Wattschnecke (*Hydrobia ulvae*) sowie vereinzelte Mangan-Konkretionen nachweisen.

### **Interpretation**

Im Radarogramm ist Rf-AL-B2 durch horizontale Reflektoren gekennzeichnet, die auf eine ungestörte Sedimentation in horizontalen Schichten hindeuten. Im Bohrkern wird diese durch eine feinschichtige Lamination von Feinsand, Silt und Ton bestätigt.

Die dunkle Sedimentfarbe, der Gehalt an organischer Substanz und Mangan lassen auf eine Sedimentation unter

reduzierenden Bedingungen schließen. Die überwiegend feinen Korngrößen von Feinsand und Silt sowie der insgesamt hohe Tongehalt zwischen 4,2 %–40,7 % sprechen für ein geringes energetisches Strömungsregime zur Zeit der Sedimentation. Gemäß der gebräuchlichen Klassifizierung nach SINDOWSKI (1973, 1979) werden die Sedimente von Rf-AL-B2 als feinsandiges Mischwatt bis Schlickwatt interpretiert. Zahlreiche Gehäuse der Wattschnecke (*Hydrobia ulvae*) bestätigen diesen Lebens- und Sedimentationsraum.

## **Rf-AL-B3 Strand-Fazies**

### **Radarfazies**

Die Radarfazies Rf-AL-B3 ist im Radarogramm in direktem Verbund mit Reflexionen der Radarfazies Rf-AL-A2 zu finden. Im Gegensatz Rf-AL-A2 verlaufen die einzelnen Reflektoren von Rf-AL-B3 horizontal, verhalten sich zueinander parallel bis subparallel und sind von mittlerer bis schwacher Amplitude. Wie schon die Radarfazies Rf-AL-A2 umfasst auch Rf-AL-B3 den Tiefenbereich zwischen 0 m und -4 m NN und kommt damit ausschließlich unterhalb des Grundwasserspiegels vor.

### **Sedimentfazies**

Sedimente der Radarfazies Rf-AL-B3 umfassen schillführende Sande aus der Kornfraktion Feinsand und Mittelsand. Die durchschnittliche Korngröße bewegt sich zwischen 1,00  $\phi$  und 2,13  $\phi$ . Mit Werten zwischen 0,60  $\phi$  und 0,97  $\phi$  sind die Sande von Rf-AL-B3 von einer mäßigen bis mäßig guten Sortierung. Die Kornverteilung ist unimodal und bimodal und zeigt nur geringe Abweichungen von der beinahe symmetrischen Schiefe. Eingeschaltete Schillhorizonte innerhalb der Strand-Fazies sind dagegen von schlechter Sortierung (1,35  $\phi$ ) und weisen eine polymodale Kornverteilung auf. In den Bohrungen ließen sich Oxidations- und Reduktionszonen innerhalb der Sedimentfazies von Rf-AL-B3 belegen. Die Oxidationszonen sind durch Rostflecken sowie die eisenhaltige Umkrustung von Sandkörnern und -konkretionen gekennzeichnet.

### **Interpretation**

Zusammen mit der Radarfazies Rf-AL-A2 umfasst die Radarfazies Rf-AL-B3 den Sedimentationsraum Strand. Rf-AL-A2 bildet im Radarogramm die in das Strandprofil eingeschalteten intertidalen Sedimentkörper (*swash bars*) ab. Diese sind für die im Untersuchungsgebiet typische unruhige Strandmorphologie bestehend aus einer Kombination aus Vollformen und Geländedepressionen (*ridge-and-runnel*-Strukturen) verantwortlich. Dagegen besteht Rf-AL-B3 aus vorwiegend parallelen, horizontalen Reflektoren, die als annähernd söhlig lagernde, leicht in Richtung Meer einfallende Strandsandschichten gedeutet werden.

Diese lassen sich mit der von SCHÄFER (2005: 228) als „Hochenergie-Parallelschichtung“ bezeichneten, internen Sedimentstruktur vieler Strände vergleichen.

## **Rf-AL-C1 Geschiebenester im Geestkern**

### **Radarfazies und Interpretation**

Kennzeichnend für die Radarfazies von Rf-AL-C1 sind zahlreiche, sich zum Teil überlagernde Diffraktionshyperbeln, die in den Georadardaten ein chaotisches, wellenförmiges Reflexionsmuster hinterlassen. Rf-AL-C1 ist ausschließlich unterhalb des Grundwasserspiegels sowie unterhalb

der Radarschichtfläche von Rf-AL-D1 nachzuweisen und kommt häufig in unmittelbarer Nachbarschaft zu den Reflexionen von Rf-AL-C2 vor.

Das Vorkommen von Rf-AL-C1 unterhalb der Radarschichtfläche von Rf-AL-D1 deutet auf Sedimentstrukturen innerhalb des pleistozänen Geestkerns hin. Chaotische, wellenförmige Reflexionsmuster in Kombination mit zahlreichen Diffraktionshyperbeln wurden auch in JAKOBSEN & OVERGAARD (2002) und VAN OVERMEEREN (1998) mit glazialen Sedimenten in Verbindung gebracht.

Auf Amrum bestehen die Sedimente der Saale-Eiszeit vorwiegend aus fluvioglazialen Schmelzwassersanden (vgl. MEIER 1987; VOIGT 1969) und stark sandigem Geschiebelehm (JESSEN 1932). Kalkhaltiger, toniger Geschiebemergel, wie er am Roten Kliff im Westen Sylts oder am Goting Kliff auf Föhr ansteht, ist auf Amrum nicht nachzuweisen (MÜLLER & FISCHER 1937).

Nach KRAUSE (1913) sind die Ablagerungen der Geest auf Amrum ursprünglich durch „Ausschlammung“ aus einer aufbereiteten Grundmoräne hervorgegangen und bestehen heute aus partiell geschichteten, geröllreichen Sanden. Der Blockreichtum Amrums im Vergleich zu den Geestkernen der Nachbarinseln spricht ebenfalls für das Hervorgehen aus einer Grundmoräne (KRAUSE 1913).

Der Geestkern der Insel Amrum zeichnet sich demnach besonders durch einen überdurchschnittlichen Geschiebereichtum aus. JESSEN (1932) weist seinerseits auf das Vorkommen sog. „Geschiebenester“ im Geestkern Amrums hin.

Diese bilden eine Konzentration von Geschieben die sich im Radargramm deutlich in Form von Diffraktionshyperbeln abbilden. Unter Rf-AL-C1 werden diese Geschiebeansammlungen innerhalb des Inselgeestkerns zusammengefasst.

### Rf-AL-C2 Geestkern-Fazies

#### Radarfazies

Die Radarfazies Rf-AL-C2 zeigt schwache, diskontinuierliche Reflexionen, die insgesamt ein chaotisches Reflexionsmuster bilden. Diffraktionshyperbeln innerhalb der Radarfazies von Rf-AL-C2 sind häufig und in der Regel chaotisch innerhalb der Radarfazies verteilt. Eine untere Grenze von Rf-AL-C2 ist in den Radargrammen aufgrund verstärkter Dämpfung mit zunehmender Untergroundtiefe nicht auszumachen. Auch Rf-AL-C2 ist ausschließlich unterhalb des Grundwasserspiegels (Rf-AL-B1) und unterhalb der Radarschichtfläche (Rf-AL-D1) nachzuweisen.

#### Sedimentfazies

Das Amrumer Pleistozän, sofern es anhand der freiliegenden Geestoberfläche, der aufgeschlossenen Kliffabbrüche und durch die durchgeführten Bohrungen bestimmt werden konnte, besteht aus Sand- und Kiesschichten mit eingelagerten Geschieben.

Die Geestkern-Fazies Rf-AL-C2 wurde durch die Bohrungen AL1, AL2 und AL6 in unterschiedlichen Tiefen abgeteuft und setzt sich aus stark glimmerhaltigen (Muskovit) Fein- und Mittelsanden zusammen. Die durchschnittli-

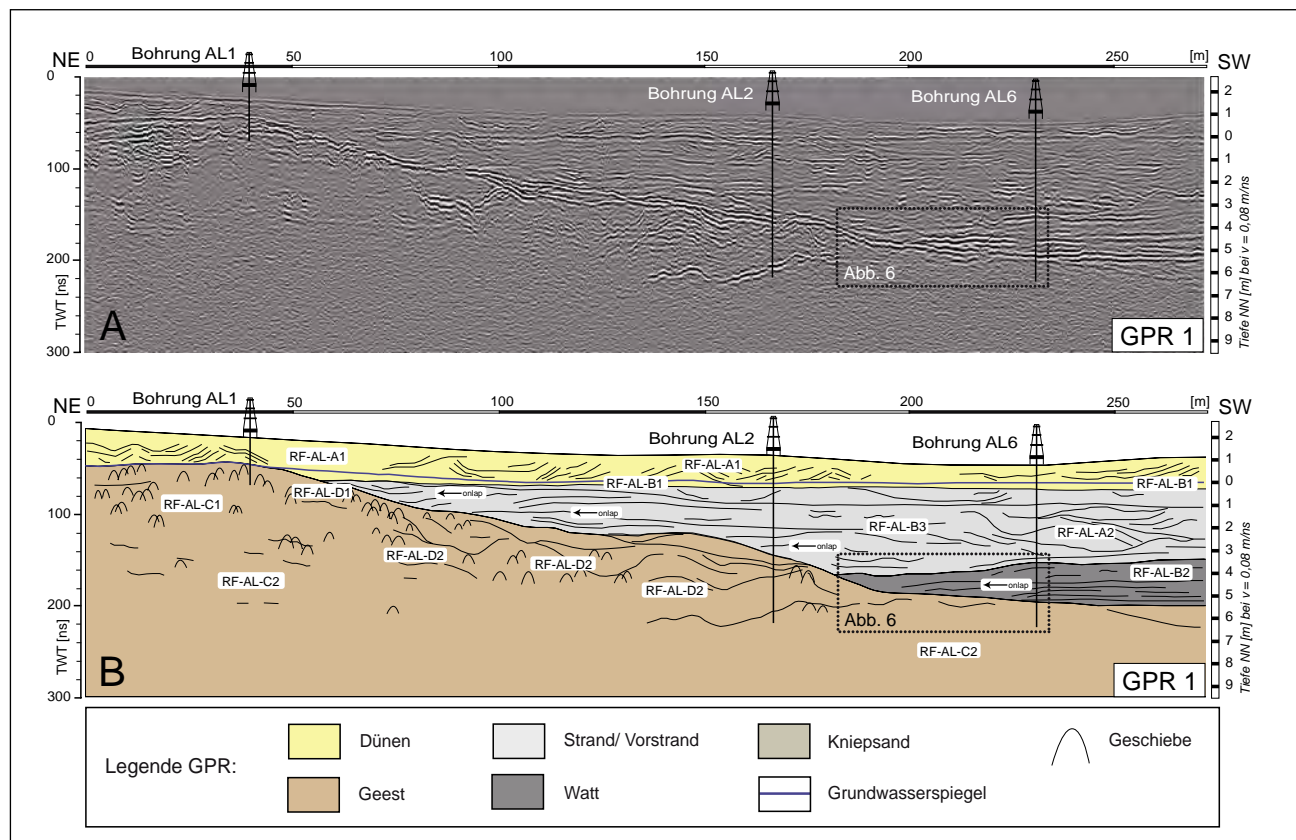


Abb. 5: GPR-Profil 1 und Position der Bohrungen AL1, AL2 und AL6. A: Für die GPR-Daten wurde ein Basis-Processing aus (static correction, subtract-mean dewow, Bandpassfilter, background removal, gain und Migration) angewandt und die Topographie mit Hilfe von dGPS Koordinaten ausgeglichen. B: Radarfazies und Interpretation von GPR-Profil 1. Die Legende gilt auch für alle weiteren GPR-Profile dieser Studie.

Fig. 5: GPR-profile 1 and position of core AL1, AL2 and AL6. A: Processing (static correction, subtract-mean dewow, bandpassfilter, background removal, gain and migration) and topographic correction of the radargramm were already applied. B: Radarfacies and interpretation of GPR-profile 1. Legend is valid for all GPR-profiles shown in this study.

che Korngröße der Geest-Sande beträgt 1,34  $\phi$  und 2,77  $\phi$ . Die Sortierung ist mit Werten zwischen 0,95  $\phi$  und 0,52  $\phi$  mäßig bis mäßig gut und ist an der Oberfläche der Geest zunehmend schlechter. Es dominieren unimodale und bimodale Korngrößenverteilungen. Vereinzelt ließen sich im Bohrkern moderat gerundete Kiese nachweisen. Muscheln oder Schillanteile konnten dagegen nicht bestätigt werden.

### Interpretation

Die Radarfazies Rf-AL-C2 entspricht der internen Struktur und Schichtung im Geestkern, die sich bedingt durch geringe Materialunterschiede in Form von Reflexionen im Radargramm abzeichnet. Ursache der internen Struktur im Geestkern können sowohl primäre fluvioglaziale Sedimentationsprozesse als auch die periglaziale Umformung des oberflächennah anstehenden Geestkerns sein.

Im Gegensatz zum zentralen Sylter Geestkern bei Kampen ist der Geestkern der Insel Amrum weniger tonhaltig und setzt sich vorwiegend aus fluvioglazialen Schmelzwassersanden und stark sandigem Geschiebelehm zusammen (vgl. JESSEN 1932; MEIER 1987; MÜLLER & FISCHER 1937; VOIGT 1969). Durch die geringen Tonanteile wurden innerhalb der Geestkern-Fazies von Amrum beachtliche Erkundungstiefen von bis zu 350 ns TWT bzw. 13,5 m Tiefe erreicht (TILLMANN & WUNDERLICH 2012).

### Rf-AL-D1 Geestkernoberfläche

#### Radarfazies

Die nach Westen einfallende Radarschichtfläche Rf-AL-D1 zeichnet sich durch einen kontinuierlichen Hauptreflektor mit starker Amplitude aus, der ausschließlich unterhalb des Grundwasserspiegels verläuft. An der Basis des Hauptreflektors sind zahlreiche Diffraktionshyperbeln konzentriert. Unterhalb der Radarschichtfläche kommen Reflexionen von Rf-AL-C1, Rf-AL-C2 und Rf-AL-D2 vor.

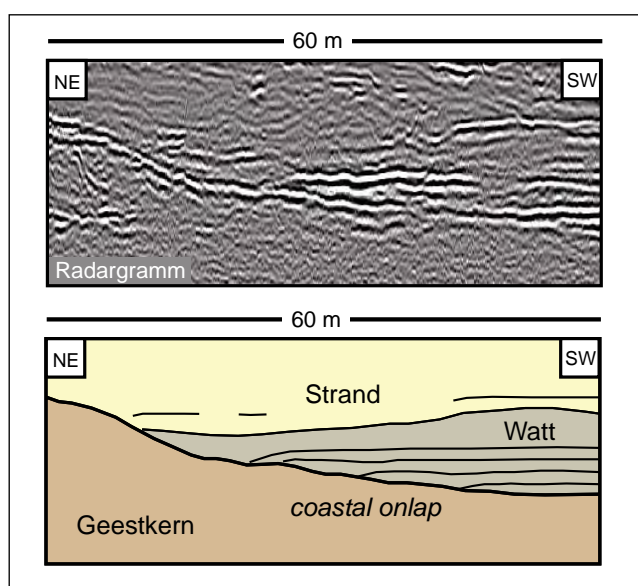


Abb. 6: Der vergrößerte Ausschnitt aus GPR-Profil 1 zeigt die coastal-onlap-Konfiguration der Reflexionen im Detail. Die Position des Ausschnitts ist im Radargramm in Abbildung 5 dargestellt.

Fig. 6: The close up of GPR-profile 1 shows the coastal-onlap configuration in detail. For position of the section see figure 5.

### Sedimentfazies

Die Radarschichtfläche von Rf-AL-D1 wurde durch die Bohrungen AL1, AL2 und AL6 durchteuft und besteht im Bohrkern aus Mittelsanden mit einer durchschnittlichen Korngröße von 1,34  $\phi$ –1,81  $\phi$ , die in Kombination mit größeren, moderat gerundeten Kiesen vorkommen. In Bohrung AL6 ließen sich im Bereich der Radarschichtfläche Wurzeln und Vegetationsreste nachweisen, die auf einen terrestrischen Sedimentationsraum bzw. eine ehemalige Landoberfläche hindeuten. Die Sedimente dieser Geestkernoberfläche sind mit Werten zwischen 1,45  $\phi$ –1,53  $\phi$  in der Regel etwas schlechter sortiert als die Sedimente, die mittels Bohrungen aus dem Inneren des Geestkerns abgeteuft wurden.

### Interpretation

Die Radarschichtfläche von Rf-SS-D3 markiert die Oberfläche des nach Westen abtauchenden Inselgeestkerns und stellt zugleich die Grenze zwischen dem Holozän im Hangenden und dem Pleistozän im Liegenden dar. Die im Radargramm an der Basis der Radarschichtfläche konzentrierten Diffraktionshyperbeln entsprechen Geschieben, die sich an der Geestkernoberfläche deflationsbedingt angereichert und dort eine Steinsohle ausgebildet haben.

### Rf-AL-D2 fossile Kliffs

#### Radarfazies

Rf-AL-D2 zeigt im Radargramm eine steil einfallende Radarschichtflächenkonfiguration mit starker Amplitude und kontinuierlichem Verlauf. Diffraktionshyperbeln lassen sich sowohl an der Basis sowie im Hangenden der Radarschichtfläche nachweisen.

### Sedimentfazies

Die Radarschichtfläche Rf-AL-D2 wurde durch die Bohrung AL2 durchdrungen. Charakteristisch für Rf-AL-D2 ist eine starke Anreicherung von moderat bis gut gerundeten Kiesen. Hinsichtlich der Korngröße besteht die Radarschichtfläche aus Grobsand (0,73  $\phi$ ), der eine schlechte Sortierung (1,26  $\phi$ ) aufweist.

### Interpretation

Unter Rf-AL-D2 werden Strukturen zusammengefasst, die an der Oberfläche des Geestkerns entstanden sind und sich in diesem an einigen Stellen lateral eingeschnitten haben. Die konkaven Erosionsstrukturen werden als fossile Kliffabbrüche im Bereich des Geestkerns interpretiert. Diese fossilen Kliffs entstanden infolge sturmflutbedingter Erosion am nach Westen abtauchenden Inselgeestkern. Brandungserosion führte am Geestkern zur Ausbildung von Brandungshohlkehlen und damit zur Kliffbildung.

Diffraktionshyperbeln zeugen sowohl an der Basis sowie im Hangenden der Radarschichtfläche von Rf-AL-D2 von Geschieben. Die an der Basis konzentrierten Hyperbeln entsprechen Geschiebeansammlungen am Fuße des Kliffs.

Schwerkraftbedingte Denudation von lockeren Geschieben und Kliffmaterial an den durch Sturmfluterosion versteilten Kliffhängen ist wahrscheinlich. Daher sind Geschiebe in Form von Diffraktionshyperbeln auch in der Hangrutschmasse enthalten. Hinsichtlich der Radarfazies und Sedimentfazies besteht kein bedeutender Unterschied

zwischen dem anstehenden Geestkern und dem in der Hangrutschmasse enthaltenen Abbruchmaterial.

## 4.2 Georadar-Profile

Eine Besonderheit der GPR-Profile im Untersuchungsgebiet sind materialbedingte Geschwindigkeitsänderungen sowohl in vertikaler als auch in lateraler Richtung. Die Ausbreitungsgeschwindigkeit innerhalb trockener Dünen- und Strandsanden unterhalb des Grundwasserspiegels verringert sich die Ausbreitungsgeschwindigkeit auf 0,09–0,07 m/ns. Innerhalb der saaleiszeitlichen Geschiebeablagerungen wurden lediglich Ausbreitungsgeschwindigkeiten von 0,05–0,08 m/ns erreicht. Auch die Wattsedimente zeichnen sich durch geringe Laufzeiten von 0,06–0,04 m/ns aus. Für die Berechnung eines einheitlichen Maßstabes wurde eine durchschnittliche Ausbreitungsgeschwindigkeit von 0,08 m/ns zugrunde gelegt. In den GPR-Profilen wurde eine maximale Erkundungstiefe von 300–400 ns TWT erreicht. Dies entspricht im Gelände einer tatsächlichen Tiefe von 8–9 m unter NN.

### GPR-Profil 1

Der obere Profildbereich wird ausschließlich von Reflexionen der Radarfazies Rf-AL-A1 eingenommen (Abb. 5). Innerhalb der Radarfazies Rf-AL-A1 verläuft der annähernd horizontale Hauptreflektor von Rf-AL-B1. Im südwestlichen Profildbereich von GPR-Profil 1, ab einer Entfernung von 60 m vom Startpunkt, finden sich im Liegenden von Rf-AL-A1 Reflexionen der Radarfazies Rf-AL-B3 und Rf-AL-A2. Während Rf-AL-B3 vorwiegend horizontale, nahezu parallele Reflektoren umfasst, beinhaltet Rf-AL-A2 zum Teil auch wellenförmige bis leicht geneigte Reflexionen. Die benachbarten Reflexionen von Rf-AL-B3 und Rf-AL-A2 kommen in einem Tiefenbereich von bis zu -3,5 m NN vor.

Die horizontalen, annähernd parallelen Reflektoren von Rf-A-B2 und Rf-A-B3 enden im Nordosten von Profil 1 mit einer deutlich ausgeprägten *onlap*-Konfiguration auf der Radarschichtfläche Rf-AL-D1. Besonders deutlich ist dies in Abbildung 6 ersichtlich, die einen vergrößerten Ausschnitt aus GPR-Profil 1 (Abb. 5) darstellt.

Der *onlap* von Rf-A-B3 vollzieht sich in einem Tiefenbereich zwischen -0,15 m und -4 m NN (60 ns–160 ns TWT). Im Liegenden von Rf-A-B3 in einer Tiefe von -4 m bis -7 m NN (160 ns–220 ns TWT) enden die Reflektoren von Rf-A-B2 ebenfalls durch eine *onlap*-Konfiguration auf der nach Westen abtauchenden Radarschichtfläche von Rf-AL-D1.

Besondere Strukturen stellen die unter Rf-AL-D2 zusammengefassten Reflexionen dar. Diese steil einfallenden Radarschichtflächen treten im Radargramm durch ihre starke Amplitude hervor und schneiden sich in die Reflexionen von Rf-AL-C2 ein. Diffraktionshyperbeln sind sowohl an der Basis als auch im Hangenden von Rf-AL-D2 konzentriert.

### GPR-Profil 3

Das GPR-Profil 3 (Abb. 7) stellt das südwestliche Profildende der GPR-Profile dar und umfasst den Profildbereich zwischen 460 m und 810 m Länge, der über den Kamm der Vordüne bis auf den Kniepsand reicht (vgl. Abb. 1).

Auch in GPR-Profil 3 wird der oberste Abschnitt von Reflexionen der Radarfazies Rf-AL-A1 eingenommen. Im südwestlichen Profildbereich ist eine verstärkte Signaldämpfung und die damit einhergehende Verringerung der Erkundungstiefe zu beobachten. Im Südosten der Düne werden im Profildbereich zwischen 725 m–745 m Länge Erkundungstiefen von -4 m NN erreicht. Ab dem Profilmeter 745 kommt es zu einer abrupten Signaldämpfung im Untergrund, die bis zum Ende des Profils deutlich zu verfolgen ist. Grund dafür ist das von Südwesten, aus dem Strand- und Überflutungsbereich des heutigen Kniepsandes, eindringende salzhaltige Meerwasser.

Der nordöstliche Profildbereich setzt sich in einer Tiefe zwischen 225 ns–165 ns TWT (0 m bis -2 m NN) aus Reflexionen der Radarfazies Rf-AL-A2 und Rf-AL-B3 zusammen. Ab einer Tiefe von 225 ns TWT dominieren Reflexionen der Radarfazies Rf-AL-A3. Diese sigmoidal bis schräg tangentialen Reflexionen sind in einzelne, wohl definierte Radarschichtpakete gebündelt, die ein Einfallen in nordöstliche Richtung aufweisen. Zwischen 540 m–610 m Profillänge enden die Reflexionsbündel von Rf-AL-A3 in einer Tiefe von 320–350 ns TWT (ca. -5 bis -6 m NN) durch eine *downlap*-Konfiguration (Abb. 8).

## 4.3 Sedimentologische Daten

### Bohrungen AL1, AL2 und AL6

In das Georadar-Transekt 1 wurden an drei Stellen die Bohrungen AL1, AL2 und AL6 eingebracht. AL1 befindet sich in 40 m Entfernung vom nordöstlich gelegenen Startpunkt des Profils und erreicht eine Bohrtiefe von 2 m unter der Geländeoberfläche (0 m NN). AL2 und AL6 wurden in 168 m und 231 m Entfernung vom Startpunkt eingebracht und wurden bis zu einer Tiefe von 8 m unter der Geländeoberfläche abgeteuft. Bohrung AL2 erreicht eine Geländetiefe von -6,25 m NN, während die maximale Geländetiefe in Bohrung AL6 -6,40 m NN entspricht. Die Ergebnisse und Korngrößenverteilung der Bohrungen AL2 und AL6 sind in den Abbildungen 9 und 10 dargestellt.

Der obere Bereich der Bohrungen AL1, AL2 und AL6 wird von holozänen Dünenansanden der Fein- bis Mittelsandfraktion eingenommen. Die äolischen Fein- und Mittelsande (2,11  $\phi$ –1,73  $\phi$ ) sind mäßig (0,81  $\phi$ ) bis mäßig gut (0,60  $\phi$ ) sortiert und zeigen insgesamt nur geringe Abweichung innerhalb des Korngrößenspektrums. Makroskopisch erkennbare Muscheln oder Schillanteile sind nicht vorhanden. Die Kornverteilung ist unimodal bis bimodal mit dem größten Anteil an der Korngröße Mittelsand. Die Schiefe variiert zwischen -0,172  $\phi$  und 0,148  $\phi$ .

Wurzeln des Strandhafers (*Ammophila arenaria*) sind in den Bohrungen AL1, AL2 und AL6 häufig und durchziehen den oberen Meter der Bohrungen AL2 und AL6. In Bohrung AL2 wurde überdies in einer Tiefe von 0,75 m ein ca. 20 cm mächtiger organisch angereicherter Boden- bzw. Wurzelhorizont nachgewiesen (Abb. 9 Bild 1), der sich bereits aufgrund des Farbunterschiedes innerhalb der hellen Dünenansanden abzeichnet.

In den dargestellten Bohrungen AL2 und AL6 gehen die äolischen Dünenansande im Hangenden ab einer Bohrtiefe von 1,30–1,40 m in die marinen Strandsande des Liegenden über (Abb. 9 und 10). Die Dünenbasis befindet sich damit

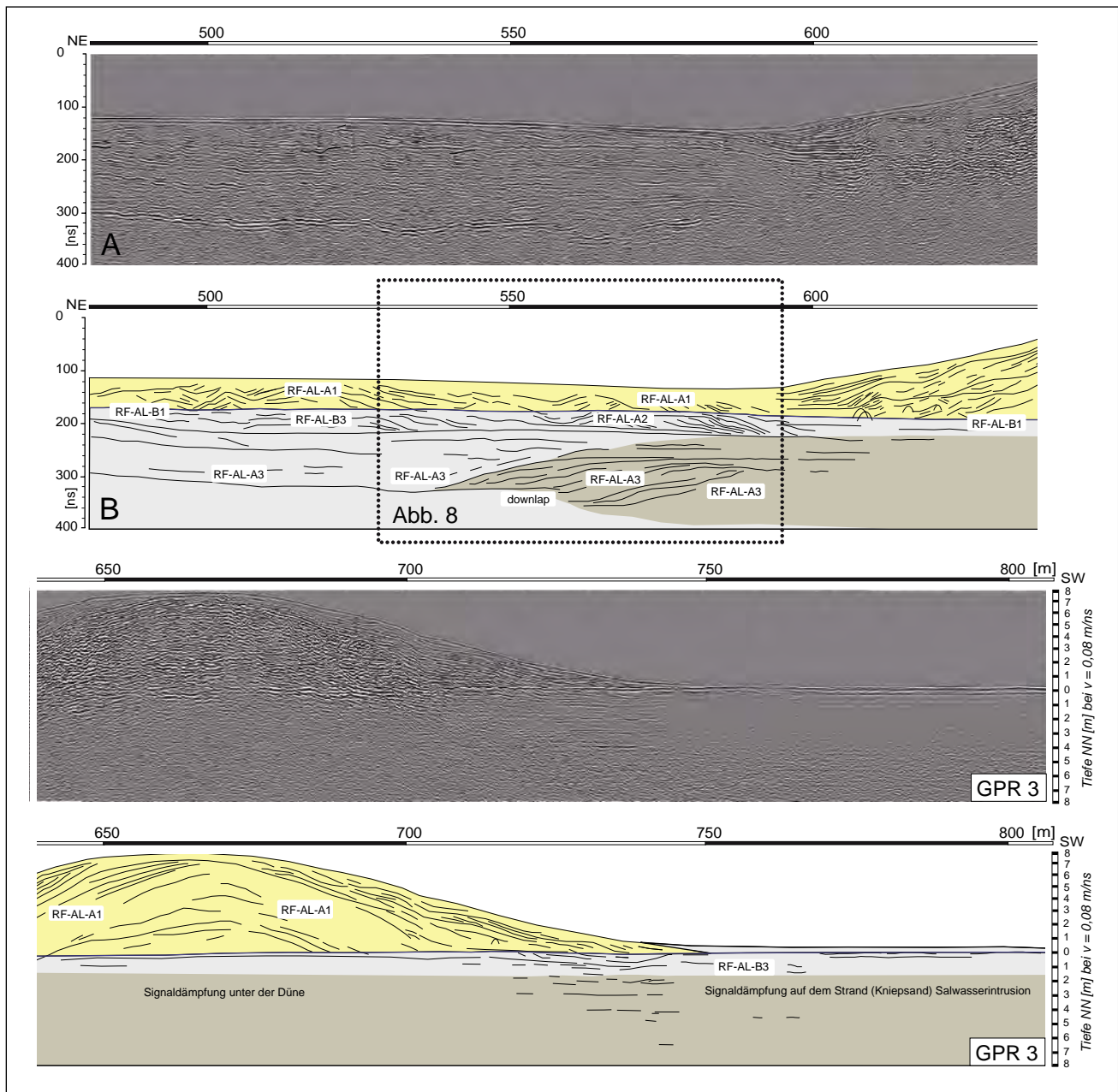


Abb. 7: GPR-Profil 3. A: Für die GPR-Daten wurde ein Basis-Processing aus (static correction, subtract-mean dewow, Bandpassfilter, background removal, gain und Migration) angewandt und die Topographie mit Hilfe von dGPS Koordinaten ausgeglichen. B: Radarfazies und Interpretation von GPR-Profil 3. Legende siehe Abbildung 5.

Fig. 7: GPR-profile 3. Processing (static correction, subtract-mean dewow, bandpassfilter, background removal, gain and migration) and topographic correction of the radargramm were already applied. B: Radarfacies and interpretation of GPR-profile 1. For legend see figure 5.

zwischen 1 m NN und -0,5 m NN und liegt damit im Bereich des heutigen Grundwasserspiegels.

Die Strand-Fazies in Bohrung AL2 zeigt im Vergleich zu Bohrung AL6 insgesamt eine stärkere Variation hinsichtlich der Korngröße und Sortierung. Im Bohrkern von AL6 lässt sich anhand der Korngröße und Sortierung kaum ein Unterschied zwischen Dünen- und Strandfazies feststellen. Es dominieren Fein- und Mittelsande (1,899  $\phi$ -2,003  $\phi$ ). Alle Proben weisen eine mäßig gute Sortierung (0,602  $\phi$ -0,612  $\phi$ ) auf und besitzen eine symmetrische bis leicht positive Schiefe. Hinweise auf ein marines Ablagerungsmilieu geben im Bohrkern AL6 lediglich die ab einer Tiefe von 2 m unter der Geländeoberfläche (ca. -0,8 m NN) vermehrt vorkommenden Muschelschalen.

In einer Bohrtiefe zwischen 3 m-5 m wechseln sich Oxidations- und Reduktionshorizonte ab. Der Wechsel von Oxidation und Reduktion spricht für die zeitweilige Überflutung des Sedimentationsraums und kann daher ebenso als Indiz für ein marines Milieu gedeutet werden.

Der Strand aus Bohrung AL2 besteht aus schillführenden Fein- und Mittelsanden (1,00  $\phi$ -2,13  $\phi$ ), die eine mäßige bis mäßig gute (0,60  $\phi$ -0,97  $\phi$ ) Sortierung und eine unimodale bis bimodale Kornverteilung aufweisen (Abb. 9). Eingeschaltete Schillhorizonte innerhalb der Strand-Fazies sind von schlechter Sortierung (1,35  $\phi$ ) und weisen eine polymodale Kornverteilung auf. Wie auch in Bohrung AL6 sind die Sedimente des Strandbereichs von AL2 durch zahlreiche Oxidations- und Reduktionsspuren gekennzeichnet. Rost-

flecken sind detailliert in Abbildung 9 (Bild 2) hervorgehoben. Überdies sind innerhalb der Strand-Fazies aus Bohrung AL2 immer wieder schwach ausgebildete Wurzelreste nachzuweisen. Im unteren Strandbereich von Bohrung AL2 lassen sich ab einer Bohrtiefe von 3 m zunehmend größere Korngrößen wie mäßig bis gut gerundete Kiese dokumentieren. Auch die Sortierung ist im unteren Strandbereich von AL2 generell schlechter als in Bohrung AL6.

Die Unterschiede der Strandsedimente von AL2 und AL6 lassen sich mit der Nähe bzw. Entfernung zu dem sich im Nordosten anschließenden Geestkliff erklären. So stellt Bohrung AL2 den kliffnahen Strandbereich dar. Dies wird vor allem durch die Anreicherung von Kiesen und Geröllen im unteren Strandbereich des Bohrkerns deutlich. Dieser Bereich entspricht morphologisch dem fossilen Klifffuß des Geestkliffs und ist so im weitesten Sinne als Teil einer Abrasionsplattform anzusehen. Die Kiese stammen ursprünglich aus dem pleistozänen Kliff und wurden durch die Brandung zu gut gerundeten Brandungsgeröllen aufgearbeitet.

Die Wurzelreste aus Bohrung AL2 deuten auf Vegetation hin, die im Vergleich zum Standort von Bohrung AL6 durch ihre Nähe zum Inselgeestkern bzw. ihrer größeren Entfernung zum Meer und der damit einhergehenden geminderten Überflutungshäufigkeit die besseren Lebensbedingungen vorgefunden hat. Es handelt sich dementsprechend um den trockenen Strand im *backshore*-Niveau.

Der sedimentologische Übergang zwischen Strand-Fazies im Hangenden und Watt-Fazies im Liegenden vollzieht sich im Bohrkern nur allmählich und kann durch einen Übergangsbereich beschrieben werden. Der Übergangsbereich setzt sich aus Fein- und Mittelsanden zusammen, die eine ähnliche Sortierung wie die überlagernden Strandsedimente zeigen. Eine eindeutige, allein auf sedimentologischen Kriterien basierende Unterscheidung von Strand-Fazies und Sandwatt-Fazies ist in diesem Abschnitt von Bohrung AL6 nicht möglich. Aufgrund der typischen Reflexionsgeometrien in den GPR-Daten wurde dieser Abschnitt durch die Radarstratigraphie dennoch eher zum Strandbereich gezählt.

Erst ab einer Bohrtiefe von 6,60 m (-3,75 m NN) wird ein deutlicher Wechsel hin zum feinkörnigen Misch- und Schlickwatt deutlich, das sich durch eine sehr fein laminierete Wechsellagerung von Feinsand, Silt und Ton auszeichnet. Das Korngrößenspektrum der Watt-Fazies variiert zwischen feinem Mittelsand (1,89  $\phi$ ) und feinem Silt (7,07  $\phi$ ). Der Anteil an reinem Ton beträgt zwischen 4,2 %–40,7 %.

Die in AL6 erbohrte Watt-Fazies vor der Westküste Amrums lässt sich in ein Schlick- und Mischwatt unterteilen. In Bohrtiefen von 6,60 m–7,35 m wurde ein Schlickwatt mit einem reinen Tonanteil von 13,5 %–40,7 % nachgewiesen. Zahlreiche Gehäuse der Wattschnecke (*Hydrobia ulvae*) sprechen ebenfalls für ein Schlickwatt, das für die zum Makrobenthos zählende und sich als Weidegänger ernährenden Schnecke einen bevorzugten Lebensraum darstellt (OTT 1996; SOMMER 2005). Die Gehäuse der Schnecken sind intakt und gut erhalten. Die graue bis schwarze Sedimentfarbe und das Vorkommen von Mangan-Konkretionen geben Hinweise auf reduzierende Bedingungen (SOMMER 2005) bzw. sind als Beleg für die Reduktionszone im Watt anzuführen (vgl. KÖSTER 1998; OTT 1996; POTT 2003; SINDOWSKI 1973).

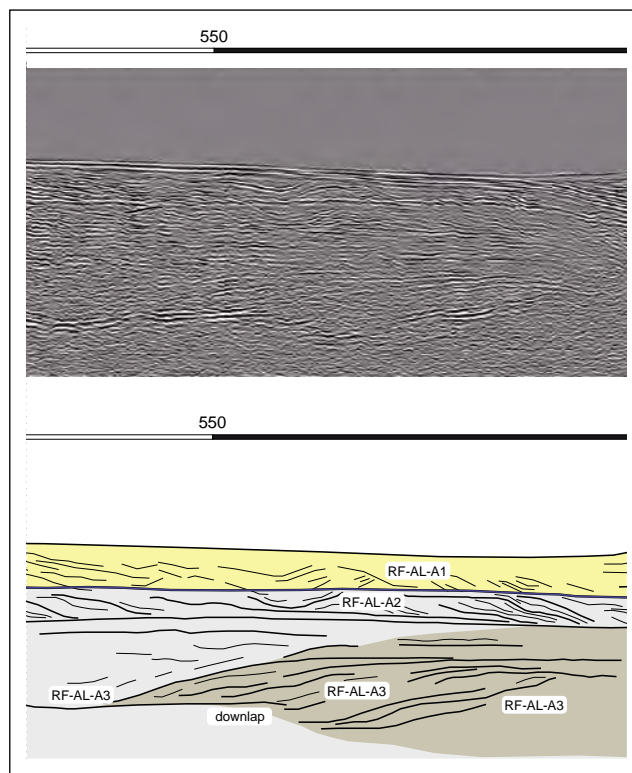


Abb. 8: Vergrößerter Ausschnitt aus GPR-Profil 3. Zur Position des Ausschnitts innerhalb des Profils siehe Abbildung 7. Der Ausschnitt zeigt die ostwärts gerichtete Wanderung des Kniepsandes.

Fig. 8: The close up of GPR-profile 3. For position of the close up see figure 7. The section shows the eastward migration of the Kniepsand.

In Bohrung AL6 wird das Schlickwatt zum Hangenden hin zunehmend grobkörniger und geht ab einer Bohrtiefe von 6,60 m in ein feinsandiges Mischwatt mit einem Tonanteil von unter 5,6 % und einer durchschnittlichen Korngröße von 2,45  $\phi$ –1,89  $\phi$  über. Die Sedimentfarbe wird zum Hangenden hin insgesamt heller (Abb. 10) und deutet auf weniger stark reduzierende Bedingungen bzw. eine bessere Durchlüftung der Sedimente mit Sauerstoff hin.

Die feinen Korngrößen von Feinsand und Silt sowie der insgesamt hohe Tongehalt sprechen für ein geringes energetisches Strömungsregime zur Zeit der Sedimentation. Infolgedessen wird angenommen, dass die Wattsedimente aus Bohrung AL6 aus einer Stillwassersedimentation an einer „Niederenergieküste“ (SCHÄFER 2005: 243) hervorgegangen sind. Solche ruhigen Sedimentationsbedingungen sind heute hauptsächlich im Bereich der Rückseitenwatten und im strömungsberuhigten Lee einer Insel oder Küstenbarriere zu finden.

Heute herrschen an der Westküste Amrums jedoch hochenergetische Strömungsbedingungen. Die erbohrten Wattsedimente müssen demzufolge unter abweichenden, ruhigeren Sedimentationsbedingungen dort zur Ablagerung gelangt sein. Es ist davon auszugehen, dass dies im Lee einer der Westküste vorgelagerten Barriere oder Sandbank geschah. Es wird daher angenommen, dass der damals noch weit vor der Küste Amrums liegende Kniepsand diese Barriere bildete und so an der Westküste für strömungsberuhigte Sedimentationsbedingungen sorgte.

Die Geest-Fazies wurde durch die Bohrungen AL1, AL2 und AL6 direkt erreicht (Abb. 9 und 10). Dabei fällt die Ober-



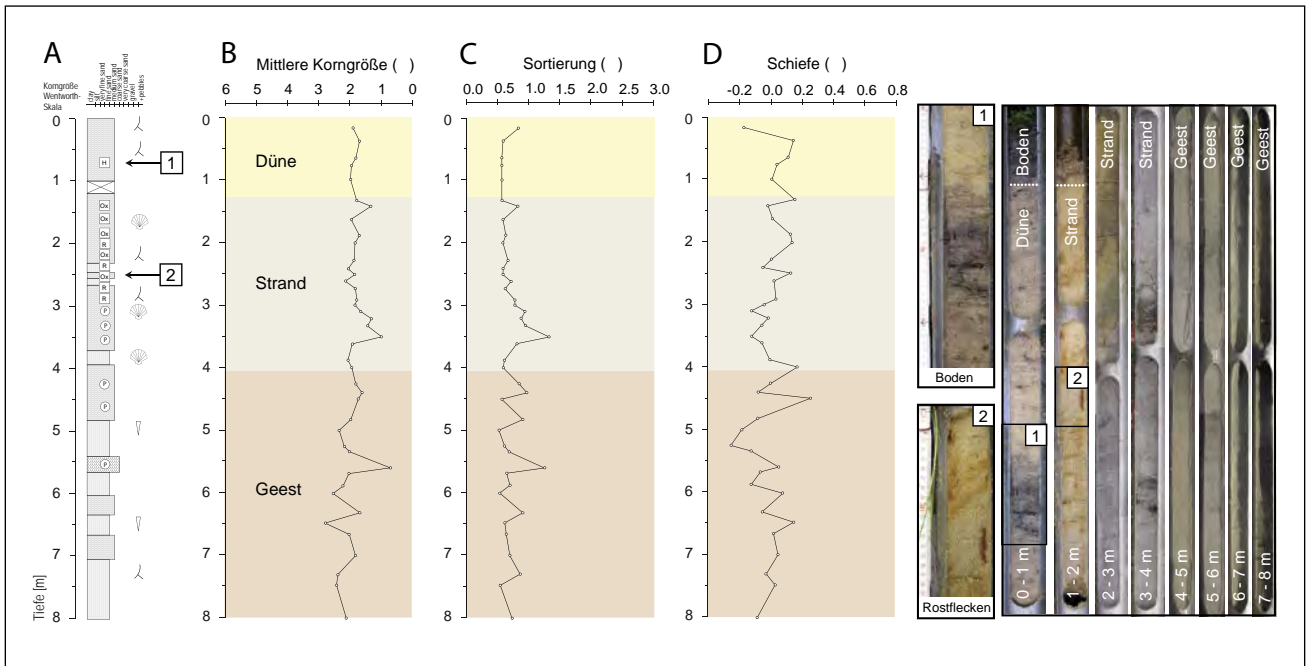


Abb. 9: Bohrkern (A) und Korngrößenstatistik (B – D) von Bohrung AL2. Legende siehe auch Abbildung 10. Bild 1 markiert einen Paläoboden innerhalb der Dünenfazies. Bild 2 zeigt eine Oxidation der Strandsedimente.

Fig. 9: Core log (A) and grain size statistic (B – D) of core AL2. Legend is provided in figure 10. Picture 1 shows a palaeosoil within the dune facies. Picture 2 shows oxidation of beach deposits.

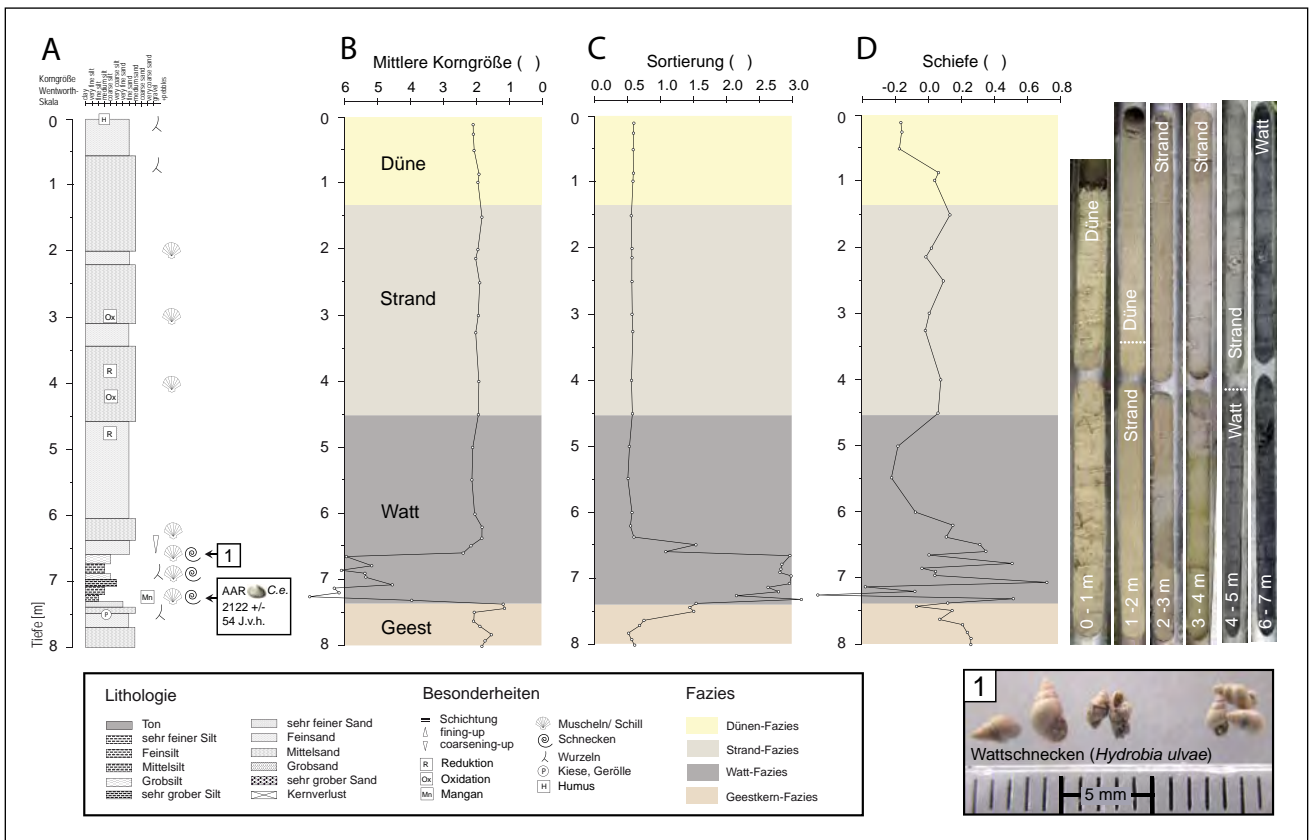


Abb. 10: Bohrkern (A), Korngrößenstatistik (B–D) und Legende von Bohrung AL6. Bild 1: Hydrobia ulvae.

Fig. 10: Core log (A), grain size statistic (B–D) and legend of core AL6. Picture 1: Hydrobia ulvae.

fläche der Geest in den Bohrungen AL1, AL2 und AL6 von Nordosten nach Südwesten ab. Während die Geest in Bohrung AL1 direkt unterhalb der Dünenfazies ansteht und in einer Bohrtiefe von 1,20 m (0,8 m NN) erreicht wurde, fällt die Oberfläche in den Bohrungen AL2 und AL6 auf 4 m und 7,45 m Tiefe ab. Damit steht die Geest in Bohrung AL2 bei -3,15 m NN und in Bohrung AL6 bei -5,25 m NN an.

Die Geestkern-Fazies setzt sich in den Bohrungen AL1, AL2 und AL6 aus stark glimmerhaltigen Fein- und Mittelsanden zusammen. Es dominiert der helle Glimmer (Muskovit). Die durchschnittliche Korngröße der Geest-Sande beträgt 1,34  $\phi$  und 2,77  $\phi$ . Die Sortierung ist mit Werten zwischen 0,95  $\phi$  und 0,52  $\phi$  mäßig bis mäßig gut und ist an der Oberfläche der Geest meist etwas schlechter. Der Übergang von der Geest-Fazies im Liegenden zur Watt-Fazies im Hangenden wird durch eine Anreicherung von groben Kornkomponenten und Kiesen begleitet, die sedimentologisch auf eine gut ausgebildete Steinsohle hinweisen.

Überdies konnten in Bohrung AL6 in einer Tiefe von 7,45 m (-5,25 m NN) Wurzeln nachgewiesen werden. Wurzeln an dieser stratigraphischen Position sprechen für den Bewuchs der Geestoberfläche und damit für ehemals terrestrische Bedingungen an dieser einstigen Landoberfläche. Die Annahme eines terrestrischen Ablagerungsraumes wird durch das Fehlen von Muscheln und Schill unterstützt.

## 5 Diskussion

### 5.1 Coastal onlap Modell

Auf der Grundlage der gemessenen GPR-Profile und durchgeführten sedimentologischen Analysen aus den Bohrungen AL1, AL2 und AL6 wurde ein stratigraphisches Modell entwickelt. Das Modell beschreibt die Landschaftsentwicklung an der Westküste Amrums in einer Zeit, bevor sich der von Westen heranwandernde Kniepsand an die Insel angelagert hat.

Die geologische Ausgangssituation ist ein leicht nach Westen abfallender Geestkern bestehend aus Endmoränenrelikten der Saale-Eiszeit. Von Westen aus folgt die Anlagerung rein mariner Sedimente, die als „*coastal onlap*“ (CATUNEANU 2002; EMERY & MYERS 1996) beschrieben werden kann. *Coastal onlap* Konfigurationen sind vor allem aus seismischen Daten vom Kontinentalhang bekannt (vgl. EMERY & MYERS 1996). In Georadardaten konnte ein *coastal onlap*-Prozess hingegen bislang noch nicht nachgewiesen bzw. unmittelbar durch Radargramme abgebildet werden.

Sedimentologisch besteht der Geestkern im Untersuchungsgebiet aus muskovithaltigen Fein- und Mittelsanden mit vereinzelt eingeschalteten Geschieben, die auf fluvio-glaziale Schmelzwassersande zurückzuführen sind (vgl. auch JESSEN 1932; KRAUSE 1913; MEIER 1987; VOIGT 1969).

Der Übergang zwischen Geestkern im Liegenden und Watt im Hangenden wird durch eine Steinsohle gebildet, die der ehemaligen Landoberfläche entspricht und zugleich die Grenze zwischen Pleistozän und Holozän darstellt. Wurzeln zeugen von terrestrischen Bedingungen und dem einstigen Bewuchs dieser Landoberfläche, die durch einsetzende Transgression allmählich von Westen her überflutet wurde.

Mit der Transgression folgten Ablagerungen eines Watts. Die Wattedimentation im Vorfeld des nach Westen abtauchenden Geestkerns stellt damit die zeitlich erste aus den o. g. Daten ersichtliche Phase dar (Abb. 11 Bild 1). Die Ablagerungen dieses Westküstenwatts konnten in den Daten bis zu einer Tiefe von -7 m NN verfolgt werden. Im Liegenden wurden zunächst horizontal geschichtete Wattedimente eines Misch- bis Schlickwatts abgelagert, die zum Hangenden hin grobkörniger werden und in ein Sandwatt übergehen. Der Sedimentationsraum wird durch Funde von Wattedschneckenhäuse in der Bohrung AL6 bestätigt.

Eine grobe zeitliche Einordnung des Westküstenwatts aus Bohrung AL6 erfolgte lediglich mittels AAR-Datierung an einer intakten Muschelschale der Herzmuschel (*Cerastoderma edule*), die aus einer Bohrtiefe von 728 cm gewonnen wurde. Auf diese Weise konnte für die Wattedimente in einer Tiefe von -5 m NN ein Mindestalter von 2122 +/-54 Jahren definiert werden. Die datierten Wattedimente zählen zu den ersten, die auf dem „ertrinkenden“ Geestkern abgelagert wurden und entsprechen damit der Position des Transgressionskontaktes. Es ist demzufolge davon auszugehen, dass die Überflutung dieses Bereiches der Westküste Amrums bereits vor 2122 +/-54 Jahren einsetzte.

Bereits FALK (2001) geht vor der Westküste Amrums von wattähnlichen Bedingungen aus. Im Vorfeld der Bohrungen AL2 und AL6 wurden nach FALK (2001) im Subboreal ausschließlich marine klastische Sedimente abgelagert, die allerdings bisher nicht näher hinsichtlich ihrer Genese identifiziert, unterteilt sowie räumlich und zeitlich abgegrenzt wurden.

So wurde nach Übereinstimmung mit FALK (2001) das Gebiet der Westküste Amrums zunehmend an der Grenze zwischen Subboreal und einsetzendem Subatlantikum von der Transgression erreicht, sodass mit steigender Überflutung Wattedimente dort abgelagert wurden. Da eine Torfbildung an der Basis der holozänen Wattedimente fehlt wird von einem raschen Vordringen der See ausgegangen. Die transgredierende Nordsee erreicht im Subboreal zuerst über das Tief der Norderaue das Untersuchungsgebiet von Südosten her (FALK 2001). Die Bohrungen AL2 und AL6 lagen im Subboreal noch außerhalb des Überflutungsbereichs.

Aus Chroniken ist zu entnehmen, dass es noch Anfang des 19. Jahrhunderts „grüne Marschwiesen“ weit westlich von Amrum gegeben haben soll, auf denen „Schafe“ weiden konnten (MEYN 1876; MÜLLER & FISCHER 1937; JESSEN 1914). Eine genaue Lagebestimmung und Position dieser „Marschwiesen“ vor der Küste Amrums wurden jedoch nie überliefert. Festgehalten werden kann dennoch, dass im Vergleich zu heute ruhigere Sedimentationsbedingungen im Vorfeld der Westküste Amrums geherrscht haben müssen.

Aus historischen Land- und Seekarten (vgl. Abb. 2) ist hingegen ersichtlich, dass sich der Kniepsand auf der Höhe des Leuchtturms bereits vor 1841 an die Insel angelagert haben muss, sodass zu und nach dieser Zeit und an dieser Stelle vor der Westküste Amrums keine „Schafe“ auf „Marschwiesen“ weiden konnten.

Glaubt man der Landkarte von JOHANNES MEYER aus dem Jahr 1648 (DANCKWERTH 1652), so lag der Kniepsand 1648 als solitäre Sandbank oder Untiefe noch weit vor der

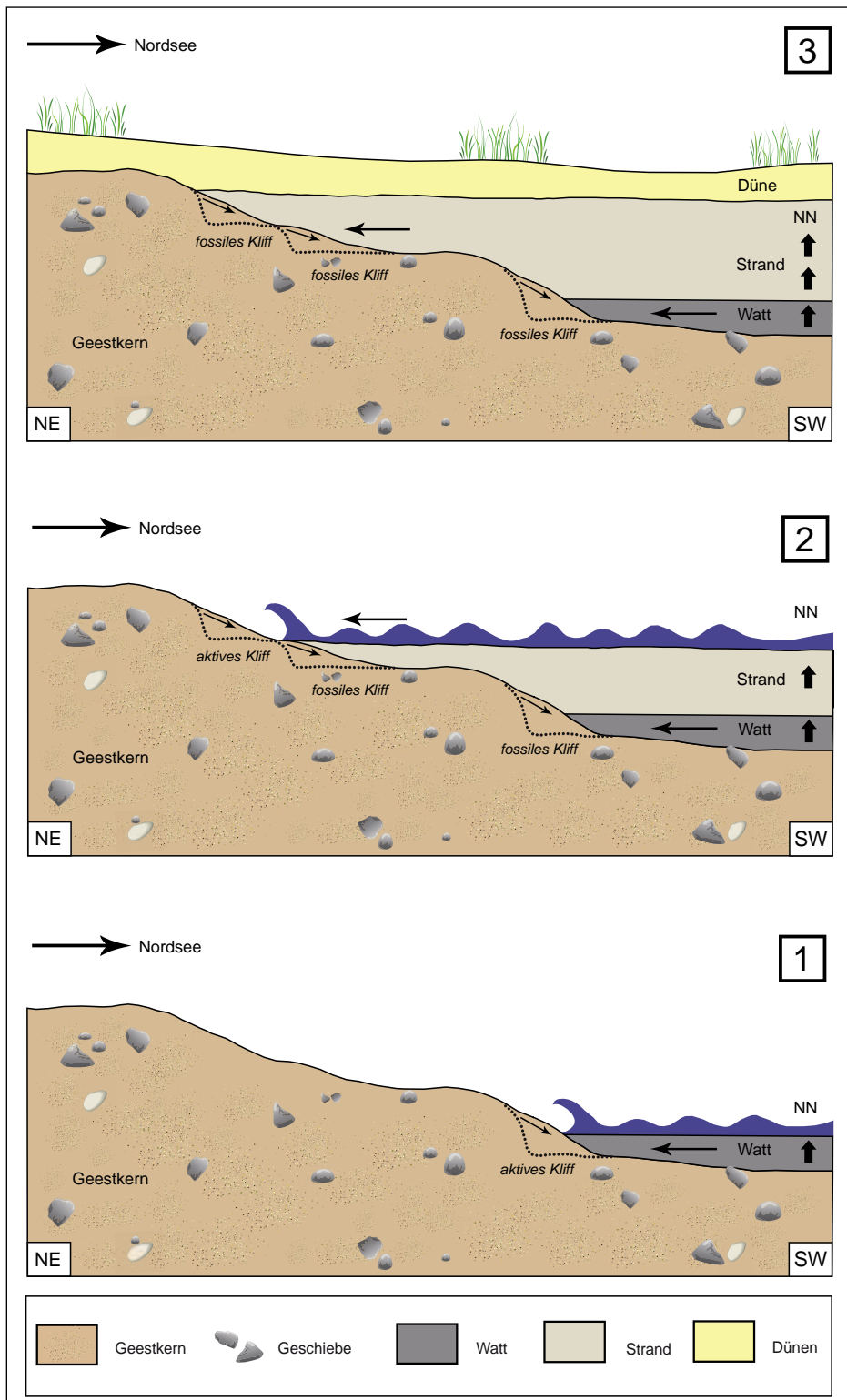


Abb. 11: Das Coastal-onlap Modell zeigt die Landschaftsentwicklung vor der Westküste der Insel Amrum. Bild 1: Im Küstenvorfeld der Insel Amrum wurden zunächst feinkörnige Wattsedimente abgelagert. Bild 2: Durch einen Anstieg des Meeresspiegels und die ostwärts gerichtete Migration des Kniepsandes folgten gröbere Strandsedimente. Sturmfluten führten gleichzeitig zur Erosion am Geestkern und schufen aktive Kliffs, die mit ansteigendem Meeres- und Sturmflutspiegel in fossile Kliffs übergingen. Bild 3: Durch die endgültige Anlagerung des Kniepsandes an die Insel wurde der nötige Sedimentinput für großflächige Dünenbildung geliefert.

Fig. 11: The coastal-onlap Model shows the landscape development in front of the westcoast of Amrum. Picture 1: Before the barrier sandbar was connected to the island, tidal flat deposits had been accumulated in a low energy environment. Picture 2: Tidal flat deposits show a general coarsening upward trend and turn into overlying coarser grained beach deposits under a rising sea level. Old cliffs formed through several storm surges are also preserved in GPR data. Picture 3: Fine to medium grained beach sand was reworked and blown out to build up some primary foredune ridges which enhance to shape the present dune area.

Westküste Amrums. Demnach können in der Zeit von 1648 bis zur Anlagerung des Kniepsandes an die Insel durchaus strömungsberuhigte Sedimentationsbedingungen geherrscht haben, die im Westküstenvorfeld zu Ausbildung von Watt- und Marschflächen führten. Denkbar ist, dass der solitäre Kniepsand seinerseits eine Barriere darstellte und so im Lee die ruhigen Sedimentationsbedingungen an der Westküste Amrums zum Teil förderte.

Eine Anlagerung des Kniepsandes an die Insel wird daher für die Zeitspanne zwischen 1648 und 1841 angenommen.

Im weiteren Verlauf der Transgression wurden die Ab-

lagerungen zunehmend größer. Im Hangenden des Westküstenwatts folgten Strandablagerungen (Abb. 11 Bild 2), die im Vergleich zum Watt eine unruhige Strandmorphologie zeigen. Die Morphologie wird durch vereinzelte in das Strandprofil eingeschaltete intertidale Sedimentkörper (*intertidal bars, swash bars*) und parallel zur Küste verlaufende Rinnen bedingt, die eine *ridge-and-runnel*-Topographie des Strandes hervorrufen.

Die verhältnismäßig große Breite sowie die flache nur leicht in Richtung Meer einfallende Topographie des Strandes und die Anlagerung von intertidalen Sedimentkörpern

lassen auf einen dissipativen Strand schließen. Sturmfluten führten neben Klifferosion am Geestkern ebenso zur Erosion und morphologischen Umgestaltung im Strandbereich. Dabei wurden die intertidalen Sedimentkörper zum Teil ausgeräumt, aufgearbeitet oder in ihrem oberen Bereich gekappt (FITZGERALD 2005; JACKSON et al. 2007), sodass zeitweilig ein ebenes Strandprofil aus horizontalen Strandsandschichten mit ausgeprägter Hochenergie-Parallelschichtung (SCHÄFER 2005) entstand.

Die fossilen Geestkernkliffs zeigen im Radargramm (Abb. 5) deutliche, durch wiederholte Abrasion infolge von Sturmfluten geschaffene Brandungshohlkehlen im unteren Kliffbereich. Gravitative Hangabtragung des oberen Kliffbereichs bzw. ein denudatives Nachrutschen von Kliffmaterial aus dem Hangenden ist anhand von Schutthalden am Klifffuß erkennbar. Diese Kliffhalden sind in den GPR-Daten durch eine Anreicherung von Diffraktionshyperbeln ersichtlich. Auch sedimentologisch lassen sich dem Bereich insgesamt gröbere Korngrößen sowie vereinzelt Brandungsgerölle nachweisen. Damit entspricht dieser dem Kliff vorgelagerte Strandbereich geomorphologisch einer Abrasions- oder Strandplattform.

Bereits MÜLLER & FISCHER (1937: 10) entdeckten westlich des Leuchtturms ein „[...] Steinplaster aus groben, teils faustgroßen Geröllen [...]“, das sie als „[...] Brandungsterasse, die aus den bei der Unterspülung nachgestürzten Geschieben [...]“ im damaligen Kliffbereich gebildet wurde deuteten.

Die Abfolge aus Geestkern, Kliff, Brandungshohlkehle, Kliffhalde und Abrasionsplattform mit Brandungsgeröllen kann als „littorale Serie“ zusammengefasst werden, die allerdings weniger stark ausgeprägt ist. Begründet ist dies auf der Tatsache, dass sich der Geestkern von Amrum an dieser Stelle nur durch gering verfestigte pleistozäne Sande mit Geschiebeeinlagerungen darstellt, der nur bedingt ein Kliff von geringer Höhe ausbildet. Dementsprechend unscheinbar ist auch die dem Kliff vorgelagerte Abrasionsplattform, die sich ebenso aus Lockersedimenten zusammensetzt.

Fossile Kliffs in unterschiedlichen Höhenpositionen im Bereich des Geestkerns lassen auf unterschiedliche Meeresspiegelniveaus schließen. Ein Anstieg des Meeres- und Sturmflutspiegels führte zur Erosion und Kliffbildung am Geestkern. So entspricht die Höhe des jeweiligen Klifffußes dem Sturmflutspiegelniveau der entsprechenden Bildungszeit (Abb. 5 und 11). Das obere fossile Kliff kann als Vorstufe des Littorina Kliffs gesehen werden. REMDE (1972) geht davon aus, dass sich das Nord- und Südende des Littorina Kliffs noch vor etwa 150 Jahren im Abbruch befand und sich erst danach zum fossilen Kliff entwickelte.

Die jüngste stratigraphische Einheit im Untersuchungsgebiet stellen die heutigen Dünen dar, die sich nach Ende der Transgression und mit Anlagerung des Kniepsandes an die Westküste der Insel großflächig bildeten und damit die Geest allmählich bedeckten (Abb. 11 Bild 3).

## 5.2 Andockmodell des Kniepsandes an die Insel Amrum

Auf der Grundlage von GPR-Profil 3 konnte ein zweites Modell entwickelt werden, das die Migration und letztendlich das „Andocken“ des Kniepsandes an den zentralen Insel-

geestkern an der Westküste Amrums beschreibt (Abb. 12).

Generell zeigt der ehemalige Weststrand der Insel Amrum noch vor Anlagerung des heutigen Kniepsandes eine leicht nach Südwesten einfallende Schichtung im Unterwasserbereich (Abb. 12 Bild 1). Auf westlicher Seite erfolgten das Heranwandern und die Annäherung des ehemals als solitäre Sandbank vor der Westküste lagernden Kniepsandes (Abb. 12 Bild 2). Die nach Nordosten gerichtete Wanderungstendenz des Kniepsandes zeigt sich in GPR-Profil 3 besonders deutlich zwischen den Profilmeter 540–610 (Abb. 7 und 8). Die sigmoidale Form der Reflexionen sowie die tangentielle Beziehung der Reflektoren deuten dort auf eine progradierende Schichtung der Sedimente hin. Die in Radarschichtpakete gebündelten Reflexionen lassen auf eine strömungsbedingte Anlagerung einzelner sandiger Sedimentpakete schließen. Die Anlagerung von Sedimenten an die „Stirn“ des Kniepsandes erforderte einen strömungsbedingten Sedimenttransport in östliche Richtung.

Zeitlich erfolgte die Bildung des ursprünglichen Strandes an der Westküste von Amrum vor der Anlagerung des Kniepsandes an die Insel. Stratigraphisch ist der Kniepsandkörper in die ursprünglichen Sedimente des Strandes vor der Westküste Amrums eingearbeitet (Abb. 12 Bild 2).

Die deutlich erkennbare *downlap*-Konfiguration aus den GPR-Daten zeigt, dass sich der Kniepsandkörper an seiner „Stirn“ auf die Strand- und Vorstrandsedimente heraufgeschoben hat. Der Bereich zwischen den Profilmeter 540 - 610 wird infolgedessen als „Andockzone“ interpretiert und stellt damit geologisch die Grenzen zwischen der Insel und dem Kniepsand dar. Die Zone zwischen dem ursprünglichen Strand und dem heranwandernden Kniepsand zeigt im Radargramm nur wenige Reflexionen. Gründe dafür können ein homogener Sedimentkörper bzw. eine nur undeutlich bis wenig ausgeprägte Schichtung innerhalb der Mischzone der Andockstelle sein. Die Vermischung der beiden Sedimentkörper wurde in der Folgezeit durch strömungsbedingte marine Vorgänge im Unterwasserbereich sowie durch äolische Vorgänge im trockenen Bereich des Strandes gefördert (Abb. 12 Bild 3).

Nach Ende der Transgression bzw. mit dem Trockenfallen der entstandenen Sandflächen wurden die Bedingungen für die anschließende großflächige Dünenbildung geschaffen. Höchstwahrscheinlich führte die Anlagerung des Kniepsandes im Westen zum Sedimentinput und stellte den Sand zur vermehrten Dünenbildung zur Verfügung (Abb. 12 Bild 4).

Die Dünen generierten sich aus dem Kniepsand, wanderten über den Strand hinweg, begruben das Geestkliff und schoben sich den flachen Geestkern hinauf. Auf diese Weise wurde das Geestkliff vor der Brandung und vor fortschreitender Abrasion geschützt und entwickelte sich allmählich zum toten Kliff.

Morphologisch ist das Geestkliff im Untersuchungsgebiet als verdeckter Teil des weiter nördlich bei „A Hörn“ oberflächennah aufgeschlossenen Littorina Kliffs zu sehen und steht mit diesem in unmittelbarer Verbindung. Bei „A Hörn“ („die Ecke“) erreicht das Littorina Kliff mit 13 m NN heute seine größte Höhe und taucht in Richtung Leuchtturm relativ kontinuierlich ab (REMDE 1972).

Auch die Dünen auf dem Geestkern Amrums wurden daher vermutlich erst nach Anlagerung des Kniepsandes

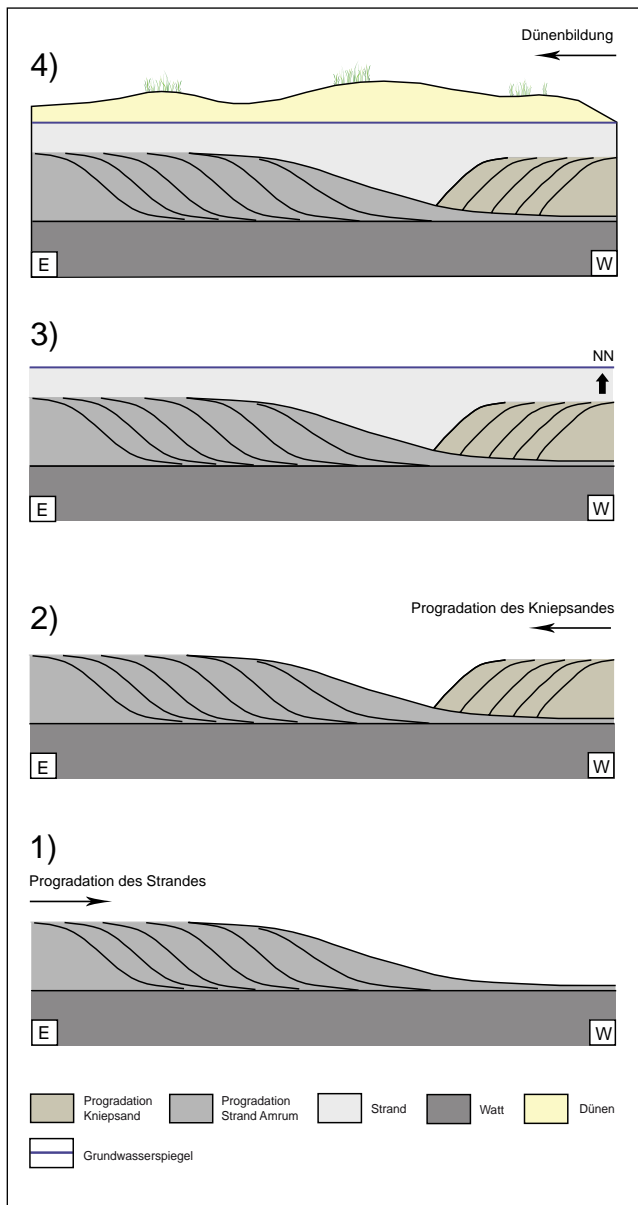


Abb. 12: Andockmodell des Kniepsandes an die Insel Amrum. Bild 1 zeigt den ursprünglichen Strand an der Westküste der Insel Amrum. Bilder 2–3 zeigen das Heranwandern des Kniepsandes an die Insel unter steigendem Meeresspiegel. Bild 4: Anschließend setzte durch vermehrten Sedimentinput großflächig Dünenbildung ein.

Fig. 12: Picture 1 shows the early beach in front of the west coast of Amrum. Pictures 2–3 show how the Kniepsand migrates to the island under a rising sea level. Picture 4: The input of sediment leads to dune building.

an die Insel gebildet. Zeitliche Anhaltspunkte für die Dünenbildung liefern die „mittelalterlichen Ackerbeete“ die als Kulturspuren unter den Dünen der Geest hervortreten (GRIPP 1968). Unweit des Amrumer Leuchtturms wurden zudem Pflugfurchen, Viehspuren und Ackerbeete entdeckt, deren Kultivierung gemäß der darin enthaltenen Scherben auf das 14. Jahrhundert zurückgeht (KERSTEN 1969; REMDE 1972). Weiter zeitlich eingrenzen lässt sich die Anlagerung des Kniepsandes an die Insel Amrum unter Beachtung historischer Land- und Seekarten. So wird ein Andocken des Kniepsandes an die Insel auf der Höhe des Untersuchungsgebietes für die Zeitspanne zwischen 1648 und 1841 angenommen.

In seiner „Chronik der friesischen Uthlande“ vermerkte C. P. HANSEN (zitiert nach MÜLLER & FISCHER 1937: 65) für das 16. Jahrhundert jedoch: „[...] Amrum war damals wie jetzt ein mehrenteils dürres Land von Heide und Dünen, klein an Fläche, aber mit großen, weit hinausreichenden, für die Seefahrer gefahrvollen Sandbänken umgeben [...]“ und schilderte auf diese Weise morphologisch die Situation sowohl des noch unweit vor der Küste liegenden Kniepsandes als auch der einsetzenden Dünenbildung. Auch HANSEN (1835: 484) teilte in einem Bericht über die Pastoratsländereien von 1629 mit: „[...] schon damals war Wiesenland durch Dünen verloren gegangen [...]“ und beschrieb damit die Auswirkungen der Dünenbildung.

Der Kniepsand stellt heute aus geomorphologischer Sicht den Strand an der Westküste der Insel Amrum dar. Dieser erstreckt sich heute südwestlich des ersten Dünenbelt. Die ursprüngliche aus den Georadar-Daten ersichtliche „Andockstelle“ des Kniepsandes an die Insel befindet sich jedoch 50 m nordöstlich der Vordüne und liegt damit heute bereits im rückwärtigen Dünenengelände. Damit befindet sich die Andockstelle bzw. die Nahtstelle zwischen Kniepsand und eigentlicher Insel etwa 200 m weiter nordöstlich als dies heute der Fall ist.

## 6 Zusammenfassung und Schlussfolgerung

Georadmessungen, Bohrungen und sedimentologischen Analysen haben sich als geeignete Methodenkombination zur Rekonstruktion der Landschaftsentwicklung an der Westküste der Insel Amrum erwiesen. Auf Grundlage der gewonnenen Daten wurden zwei stratigraphische Modelle entwickelt, welche die geologisch-geomorphologischen Prozesse und Sedimentationsbedingungen im Westküstenvorfeld erklären.

Das erste Modell zeigt die Landschaftsentwicklung an der Westküste Amrums und beschreibt die Sedimentationsbedingungen, die im Vorfeld der Westküste herrschten, bevor der Kniepsand an die Insel heranwanderte. Die geologische Ausgangssituation stellt der leicht nach Westen abfallende Inselgeestkern bestehend aus Endmoränenrelikten der Saale-Eiszeit dar. Von Westen aus folgte eine Anlagerung rein mariner Sedimente (*coastal onlap*).

Auf der Landoberfläche des ertrinkenden Geestkern wurden zu Beginn der Transgression feinkörnige Sedimente eines Misch- bis Schlickwatts abgelagert.

Diese zeugen von Stillwassersedimentation und ruhigeren Strömungsbedingungen als diese heute an der hochenergetischen Westküste Amrums herrschen. Es ist davon auszugehen, dass der damals noch weit vor der Küste Amrums liegende Kniepsand eine Barriere bildete und so an der Westküste für strömungsberuhigte Sedimentationsbedingungen sorgte. Der Transgressionskontakt konnte mit Hilfe einer mittels der AAR-Methode auf 2122 +/-54 Jahren datierten Muschelschale der Gattung *Cerastoderma edule* bestimmt werden. In Übereinstimmung mit FALK (2001) wurde das Gebiet der Westküste Amrums zunehmend an der Grenze zwischen Subboreal und einsetzendem Subatlantikum von der Transgression erreicht.

Im Hangenden des Westküstenwatts folgten Ablagerungen eines Strandes, der eine ausgeprägte *ridge-and-runnel*-Topographie aufgrund von zahlreichen in das Strandpro-

fil eingeschalteten intertidalen Sedimentkörpern aufweist. Durch Sturmfluterosion am Geestkern konnten sich Brandungshohlkehlen ausbilden, die zur Kliffbildung führten. Fossile Kliffs in unterschiedlichen Höhenpositionen lassen auf unterschiedliche Meeresspiegelniveaus schließen.

Das zweite Modell beschreibt die Migration und den Andockmechanismus des der Insel im Westen vorgelagerten Kniepsandes an Amrum. Es erfolgte eine strömungsbedingte Anlagerung von Sedimenten an die „Stirn“ des Kniepsandkörpers. Dies führte allmählich zur Progradation und zum Heranwandern des Kniepsandes an die Insel. Dabei schob sich der Kniepsand zum Teil auf die zuvor an der Westküste Amrums abgelagerten, stratigraphisch älteren Strand- und Vorstrandsedimente hinauf. Die einstige Andockstelle befindet im Vergleich zu heute etwa 200 m weiter im Nordosten. Eine Anlagerung des Kniepsandes an die Insel wird für die Zeitspanne zwischen 1648 und 1841 angenommen.

Mit dem Trockenfallen der entstandenen Sandflächen nach Ende der Transgression wurden die Bedingungen für die anschließende großflächige Dünenbildung geschaffen, die durch den Sedimentinput aus dem Kniepsand zunehmend generiert werden konnten. Mit fortschreitender Dünenbildung wurde das Geestkliff vor der Brandung und vor fortschreitender Abrasion geschützt und entwickelte sich allmählich zum toten Kliff.

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# The inventories of archaeological horizons 4 and 3 and the loess section of Grub/Kranawetberg, a Gravettian site in Lower Austria

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## Abstract:

Excavations at the Gravettian site Grub/Kranawetberg from 1993 to 2011 exposed four archaeological horizons (AH) separated by sterile loess deposits. The lowest AH 4 contains a number of features, consisting of two hearths which both are surrounded by small pits. The overlying AH3 is separated from AH4 by 8 to 10 cm of loess. In AH3 there are no features. Above AH3 there are two more AHs (AH2 and AH1) with significantly lower find densities.

Chronologically both AH4 and 3 are very close (ANTL-WEISER et al., 2010) but there are big differences in the assemblages and the presence/absence of features. According to the present state of research the assemblages of the two AHs seem to reflect the presence of different groups using this territory possibly under changing environmental conditions. In 2010 and 2011 a series of samples for an IRSL- and OSL-dating program (ZÖLLER et al., this volume) has been collected from two deep trenches in the east of the excavated area. Results suggest an occupation of AH4 to AH1 between 30ka BP and 27ka BP. The site gives an insight not only into an important part of the cultural development before the Last Glacial Maximum but possibly also into climatic changes during a longer time span of the Upper Pleniglacial period in this part of Austria.

## Das Inventar der archäologischen Horizonte 4 und 3 und das Lössprofil von Grub/Kranawetberg, einem Lagerplatz des Gravettien in Niederösterreich

## Kurzfassung:

Die Ausgrabungen an der Gravettienfundstelle Grub/Kranawetberg bei Stillfried erbrachten vier archäologische Horizonte. Der unterste (AH4) enthält zwei Herdstellen, die beide von einer Reihe von Grübchen umgeben sind. Der darauf folgende Horizont (AH3) ist durch eine 8–10 cm mächtige Lössschicht von AH4 getrennt. In AH3 gibt es keine evidenten Strukturen. Über AH3 liegen zwei weitere archäologische Horizonte mit lediglich einigen verstreuten Funden. Chronologisch sind beide Horizonte zwar nahe beisammen, sie sind aber sehr unterschiedlich, was die Strukturen und das Inventar angeht. Zum gegenwärtigen Zeitpunkt scheinen die beiden Horizonte die Anwesenheit von zwei Gruppen widerzuspiegeln, die dieses Gebiet möglicherweise unter sich ändernden Umweltbedingungen nutzten.

In den Jahren 2010 und 2011 wurde eine Reihe von Proben für IRSL und OSL Datierungen an zwei langen Profilen im Osten der ausgegrabenen Fläche genommen. So gesehen gibt der Platz nicht nur Einblick in einen bedeutenden Abschnitt der kulturellen Entwicklung vor dem Kältemaximum der letzten Eiszeit, sondern auch in die klimatischen Veränderungen des Hochglazial in diesem Teil von Österreich.

## Keywords:

*Gravettian, cultural changes, raw material, settlement structures, personal adornments*

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## 1 Introduction

After surface collections of Palaeolithic artefacts on the fields of the Kranawetberg, a hill west of the village Grub near Stillfried in the March valley in the northeast of Lower Austria (Fig.1), the Prehistoric Department of the Natural History Museum Vienna started archaeological investigations in 1993 (ANTL et al. 1997). From 1993 to 2011 excavations at the site Grub/Kranawetberg exposed two different areas of activity (ANTL & FLADERER 2004). First, from 1993 to 1995, a bone accumulation with remains of mammoth, rhino, giant deer, wild horse and reindeer was excavated (ANTL et al. 1997; BOSCH et al. 2012). Then, approximately 20 metres to the east an area with hearths and high find

density was exposed (ANTL et al. 1997; ANTL & FLADERER 2004). The focus of this paper shall be put on this second area.

## 2 Methods

**Fieldwork methods:** First, we removed the recent top soil (30 to 40cm) and 120 to 180 cm of sterile loess covering the deposits containing the AHs 1 to 4. Then, AHs were excavated following the lithostratigraphic boundaries. All objects larger than 0.5cm were recorded three-dimensionally, i.e. piece-plotted. Until 2001, this was done manually by drawing maps in scales of 1:10 and 1:5 to record the location, inclination and position (dorsal or ventral down) of all find



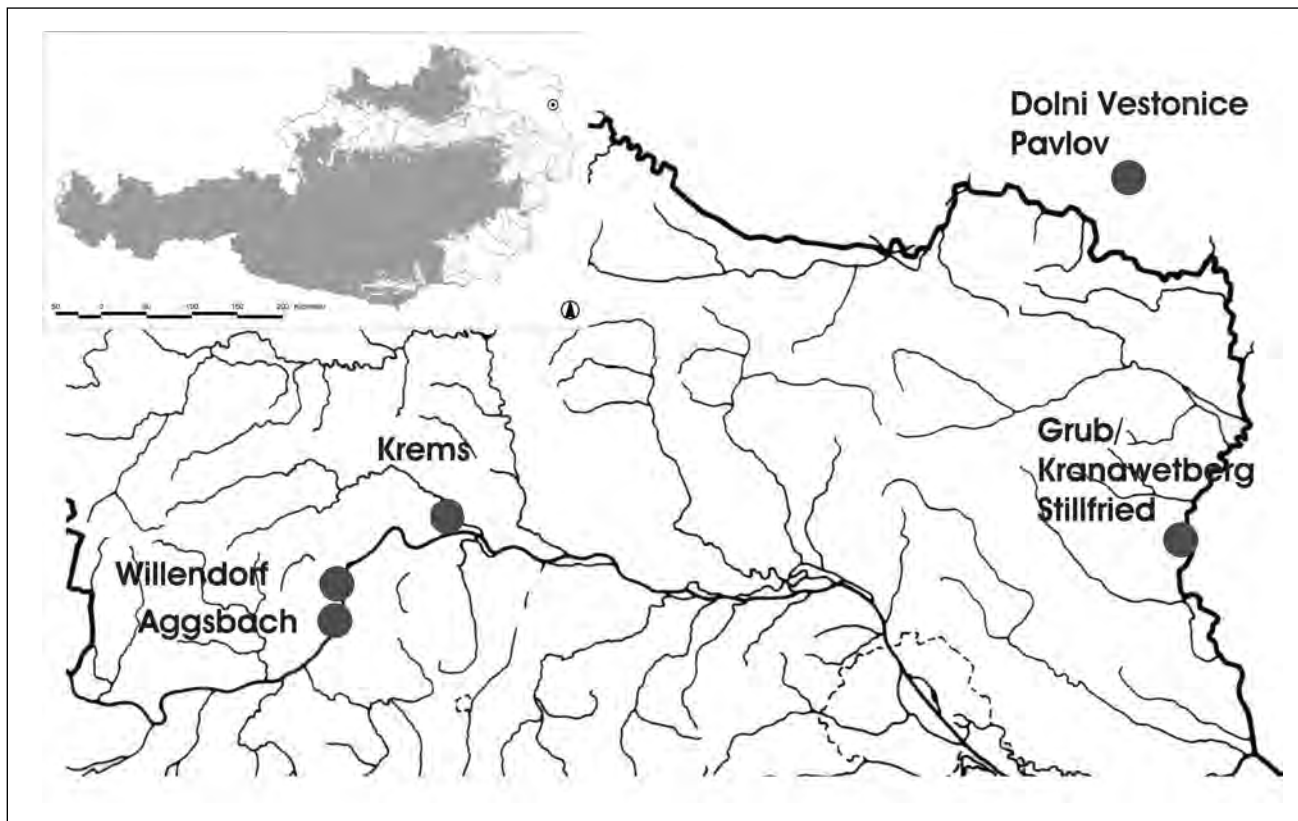


Fig. 1: Map of Austria with the geographical position of Grub/Kranawetberg (small inset) and a more detailed view to the north of Lower Austria and Southern Moravia with the geographical position of Grub/Kranawetberg and other sites mentioned in the text.

Abb. 1: Österreichkarte mit der geographischen Lage von Grub/Kranawetberg (kleines Rechteck) und eine Detailansicht des nördlichen Niederösterreich und Südmähren mit der Lage von Grub/Kranawetberg und anderen im Text erwähnten Fundstellen.

categories. Starting in 2002, we implemented a digital documentation system using totalstations, a field database, and rectified photos (NIGST et al. 2004; 2010). The final product is an interactive database and GIS of the site. Additionally to the digital excavation documentation system, we kept a separate diary (by the archaeologist in charge of the excavation) containing various sorts of further remarks, drawings with the position of measured points, finds in special context and instant remarks concerning structure of the sediment and discussion of the situation as a whole. Parallel to the measurements minutes are kept containing the reference points and the degree of possible deviations. All sediment removed is collected per quarter-square meter and wet-sieved with a grid size of 1 mm.

**Laboratory methods:** As to the lithic raw material microscopic analysis of a reference sample of one thousand pieces was carried out by Alexander Binsteiner. As a first step to a detailed technological analysis a preliminary type list of the lithic material has been established. As far as the fauna is concerned a detailed analysis of the faunal remains of the bone accumulation was made by F. Fladerer and M. Bosch (ANTL-WEISER, FLADERER, NIGST & VERPOORTE 2010; BOSCH, NIGST, FLADERER & ANTL-WEISER 2012). The malacological analysis was based on sediment samples from series at the western periphery of the settlement structures. Different types of molluscs and their preferred habitat were described and the number of each species was listed by F. Stadler (ANTL-WEISER, FLADERER, PETICZKA, STADLER & VERGINIS 1997).

In the East of O16 and O19 series of sediment samples from 1.5 to 4m below surface were taken (Fig. 2). The samples for IRSL and OSL dating were taken parallel to the sediment samples in O16 and O19.

### 3 Results

Within the area with hearths four AHs can be distinguished. Most of the archaeological material is concentrated within AH 4 and 3. Both AHs are clearly separated by 8 to 10 cm of loess; only in the area of hearth II (Fig.3) AH 4 and 3 are extremely close.

All features like hearths and pits belong to AH4. Hearth I and hearth II show repeated use and seem to be covered by AH4 (Fig. 4). As the distribution of finds in AH4 corresponds to the distribution of features (NIGST, 2003; 2004; NIGST & ANTL-WEISER 2012) it is clear that the features and AH4 constitute a unit but it will be a matter of further discussions in which way post-depositional processes possibly contributed to this impression. Above the shallow depression of hearth II there are several rather big bones on the top of AH4, which were not completely covered by loess and therefore must have been visible during the initial occupation of AH3.

Both hearths are surrounded by a series of small pits. On the bottom of these pits there are often rather big stone tools or stones with traces of grinding. The areas with pits are approximately 5 to 6 m in diameter and about 7.5 to 8 m apart from each other (Fig. 3). Between the two concentra-



Fig. 2: Grub/Kranawetberg: Sequence of AHs and loess in O19 with sample locations for pollen analysis on the left and sedimentological analysis on the right.

Abb. 2: Ostprofil in O19 mit Spuren der Probenahme für Pollenanalyse links und für Sedimentanalyse rechts.

tions of pits there is a continuous high density of finds. The base of AH4 is slightly folded in this area as a consequence of periglacial processes. Apart from the hearths different traces of fire as burned sediment and big pieces of charcoal burned *in situ* could be detected, especially in the north and northwest of the eastern concentration.

Around hearth I in the west of the excavated area a sudden decrease of find density has been observed 0.5 to 1 m outside the pits to the north, west and south, whereas the find density around hearth II is only decreasing to the north. South of the pits of hearth II there is an area with

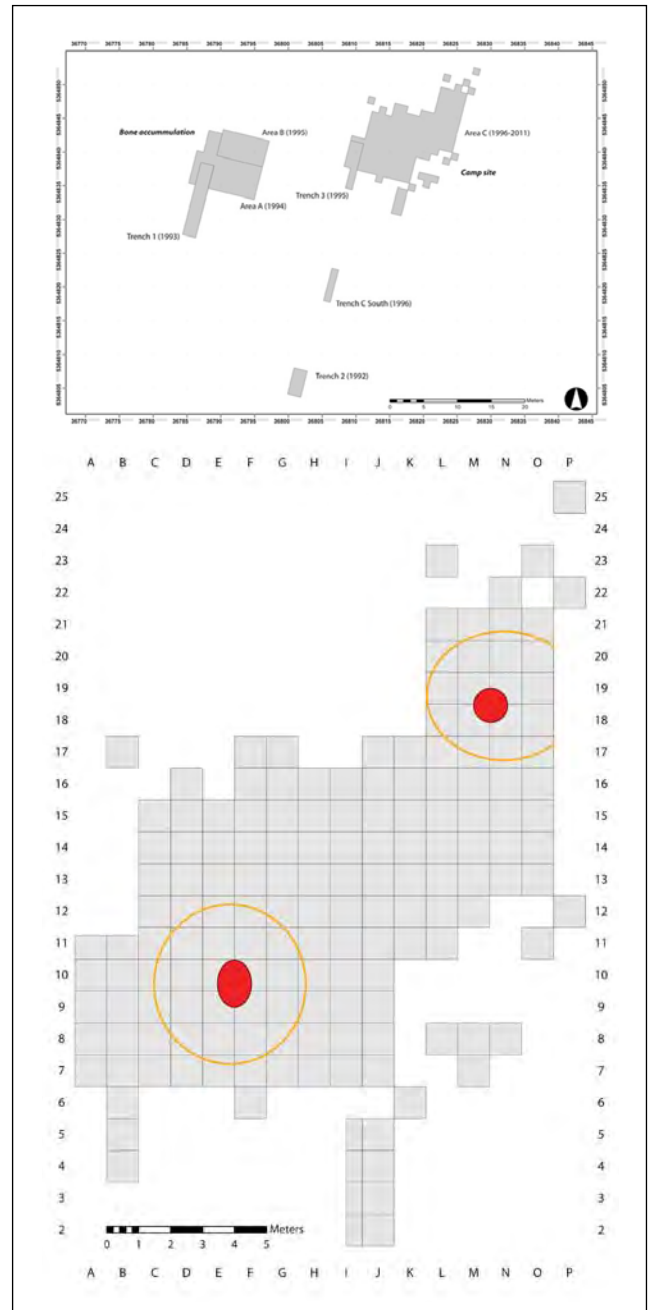


Fig. 3: Grub/Kranawetberg: General map of the excavated area (small inset) with a more detailed view of the area with settlement structures after Ph. R. Nigst, showing the two hearths (in red) and the area with pits around the hearths (in orange).

Abb. 3: Übersichtsplan des ausgegrabenen Bereiches (kleines Rechteck) mit einer Detailsicht des Bereiches mit den Siedlungsstrukturen nach Ph. R. Nigst, die die beiden Feuerstellen (Rot) und den Bereich der Grübchen um die Feuerstellen (Orange) zeigt.

a lower find density than in the concentration itself. The eastern and western border of this concentration has not been fully excavated. In some places AH4 is separated by thin lenses of loess into two small layers which indicates that AH4 consists at least of two occupations, the hearths even show up to four periods of use, which is clearly visible by the superposition of four different phases of firing.

The centre of overlying AH3 is only 2 m to the NW of AH4's hearth II. In AH3 there are no evident structures. Below the centre of AH3 there are traces of fire like burnt bones, large (4–5 cm) pieces of charcoal and patches of

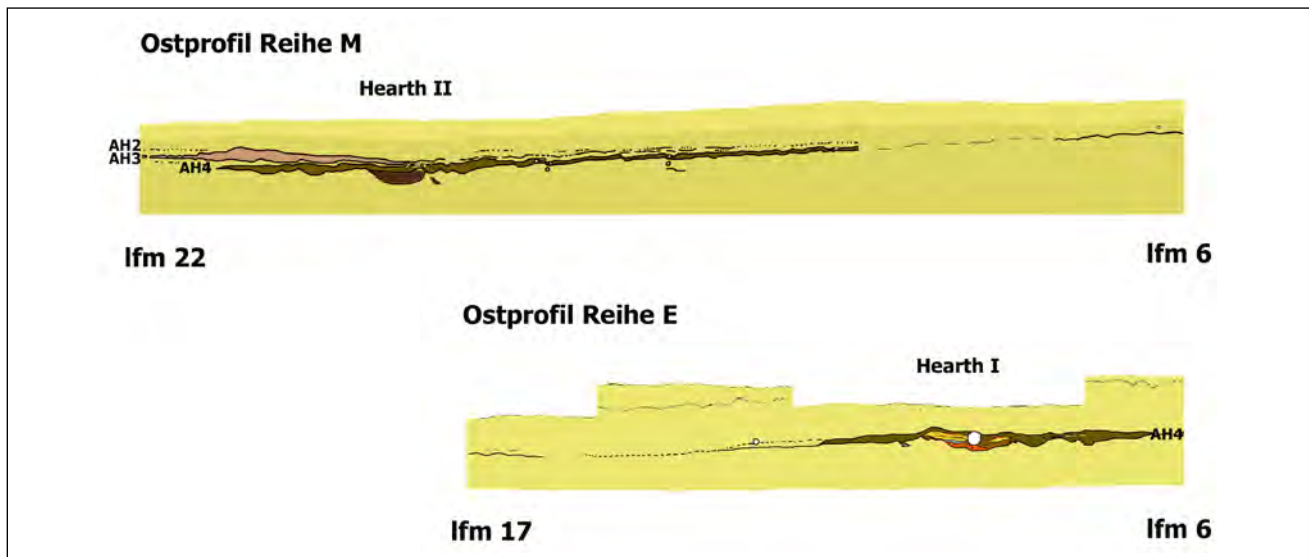


Fig. 4: Grub/Kranawetberg: Section through hearth I in the east of row E from south to north and through hearth II in the east of row M from south to north.  
 Abb. 4: Schnitt durch Feuerstelle I, Ostprofil Reihe E von Süd nach Nord und Schnitt durch Feuerstelle II, Ostprofil Reihe M von Süd nach Nord.

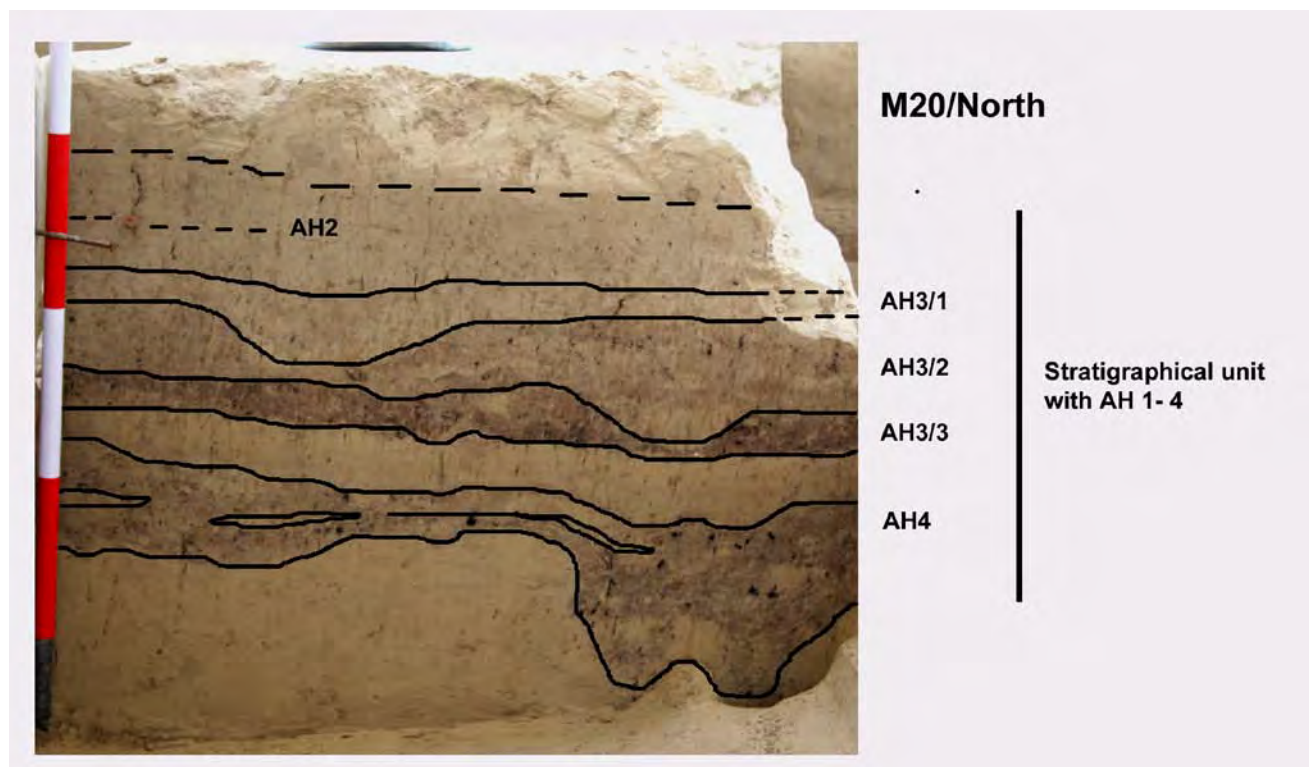


Fig. 5: Grub/Kranawetberg: North-section of square M20 showing the stratigraphic positions of AH1 to AH4.  
 Abb. 5: Nordprofil von M20 mit der stratigraphischen Lage von AH 1–4.

burnt sediment, but there no structured hearths comparable to those of AH4 have been found. The occupation of AH3 covers the whole excavated area. In the west (row A to G) (Fig. 3) AH3 is a horizon with only some scattered bones and artefacts. From row I (Fig. 3) to the east and especially to the northeast find density in AH3 increases. AH3 is associated with a clearly visible light brown horizon, which can be distinguished from the loess, from row J to the east and row 10 to the north (Fig.3). Within square M21 (Fig. 3), AH3 is about 16 cm thick and decreases dramatically in thickness in M22 to approx. 5 cm. In M21 AH3 can be divided into three different events of occupation (Fig. 5).

Above AH3 there are two more archaeological horizons (AH2 and AH1) with only some scattered finds. These horizons yielded individual finds in the western part of the excavated area, e.g. a fragmented mammoth tusk in square G11. In the east (squares L-O/19-21) especially AH2 yielded a more regular scatter of finds. Above AH1, the sediment structure changes considerably. Whereas the layer containing the archaeological horizons with a maximum thickness of 0.8 m shows many traces of organic activity the loess above is characterized by a brighter colour and higher density of the sediment.

Tab. 1: Grub/Kranawetberg: radiocarbon dates for AH 4 and 3. Abbreviations: AH: Archaeological Horizon, lab.nr.: radiocarbon dating laboratory number.

Tab. 1: Grub/Kranawetberg: Radiokohlenstoff-Alter für AH4 und 3. Abkürzungen: AH: Archäologischer Horizont, lab.nr.: Labornummer des Radiokohlenstoff-Datierungslabors.

AH	material	pre-treatment	lab.nr.	<sup>14</sup> C date	1 sigma	reference
AH3	charcoal	ABA	GrA-28183	24,780	+/-140 BP	ANTL-WEISER ET AL., 2010
AH3	charcoal	ABA	GrA-28184	25,640	+/-160 BP	ANTL-WEISER ET AL., 2010
AH3	charcoal	ABA	GrA-28185	25,010	+/-150 BP	ANTL-WEISER ET AL., 2010
AH4/hearth I	charcoal	ABA	GrA-9066	24,830	+/-230 BP	ANTL-WEISER ET AL., 2010
AH4/post hole	charcoal	ABA	GrA-9065	24,930	+/-240 BP	ANTL-WEISER ET AL., 2010
AH4	charcoal	ABA	GrA-9063	24,620	+/-230 BP	ANTL-WEISER ET AL., 2010
AH4	charcoal	ABA	VERA-364	25,300	+/- 90 BP	ANTL-WEISER ET AL., 2010

The radiocarbon dates suggest that AH4 and AH3 had been deposited within a relatively short interval (Table I). The above mentioned list shows that the dates of AH4 and AH3 overlap clearly. Statistically the dates of AH3 appear even older than those of AH4. This is partly due to one rather old date from AH3, which is the oldest one of all. One reason for this discrepancy might be found in the relatively small number of items dated so far. Also the use of collected wood of different ages cannot be excluded and has a strong effect on a small series. Nevertheless on the base of given data we have to expect a rather short time span between the two horizons. In spite of the rapid succession of AH4 and AH3 there are big differences concerning the archaeological material of the two AHs. Apart from differences concerning the presence/absence of features, the raw material used for lithic artifacts is completely different. From AH4 we have mainly flints and cherts whereas from AH3 radiolarites are the most abundant raw material type. AH4 and AH3 of Grub Kranawetberg yielded more than 1000 formal lithic tool types, between 4000 and 5000 blades and bladelets and thousands of flakes. At the present state of analysis it is not yet possible to give a detailed list of formal tools of each AH. This is due to the fact that between 1995 and 2001 artefacts had been drawn manually on plans which are not yet fully digitalized. Nevertheless an estimate of the distribution of formal tools from AH1 to AH4 might already give an impression of the high density of finds in AH4 (Fig. 6). Among the formal tool types of AH4, microgravettes are dominant; in AH3 there are mainly backed bladelets and small bladelets with fine retouching. Two complete and one fragmented notched backed point have been excavated in AH4.

Until now we found 245 beads and pendants of ivory, which are restricted to AH4. The number of ivory objects is still incomplete because the majority of the tiny beads (most of them 6–8 mm) was found within the remains of the sediment after wet sieving. As there are still a lot of samples yet to be sieved and a big amount of remains to be separated into the different find categories (silex, stone, bones, grains of coloured material, charcoal etc. which are smaller than 0.5 cm) a detailed list of bead types at the

present state of analysis would be premature. Personal adornments made of mollusc shells (174 pieces) as well as remains of colour are abundant in both AH3 and AH4. Also tools of antler and bone are present in AH4 as well as in AH3. Worked pieces of ivory are more typical for AH4. Although the analysis of the faunal remains has yet to come, it seems that there is a great abundance of mammoth in AH4 and of reindeer and horse in AH3.

#### 4 Discussion

Chronologically both AH4 and AH3 are very close (ANTL-WEISER et al. 2010), but there are big differences concerning structure and contents. Compared to the settlement structures in AH4 we have only a concentration of finds with a clear centre in AH3. There are evident differences in raw material procurement and in personal adornments. Remains of mammoth seem more abundant in AH4 but detailed zooarchaeological analyses concerning preferred game have yet to be conducted.

First analyses of spatial organization around hearth I (NIGST 2004; NIGST & ANTL 2012a; 2012b) showed different zones of artefact distribution. One zone with a high density of artefacts was interpreted as the place of a dwelling while the other zones were interpreted as the periphery of it. This interpretation was also supported by the distribution of pits around hearth I. As to hearth II there are similar structures, which have yet to be analyzed. Comparable structures with central hearths and pits have been excavated in Dolní Věstonice II, occupation unit 3 (Southern Moravia) (SVOBODA 2003; NIGST & ANTL 2012) and during recent excavations at Krems Wachtberg around the big central hearth (Lower Austria) (EINWÖGERER, et al. 2006; EINWÖGERER 2010; NIGST & ANTL 2012).

A very interesting aspect is that there is evidence for AH4 to be contemporaneous with the bone accumulation (BOSCH 2009; BOSCH et al. 2012). This is based on the fact that an upper left first molar of mammoth, found in the AH4, belongs to the same maxilla as a right upper first molar found at the bone accumulation. This observation fits very well with the intense use of ivory in AH4.

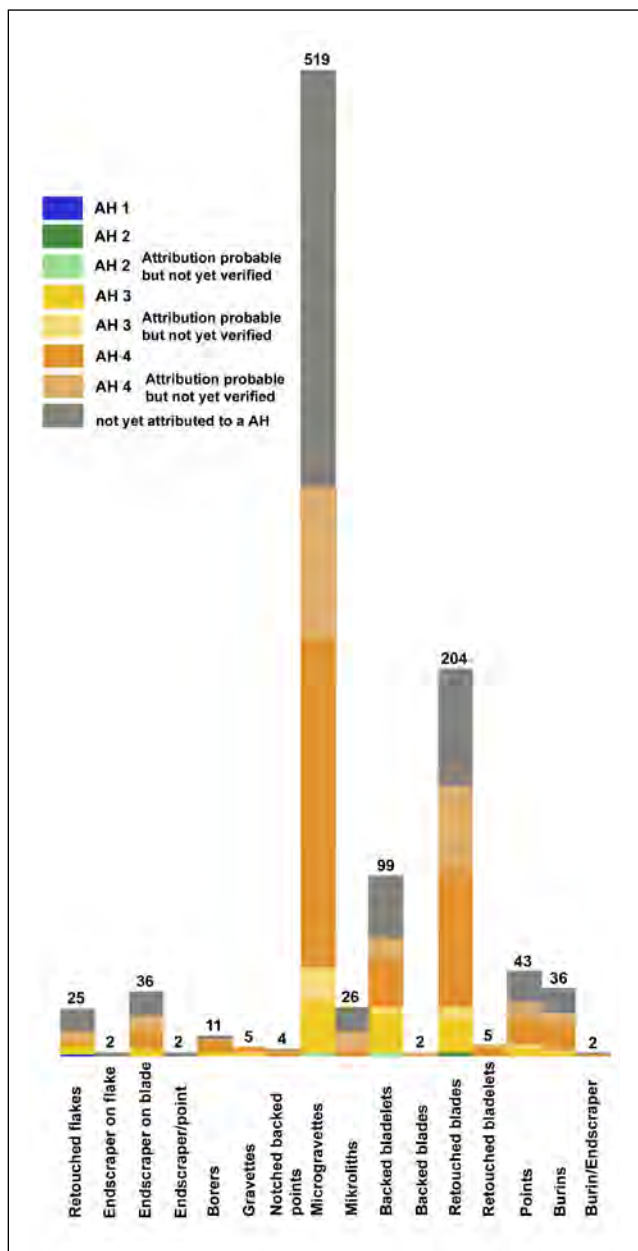


Fig. 6: Grub/Kranawetberg: Distribution of formal lithic tool types from AH1 to AH4.

Abb. 6: Verteilung der retuschierten Geräte auf die archäologischen Horizonte 1–4.

The radiocarbon dates as well as the material culture of AH4 suggest a position of the occupation at the end of the Pavlovian. Apart from the lithic raw material there are also some varieties of personal adornments which support the assumption of closer connections to the north in the course of AH4's occupation, but it must be mentioned that the role of the gravels of the river March as a source of Moravian chert used by the humans occupying Grub/Kranawetberg has yet to be analyzed.

As to the lithic material of AH3, the Carpathian Mountains in the east are a possible source for the radiolarites. The nearly exclusive use of radiolarites, resembling material from the Carpathian Mountains, may indicate increased relations to the East. Having a look at the personal adornments used in AH4 and AH3, we can observe a significant change, which supports the idea that different groups of people are

behind the formation of AH3 and AH4. Further analyses shall clarify the role of the site in a bigger regional context. It will be necessary to discuss to which extent the change in raw material procurement from AH4 to AH3 might correspond to a shift of population to the east at the end of the Pavlovian as has been suggested (OTTE, 1993; OTTE, 2004; ESCUTENAIRE et al., 1999). This shift of population as a consequence of the expansion of the Fennoscandinavian glacier and the climatic deterioration in central Europe is also being discussed by J. Svoboda (SVOBODA, 2000; SVOBODA, 2007) for the time after 25 ka BP.

Within the excavated area the occupations of AH2 and AH1 consist only of some scattered finds. There are some individual finds at this level in the western part of the site and a slight increase of find density in the eastern part. As we dispose of sufficient charcoal from these horizons we will be able to date these occupations in the near future. The analyses will show how near to the Last Glacial Maximum the cultural sequence from the Kranawetberg ended. The calibrated dates for the youngest sample in the profile of Zöllner (ZÖLLNER et al. in this volume) from O19 (Fig. 2) support this presumption.

First malacological analyses suggest a cold climate in an environment of some bushes and trees (ANTL-WEISER et al. 1997). According to F. Stadler there are indications for warmer climatic conditions shortly before the deposition of the AHs or the beginning of warmer conditions (ANTL-WEISER et al. 1997; ANTL et al. 2010). As warmer phases are observed at Pavlov and Willendorf II/8 (HAESAERTS et al. 1996; HAESAERTS et al. 2004) around 25.5ka BP we can expect that warmer conditions prevailed mainly before the deposition of the AHs at Grub/Kranawetberg.

In order to examine the role of climate change during the occupations after AH4 a series of samples for sediment, pollen and mollusc analyses have been collected. Apart from samples within the geological unit containing the archaeological horizons sampling comprised a length from 150 to 37 cm below surface. In 2010 and 2011 a series of samples for IRSL- and OSL-dates have been collected there. The upper parts from surface to -150 cm were sampled in various sequences in the west of the dwelling area as well as at the bone accumulation (ANTL et al. 1997, 1999). A drilling program executed by S. Verginis in 1996 in the west of the dwelling structures proved that there is at least a sequence of 7 m of loess at this place (ANTL et al. 1997, 1999).

## 5 Conclusion

Comparing AH4 and AH3 of Grub/Kranawetberg we could observe a clear change in settlement structures, raw material procurement and personal adornments as well as some differences concerning the frequency of certain tool types. According to the present state of research the assemblages of the two AHs seem to reflect the presence of different groups using this territory possibly under changing environmental conditions, one of them showing closer contacts to the north and the other to the east. The strongest argument for the preference of the existence of different groups seems to be the clear change in personal adornments within a short period of time. Chronologically the sequence of Grub/Kranawetberg overlaps with the sequence of Willen-

dorf II starting with AH4 and AH3 with radiocarbon dates between those of AHs 8 and 9 of Willendorf II (Haesaerts et al. 1996). AH2 and AH1 of Grub/Kranawetberg are younger than Willendorf II/AH9 and thus extend the cultural sequence of the Gravettian in Lower Austria.

Combined with a thick loess deposit below the AHs Grub/Kranawetberg gives an insight not only into an important part of the cultural development before the Last Glacial Maximum but also into climate changes during a longer time span of the Upper Pleniglacial period in this part of Austria.

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# Luminescence chronology of the Grub-Kranawetberg site, Austria

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**Abstract:**

Eight samples from a loess profile at the open air site of Grub-Kranawetberg (Lower Austria) were dated by Infrared Stimulated Luminescence (IRSL) of the polymineral fine-grain (4–11 µm) fraction and Optically Stimulated Luminescence (OSL) of quartz grains in the size range of 38–63 µm. Calibrated radiocarbon ages of ca. 30 ka BP on charcoal from the main archaeological layer AH4 agree with the IRSL and OSL age estimates within error limits. Anomalous fading of the feldspar-dominated IRSL from fine grain fraction could not be detected in the laboratory and the MAAD-IRSL ages were therefore not corrected. Most IRSL and OSL ages agreed within error limits, however, some IRSL ages were significantly underestimated as compared to the OSL ages, and in general the mean of the OSL ages was higher compared to the mean of individual IRSL ages. The OSL ages place the deposition of the exposed loess including an interstadial pedocomplex (“Stillfried B”) to between ca. 47 ka and ca. 30 ka. Potential pedostratigraphic correlations with some European loess areas as well as palaeoecological implications on Upper Palaeolithic human occupation patterns are discussed.

**Lumineszenz-Chronologie der Fundstelle Grub-Kranawetberg, Österreich****Kurzfassung:**

Das Alter von acht Proben aus einem Lössprofil der Freilandfundstelle Grub-Kranawetberg (Niederösterreich) wurde mittels der Lumineszenzmethoden der IRSL-Datierungen an der polymineralischen Feinkornfraktion und der OSL-Datierungen an der Mittelkornfraktion von Quarzen bestimmt. Kalibrierte Radiokohlenstoffalter von ca. 30 ka für den archäologischen Hauptfundhorizont AH4 werden innerhalb der Fehlergrenzen durch IRSL- und OSL-Alter bestätigt. Anomales Ausheilen der feldspat-dominierten Feinkornfraktion konnte im Labor nicht nachgewiesen werden, weshalb die MAAD-IRSL Alter nicht korrigiert wurden. Während die meisten IRSL- und OSL-Alter innerhalb ihrer Fehlergrenzen übereinstimmen, unterschätzen einige IRSL-Alter die OSL-Alter aus bisher unbekanntem Grund signifikant, wobei die Mittelwerte der OSL-Resultate tendenziell älter als die IRSL-Datierungen sind. Nach den OSL-Altern wurde der aufgeschlossene Löss einschließlich eines interstadialen Bodenkomplexes („Stillfried B”) im Zeitraum zwischen ca. 47 ka und ca. 30 ka gebildet. Mögliche pedostratigraphische Korrelationen mit einigen europäischen Lössgebieten sowie paläoökologische Überlegungen zur menschlichen Besiedlung im Jungpaläolithikum werden diskutiert.

**Keywords:**

*Luminescence dating, Grub-Kranawetberg, loess, palaeoecology, Stillfried B, correlations*

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## 1 Introduction

The Lower Gravettian open air site of Grub-Kranawetberg (Lower Austria) has been studied since the mid-1990. The site is described and discussed in detail in ANTL-WEISER (this volume). The age of the archaeological horizons, in particular of the main horizon AH4 attributed to the Gravettian period, is constrained by several radiocarbon ages. Further minor archaeological assemblages located a few cm above AH4 and labelled as “first upper AH” are followed by the “second upper AH” (ANTL-WEISER 2008). Another human occupation was detected 20 cm below AH 4 (Lower layer in Table 4). A section of ca. 210 cm thick brownish loess loam was observed from ca. 30 cm below AH4 and is interpreted as a pedocomplex, probably related to Stillfried-B. A more detailed pedological analysis is however pending. The timing of the deposition of the under- and overlying

loess, the occurrence of potential sedimentation hiatuses, and relative sedimentation rates were unknown so far. A contribution to these questions is attempted here by luminescence dating within the framework of an interdisciplinary team using a diachronic landscape reconstruction approach (WATERS 1996), which examined the entire profile of the 2010 and 2011 excavations. The present contribution tries to unravel the chronostratigraphy and geochronology of the site within the Danube loess sequence of Austria and relates these to palaeoecological interpretations.

## 2 Sampling

The Kranawetberg-site is situated west of the village Grub on the western side of the March River at N 48°25'08" and E 16°50'02", altitude 185 m a.s.l. Samples for optical dating (IRSL, OSL) were extracted from the excavation site (for lo-

cation see ANTL-WEISER 2008) by one of us (W.A.-W.) during the summer of 2010. Coordinates of the sampled profile O16 relative to the reference point of the excavation are  $x=15.07$  and  $y=15.86$  m. Steel cylinders of 100 cm<sup>3</sup> volume were hammered horizontally into the loess wall and sealed by light-tight lids. One of the main research focus was the verification of the radiocarbon dating of the main archaeological horizon (AH 4). Luminescence dating of such a horizon, that would have been mixed by the prehistoric anthropogenic activities, is considered as difficult. Thus, the sampling was performed at positions in 'clean' loess directly above and below this horizon. These were with the aim to provide bracketing ages. Further to understand the sedimentation history, additional samples were taken wherever feasible, in the entire available profile.

### 3 Sample preparation and IRSL/OSL measurements

Sample preparation for luminescence measurements were carried out in the dark laboratory under subdued red diode light ( $640\pm 20$  nm). Approximately 1 cm of material from each end of the cylinders was scraped off and used for dosimetry measurements using thick source alpha counting to determine U and Th contents (ZÖLLER & PERNICKA 1989) and ICP-MS (for K contents). Samples were then pre-treated with 1N HCl and 30% H<sub>2</sub>O<sub>2</sub>, followed by wet sieving and extraction of the 4–11 µm polymineral fine-grain fraction in Atterberg cylinders using Stoke's law. IRSL dating was carried out using the MAAD (Multiple Aliquot Additive Dose) protocol following MAUZ et al. (2002), with further experimental details given in FUCHS et al. (2008). IRSL measurements were carried out on a Risø TL/OSL-DA15 reader equipped with a Chroma D410 detection filter (transmission over  $425\pm 15$  nm band). The use of a MAAD protocol ensured that sensitivity changes were not an issue. All aliquots, including the ones for the fading experiment, were preheated at 220°C for 300 s before IRSL measurement. No normalization was required due to good reproducibility of the luminescence signals of the 5–6 aliquots for each of the additive dose points of 70, 140, 280 and 560 Gy. For data processing and calculation of equivalent doses by exponential fitting and extrapolation we used the Analyst (3.07b) software supplied with the Risø reader and followed MAUZ et al. (2002). All aliquots used for the construction of the MAAD growth curve (Figure 5) were stored at 70°C for one week between laboratory irradiation and IRSL measurements, to allow for the decay of short lived IRSL signals and also permit for the occurrence of any potential short term anomalous fading. Such a procedure in general equals storage of 4 weeks at room temperature (BERGER 1988, p. 26). A fading test was carried out by measuring the IRSL of 5 additional aliquots which had received the largest additive laboratory dose (ca. 560 Gy) immediately after irradiation and comparing this IRSL signal intensity with the signal from stored aliquots of the same dose group, which were used for the construction of the MAAD growth curve. The percentage of fading was calculated as the ratio of the intensities  $(N+\beta+\text{delay})/(N+\beta)$  of the IRSL-signal for each sample except BT 1009, which failed due to machine error.

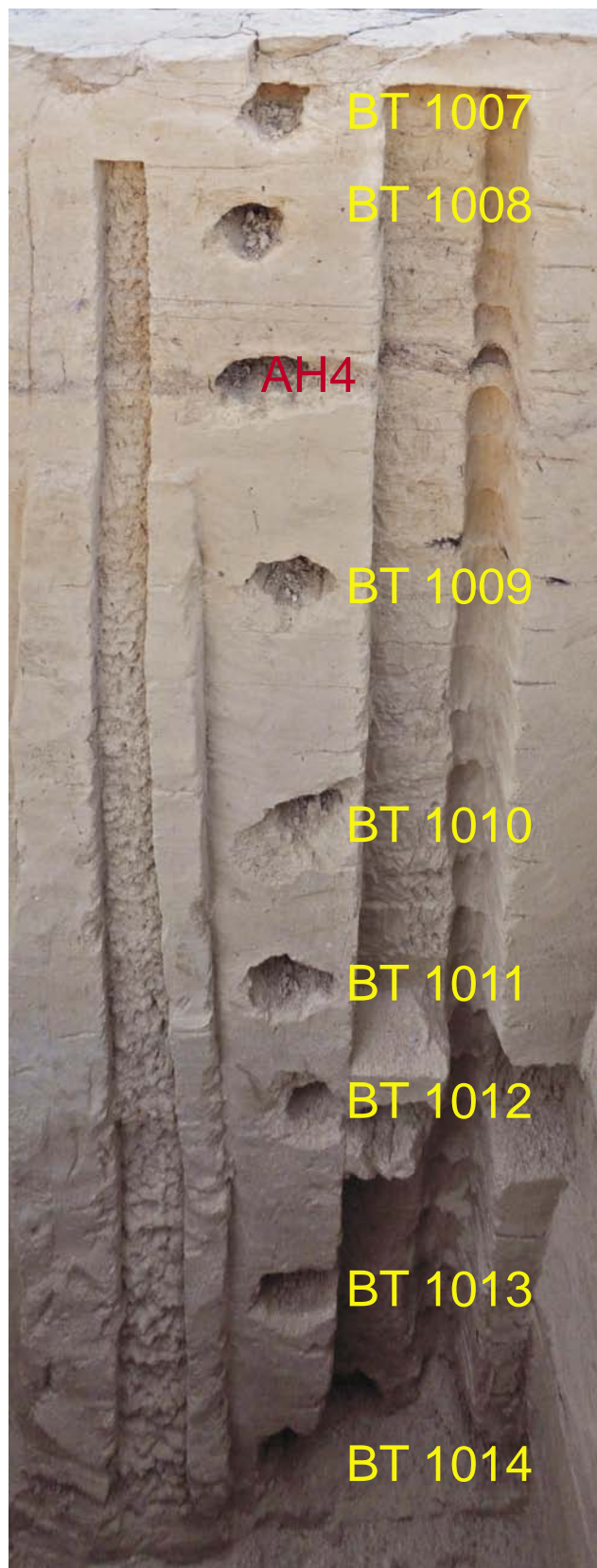


Fig. 1: Loess section of the Grub-Kranawetberg excavation 2010 and sample numbers. Note the reddish-yellow hue below and partly also above the main Archaeological Layer AH 4 and the increasing reddish-brown hue below sample BT 1012 indicating weak interstadial soil formation.

Abb. 1: Lössprofil von der Grabung Grub-Kranawetberg 2010 und Probennummern. Man beachte den rötlich-gelblichen Farbton unterhalb und teilweise auch oberhalb des Archäologischen Fundhorizontes AH 4 sowie den zunehmend rötlich-braunen Farbton unterhalb der Probe BT 1012 als Hinweise auf schwache interstadiale Bodenbildung. (Foto: W. Antl-Weiser)



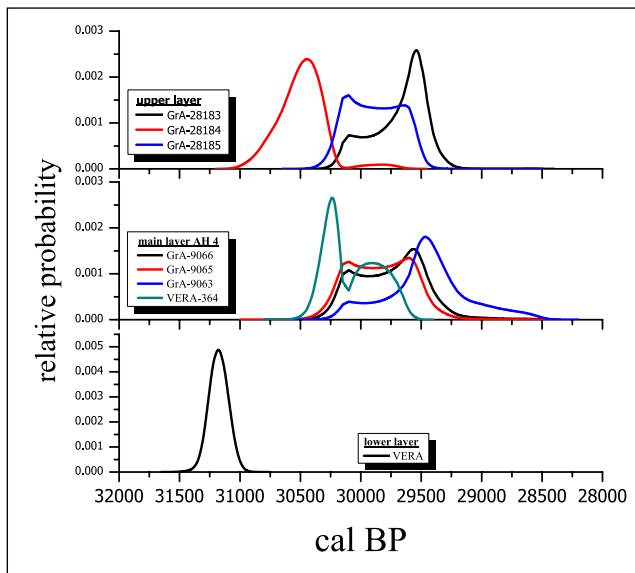


Fig. 2: Cumulative (black line, not normalized) and single (coloured) probability distributions of calibrated  $^{14}\text{C}$  ages from the Grub-Kranawetberg site (Calib 6.11 with IntCal09).

Abb. 2: Kumulative Wahrscheinlichkeitsverteilungen (schwarz, unnormiert) und einzelne Wahrscheinlichkeitsverteilungen (farbig) der kalibrierten  $^{14}\text{C}$ -Alter von der Fundstelle Grub-Kranawetberg (Calib 6.11 mit IntCal09).

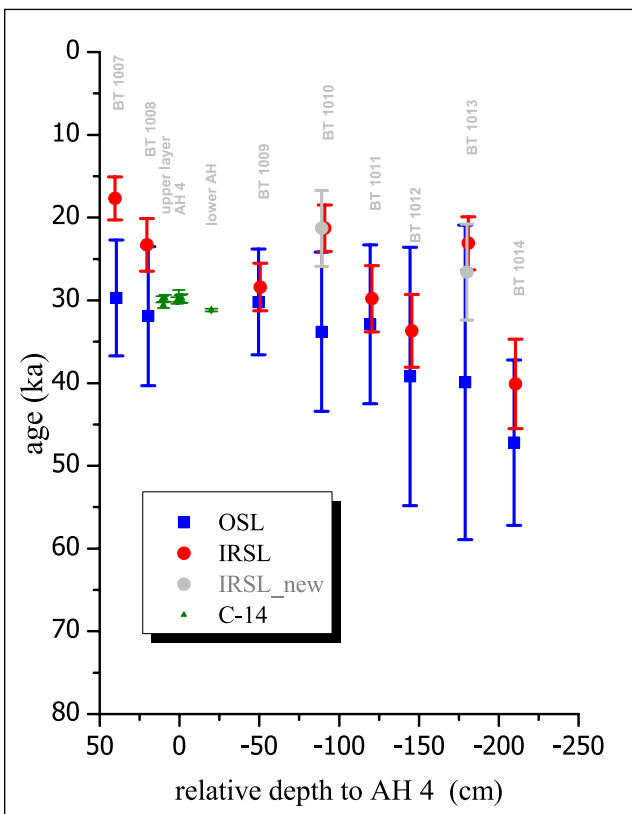


Fig. 3: Age-depth plot of  $^{14}\text{C}$  (calibrated), OSL and IRSL ages ( $2\sigma$ ). Sample depth positions were slightly shifted for clarity.

Abb. 3: Alters-Tiefenplot von  $^{14}\text{C}$  (kalibriert), OSL- und IRSL-Altern ( $2\sigma$ ). Probenpositionen wurden zur besseren Sichtbarkeit leicht versetzt dargestellt.

The quartz extracts from the middle grain size fraction (38–63  $\mu\text{m}$ ) were obtained by wet sieving and etching in hydrofluoric acid ( $\text{H}_2\text{SiF}_6$ ) following FUCHS et al. (2005) for OSL measurement by blue light stimulation. Feldspar contamination was ruled out by IRSL measurements yielding intensities at background level. OSL measurements followed the SAR protocol (MURRAY & WINTLE 2000) and experimental details as described in FUCHS et al (2012), including analytical details like acceptance criteria etc. A preheat of  $200^\circ\text{C}$  for 10 s was applied. Due to the scarcity of quartz middle grains size, the study was hampered by the low number of aliquots available for  $D_E$ -determination. Results from only 10–22 aliquots (excluding 0–10% rejection) could be obtained, and these resulted in relatively larger uncertainties and precluded elaborate statistical analysis of the  $D_E$  data.

## 4 Results

### 4.1 Dosimetry Results

Dose-rates were calculated assuming secular equilibrium of U decay series and using conversion factors given by GUERIN et al. (2011). The cosmic ray contribution was calculated after PRESCOTT & HUTTON (1994), using appropriate positional coordinates and approximate estimate of average sediment thickness. A moisture content of  $\delta=1.15\pm 0.05$  (see ZIMMERMAN 1971, LANG ET AL. 2003, ZÖLLER et al. 2013) was assumed for all samples with respect to hydromorphic features. This was despite the fact that the measured actual moisture content was lower with values between 1.09 and 1.11. An alpha efficiency factor (a-value) of  $0.08\pm 0.02$ , derived as an average from numerous data from loess (see e.g. LANG et al. 2003; LOMAX et al. 2013, take  $0.07\pm 0.02$ ) was assumed for all polymineral fine-grain samples. An a-value of  $0.035\pm 0.05$  was employed in dose-rate calculations for the quartz middle grain fraction (cf. LAI et al. 2008).

Analytical results for concentrations of radio-elements are given in Table 1 and the calculated dose-rate (in Gy/ka) for moist samples and the cosmic ray dose contributions are provided in Table 2 (data in all tables with  $1\sigma$  uncertainty).

### 4.2 IRSL dating results

All IRSL samples yielded shine-down plateaus suggesting sufficient bleaching of the IRSL signal during and prior to deposition of the loess. The fading tests resulted in ratios between  $1.04\pm 0.08$  and  $0.98\pm 0.06$ . Therefore, no correction for athermal fading was applied.

Equivalent doses together with effective dose-rates and the resulting IRSL ages are presented in Table 3 together with their  $1\sigma$  uncertainty. While IRSL results generally provide increasing ages with depth, the IRSL ages of samples BT 1010 and 1013 do not conform with this trend and were significantly younger than their respective overlying samples (Figure 3). The location on a plateau and the sample being buried under several meters of loess, rules out major sedimentological disturbances as a cause. According to the fading tests, short-term athermal fading can be precluded

as a possible reason.<sup>1</sup> The dosimetry is not different from the other samples and radioelement concentrations (Table 1) and thus do not indicate different sediment source areas, which could have resulted in the vastly different alpha sensitivity necessary for such age discrepancies. No explanation can be presently put forward, and therefore these age estimates were considered as outliers, and excluded from further interpretation, pending further research. Given such an ambiguity, OSL dating using pure quartz separates from the identical sample material was carried out in order to verify the reliability of IRSL dating.

### 4.3 OSL dating results

Effective dose-rates for middle grain quartz are lower than for feldspar-dominated fine-grains due to significantly lower  $\alpha$ -values. The equivalent doses together with effective dose-rates and resulting OSL ages are presented in Table 5 with  $1\sigma$  uncertainties. It is noteworthy that samples BT 1010 and 1013, which provided suspicious IRSL dating results, yielded OSL ages which are consistent with OSL ages for samples stratigraphically located above and below, respectively. Furthermore, no significant age inversion of OSL ages with respect to stratigraphy was observed, and from this point of view none of the OSL ages is therefore suspected to be underestimated.

### 5 Comparison of OSL and IRSL dating results

Radiocarbon ages for the main archaeological layer (ANTL-WEISER et al. 2010, NIGST & ANTL-WEISER 2012) were calibrated with the "Calib Rev 6.1.1" program (STUIVER & REIMER 1993), employing the IntCal09 calibration curve (REIMER et al. 2009). Calibration results are given in Table 4 and probability distributions are plotted in Figure 2. For the main archaeological layer AH4 we calculated a summed probability  $^{14}\text{C}$  age of 30405 to 29162 cal BP  $2\sigma$  (see Table 4).

An age-depth plot of IRSL, OSL and calibrated  $^{14}\text{C}$  ages is shown in Figure 3. Two IRSL ages from samples above AH4 of  $23.3 \pm 1.6$  ka (BT 1008) and  $17.7 \pm 1.3$  ka (BT 1007) are younger than the calibrated  $^{14}\text{C}$  ages, while IRSL results down to 150 cm below AH4 are still in accordance within uncertainties. The OSL ages are all in agreement with radiocarbon data, except for the lowermost sample (BT 1014), which is significantly older. The OSL ages are all consistent with the stratigraphy and samples BT 1008 and 1009, which are bracketing AH4, agree with the radiocarbon ages. In general, uncertainties for OSL ages (middle grains) are remarkably larger than for IRSL dating due to relatively large scatter of equivalent doses obtained from single aliquots. This is in accordance with previous observations (e.g., FUCHS et al 2012). In Figure 4, the OSL and the

accepted IRSL dating results with their  $1-\sigma$  errors are plotted together with the stratigraphy of the sampled profile. Whereas OSL and the accepted IRSL ages agree within the error limits for all samples below AH4, this is not the case for sample BT 1007 located stratigraphically above AH4.

As mentioned above, significant anomalous fading as a reason for the age underestimates of this IRSL age with respect to the OSL age is precluded. The fading experiment would have revealed at least short term fading as it equals

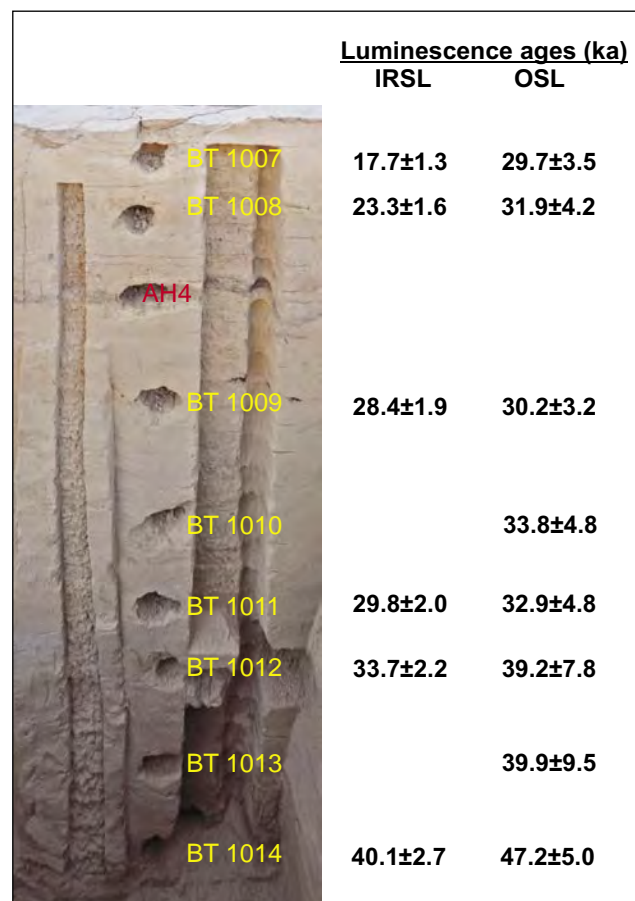


Fig. 4: Sample positions along the profile and comparison of OSL and IRSL dating results.

Abb. 4: Positionen der Proben im Profil und Vergleich von OSL- und IRSL-Datierungsergebnissen.

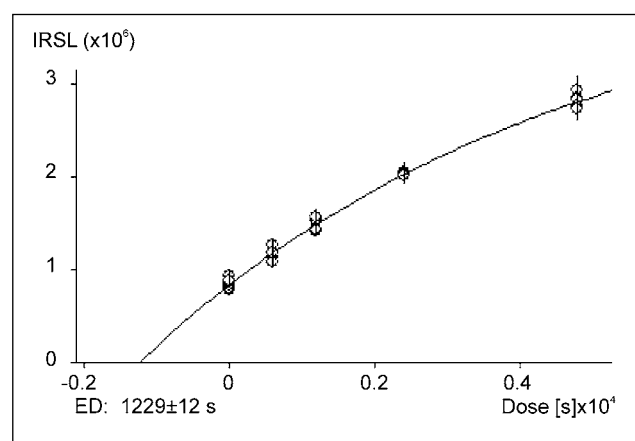


Fig. 5: Additive IRSL dose-response curve (MAAD) for sample BT 1012.

Abb. 4: Additive IRSL-Wachstumskurve (MAAD) für Probe BT 1012.

1) The measurements for these two samples were repeated in order to confirm the results. An identical equivalent dose (ED) was obtained for BT 1010\_new ( $76.8 \pm 3.6$  Gy) and, thus, the identical apparent age. In case of BT 1013 the ED changed significantly from  $91.9 \pm 1.3$  Gy to  $105.7 \pm 2.3$  Gy for reasons unknown so far (BT 1013\_new), whereas the apparent ages of  $23.1 \pm 1.6$  and  $26.6 \pm 2.9$  ka, respectively, agree within uncertainties. Nevertheless, the apparent IRSL age of BT 1010\_new remains a significantly underestimated outlier in an internal consistency check, where the identical systematic uncertainties are excluded.

Tab. 1: Radioelement concentrations.

Tab. 1: Konzentrationen der Radioelemente.

Sample no.	Th	$\delta$ Th	U	$\delta$ U	K	$\delta$ K
	[ppm]	[ppm]	[ppm]	[ppm]	[%]	[%]
BT 1007	11.49	1.44	3.73	0.44	1.54	0.15
BT 1008	9.12	1.31	4.10	0.40	1.59	0.16
BT 1009	10.22	1.11	4.19	0.34	1.58	0.16
BT 1010	7.63	0.97	4.15	0.30	1.65	0.17
BT 1011	9.27	0.95	3.67	0.29	1.76	0.18
BT 1012	11.11	1.06	4.12	0.32	1.82	0.18
BT 1013	11.01	1.14	3.49	0.35	1.75	0.18
BT 1014	9.35	1.14	3.32	0.35	1.78	0.18

Tab. 2: Dose-rates for moist samples assuming  $\delta=1.15$

Tab. 2: Dosisleistungen für feuchte Proben bei  $\delta=1.15$

Sample no.	$D_{\alpha \text{ moist}}$	$\delta(D_{\alpha \text{ moist}})$	$D_{\beta \text{ moist}}$	$\delta(D_{\beta \text{ moist}})$	$D_{\gamma \text{ moist}}$	$\delta(D_{\gamma \text{ moist}})$	Depth	$D_{\text{cosm}}$	$\delta(D_{\text{cosm}})$
	[Gy/ka]	[Gy/ka]	[Gy/ka]	[Gy/ka]	[Gy/ka]	[Gy/ka]	[m]	[Gy/ka]	[Gy/ka]
BT 1007	9.96	0.66	1.78	0.15	1.15	0.10	1.5	0.195	0.02
BT 1008	8.09	0.55	1.81	0.15	1.10	0.09	1.7	0.193	0.02
BT 1009	9.00	0.61	1.84	0.15	1.15	0.09	2.4	0.186	0.02
BT 1010	6.88	0.48	1.82	0.15	1.06	0.08	2.8	0.182	0.02
BT 1011	8.15	0.55	1.88	0.16	1.10	0.08	3.1	0.179	0.02
BT 1012	9.71	0.65	2.02	0.17	1.23	0.09	3.4	0.176	0.02
BT 1013	9.54	0.63	1.89	0.16	1.16	0.09	3.7	0.173	0.02
BT 1014	8.16	0.55	1.85	0.16	1.08	0.09	4.0	0.170	0.02

Tab. 3: Equivalent doses (ED), effective dose-rates (Gy/ka) and IRSL ages (ka). Ages in italics are considered as outliers and discarded from further discussion.

Tab. 3: Äquivalenzdosen (ED), effektive Dosisleistungen und IRSL-Alter (ka). Alter in Kursivschrift werden als Ausreißer betrachtet und in der weiteren Diskussion nicht berücksichtigt.

Sample no.	ED	$\delta$ ED	$D_{\text{eff}}$	$\delta(D_{\text{eff}})$	Age	$\delta$ Age
	[Gy]	[Gy]	[Gy/ka]	[Gy/ka]	[ka]	[ka]
BT 1007	69.4	2.1	3.93	0.28	<b>17.7</b>	<b>1.3</b>
BT 1008	87.5	2.2	3.75	0.25	<b>23.3</b>	<b>1.6</b>
BT 1009	110.7	1.5	3.90	0.26	<b>28.4</b>	<b>1.9</b>
BT 1010	76.8	1.5	3.61	0.22	<i>21.3</i>	<i>1.4</i>
_new	76.8	3.6			<i>21.3</i>	<i>2.3</i>
BT 1011	113.5	1.7	3.81	0.25	<b>29.8</b>	<b>2.0</b>
BT 1012	141.6	1.4	4.20	0.28	<b>33.7</b>	<b>2.2</b>
BT 1013	91.9	1.3	3.98	0.27	<i>23.1</i>	<i>1.6</i>
_new					<i>26.6</i>	<i>2.9</i>
BT 1014	150.4	1.7	3.75	0.25	<b>40.1</b>	<b>2.7</b>

4 weeks of room temperature storage. We cannot preclude, however, mid- or long-term fading (XIE & AITKEN 1991) which cannot be detected in the laboratory within the available time. Inspection of the shapes of luminescence decay and growth curves (Figure 5) did not reveal any clues, where especially the latter are well below the onset of saturation and therefore do not appear to be critical. The same applies to possible discrepancies in alpha sensitivity. Provided the uniformity of the source material, which is indicated by the lack of variability in radioelement concentration, a significant discrepancy in the alpha sensitivity is considered most unlikely as reason. We like to mention, however, that during sampling in the field we observed crotonas extending downwards almost to the upper archaeological layer and causing some problems during sampling. Furthermore, the area was used as a vineyard in previous times and remnants of deep roots of grapevines were detected in the uppermost loess. Even if sampling tried to steer clear of crotonas and tracks of former grapevines we cannot be totally certain to have entirely avoided sampling material affected by bioturbation. Thus, the obtained age, possibly including the other sample located above AH4 (BT 1008) may be a mixed age and thus underestimating the original deposition age of the loess. It would be a surprise, however, if only fine grains contaminated the loess due to bioturbation and not coarse silt grains used for OSL dating. This question will be studied further using soil micromorphology during ongoing research by other members of the working group.

The IRSL ages from above AH4 may suggest erosional hiatuses or strongly reduced sedimentation rate. Although traces of solifluction were detected in this part of the section, other possible reasons for these rather young apparent ages need to be identified. As far as AH4 and deeper horizons are concerned, we can state that within uncertainties IRSL and OSL ages are consistent with calibrated radiocarbon ages on charcoal which yielded congruent results from different laboratories (ANTL-WEISER et al. 2010, NIGST & ANTL-WEISER 2012) and an age of ca. 30 ka for AH4 is confirmed by luminescence dating results. The chronostratigraphic position of the loess above the upper archaeological layer so far remains a matter of debate.

## 6 Discussion

The luminescence dating results for the Grub-Kranawetberg site have implications for the regional loess stratigraphy and the supposed relation of human occupation patterns to specific palaeoecologies. If the uppermost OSL ages correctly reflect the deposition ages of the loess above AH4 the entire exposed loess sequence presumably accumulated during a time period corresponding to a late phase of Marine Isotopic Stage MIS 3. The two uppermost IRSL ages, however, argue for deposition of the loess above the upper archaeological layer during MIS 2.

### Constraints for palaeoecology

The archaeological layer AH4 overlies a brownish interstadial palaeosol which appears to be coeval with the nearby "Stillfried B" palaeosol at its type locality. The loess in which AH4 is embedded exhibits, however, features of hydromor-

phic soils and of solifluction (as so far no indication of harsh permafrost conditions was observed in this layer we were cautious in using the term gelifluction). These observations as well as preliminary malacological results (ANTL-WEISER et al. 1997) indicate a climatic deterioration (cooling) after the thermal optimum of the interstadial for the time of occupation. The presence of humans in such palaeoecological conditions is usually neglected, based on an oversimplifying geodeterministic view that in general Upper Palaeolithic humans inhabited Central Europe at the onset or during thermal optimum of interstadials and withdrew at the onset of climatic cooling. Instead, moister and cooler climate at the onset of a stadial could have triggered higher bioproductivity of the landscape with respect to a summer-dry loess steppe and, thus, better conditions for game (e.g., ZECH & HINTERMAIER-ERHARD 2002, p. 12; SCHULTZ 2002). MÖLLER & THANNHEISER (2011) point out that tundra soils of moist habitat can produce a high amount of below-ground phytomass "which is important for plant growth and for building up soil organic matter" (p. 255).

In many classical loess-palaeosol sequences of Central Europe the time span between ca. 30 and ca. 26 ka is poorly resolved due to erosion (e.g., ZÖLLER & SEMMEL 2001, ZÖLLER et al. 2004). Convincing evidence for rapid loess sedimentation starting around 30 ka ago (i.e., late MIS 3) with intercalation of up to 4 or 5 tundra gley soils was reported from the Nussloch site, Germany (ANTOINE et al. 2009) and from the Dolní Věstonice (with less thickness) in, Czech Republic (FUCHS et al. 2012), as also from the Krems-Wachtberg site, Austria (LOMAX et al. 2013, ZÖLLER et al. 2013). In the upper part of the relevant section at Nussloch, ANTOINE et al. (2009) observed "cryo-desiccation micro cracks" (layer 24, see legend to their Figure 2). Even if Nussloch is situated ca. 630 km WNW of Grub the drastic climatic cooling starting around 30 ka ago must have affected the surroundings of Grub and Stillfried as well. LÜTHGENS & BÖSE (2011) suggest that the ice advance of the "Brandenburg phase" (local glacial maximum) of the last glaciation (Weichselian) occurred during the time interval under question. This dramatic palaeoecological event in northern Central Europe may be reflected some Central European loess areas.

### Constraints for Austrian and European loess stratigraphy

The chronostratigraphic position of the "Stillfried B" palaeosol has been a matter of debate (TERHORST et al. 2011). At the "classical" site in the north-western corner of the former Stillfried brickyard (N 48° 24' 32,4", E 16° 50' 31,6") the substratum of the weakly developed "Stillfried B" soil was dated 29.8±3.7 ka by TL (ZÖLLER et al. 1994). It directly overlies Rissian loess (HAMBACH & ZÖLLER, unpublished results) from which it is separated by an erosional hiatus. For this reason, the "Stillfried B" pedocomplex is supposed to have developed only with its youngest part at this site. PETICZKA et al. (2010) described a "revised profile" from the western part of the abandoned brickyard which appears stratigraphically more complete and resembles the stratigraphy of the Grub-Kranawetberg excavation. Using field methods mentioned by PETICZKA et al. (2010) the authors distinguish 3 fossil BC horizons at their "Stillfried B 2005" site which corresponds to 3 buried interstadial soils at Grub-Kranawetberg. Unfor-

Tab. 4: Uncalibrated and calibrated ( $2\text{-}\sigma$ ) radiocarbon ages (Calib Rev 6.1.1 & IntCal09).

\* no laboratory number available

Tab. 4: Unkalibrierte und kalibrierte ( $2\text{-}\sigma$ ) Radiokohlenstoff-Alter (Calib Rev 6.1.1 & IntCal09).

\* keine Labornummer verfügbar

Archaeological	relative	Laboratory	uncalibrated	error	calibrated
layer	depth	code	age [aBP]	[a]	age range [a calBP]
Upper layer	10	GrA-28183	24780	140	30172 - 29346
	10	GrA-28184	25640	160	30939 - 30191
	10	GrA-28185	25010	150	30254 - 29501
Main layer AH4	0	GrA-9066	24830	230	30284 - 29257
	0	GrA-9065	24930	240	30321 - 29359
	0	GrA-9063	24620	230	30189 - 28748
	0	VERA-364	25300	90	30409 - 29645
Lower layer	-20	VERA*	26700	120	31336 - 31013

Tab. 5: Equivalent doses (ED), effective dose-rates (Gy/ka) and quartz OSL ages (ka)

Tab. 5: Äquivalenzdosen (ED), effektive Dosisleistungen und Quarz-OSL-Alter (ka)

Sample no.	ED	$\delta_{ED}$	$D_{eff}$	Age	$\delta_{Age}$
	[Gy]	[Gy]	[Gy/ka]	[ka]	[ka]
BT 1007	101	9	3.40	<b>29.7</b>	<b>3.5</b>
BT 1008	106	11	3.31	<b>31.9</b>	<b>4.2</b>
BT 1009	103	7	3.41	<b>30.2</b>	<b>3.2</b>
BT 1010	108	13	3.23	<b>33.8</b>	<b>4.8</b>
BT 1011	111	14	3.37	<b>32.9</b>	<b>4.8</b>
BT 1012	144	28	3.68	<b>39.2</b>	<b>7.8</b>
BT 1013	138	27	3.47	<b>39.9</b>	<b>9.5</b>
BT 1014	156	16	3.31	<b>47.2</b>	<b>5.0</b>

tunately, no dating results have been published so far from the “Stillfried B 2005” site. The time of this pedocomplex at the Grub-Kranawetberg excavation is bracketed by our IRSL ages between  $28.4\pm 1.9$  ka and  $40.1\pm 2.7$  ka and by our OSL ages between  $30.2\pm 3.2$  ka and  $47.2\pm 6.0$  ka. Thus, the Grub-Kranawetberg site and the “Stillfried B 2005” site both appear well-suited to complete the Austrian loess stratigraphy in the period between ca. 40 and ca. 30 ka (MIS 3, upper part) by increasing the number of observed soils from one (Stillfried B) to a complex of three paleosols. Furthermore, our new dating results can inform the discussion on the correlation of the Stillfried B palaeosol with loess stratigraphies from other European areas, such as Southern Moravia (FUCHS et al. 2012), Serbia (FUCHS et al. 2008), Hungary (ZÖLLER et al. 1994, THIEL et al. 2013), Austria (THIEL et al. 2011), Ukraine (ROUSSEAU et al. 2011), and Germany (ANTOINE et al. 2009, KADEREIT et al. 2013). The pedo-complex at Grub-Kranawetberg below AH4 and the revised “Stillfried B 2005” section appear to be coeval with interstadial soils or pedocomplexes in the mentioned areas. But with respect to palaeo-moisture conditions, palaeo-temperatures and sedimentation rates in

the different regions the palaeosols manifest different soil typologies ranging from tundra-gleys or (sub-) arctic cambisols to chernozems. It may prove of particular interest that the range of apparent OSL ages bracketing the lowermost soil of the Grub-Kranawetberg pedocomplex (samples BT 1013 and 1014) includes the TL age of  $43\pm 4$  ka for the substratum of the “MF1” palaeosol (chernozem) in Hungary (ZÖLLER et al. 1994, cf THIEL et al. 2013). On the other hand, high sedimentation rates and locally dry climate impeded well-distinguishable interstadial palaeosols at the Krems-Wachtberg site. Therefore, an equivalent of the Stillfried B pedocomplex cannot be clearly identified so far at Krems-Wachtberg, although the same age range is covered (TERHORST et al. 2013).

## 7 Conclusions

The application of an IRSL-MAAD protocol for polymineral fine grain loess has once more (as in e.g. LANG et al. 2003; FUCHS et al. 2008) yielded age estimates congruent in their majority with OSL and radiocarbon dating. While this in itself indicates the absence of anomalous fading of

the feldspar luminescence component used, a fading test neither did show any loss of signal, and it appears that the occurrence of feldspar fading might be related to detection (here: blue band wavelength) and protocol (here: MAAD) used. OSL ages from quartz middle grains are, however, stratigraphically consistent without exception and also agree with radiocarbon dating results.

Luminescence dating confirms an age of ca. 30 ka for the main archaeological horizon AH4. According to proxy data, the main human occupation documented in level AH4 occurred after the thermal optimum of an interstadial at the onset of colder and moister conditions. A threefold complex of interstadial soils underlies the archaeological layers of Lower Gravettian age. This pedocomplex developed in a time span ranging from  $\geq 40$  ka to ca. 30 ka and includes the palaeosol known as “Stillfried B” soil. The pedocomplex at the Grub-Kranawetberg site appears, however, coeval with a nearby pedocomplex introduced as “Stillfried B 2005” (PETICZKA et al. 2010) and with interstadial palaeosols or pedocomplexes of different typologies in various European loess regions. The Grub-Kranawetberg site is, thus, well-suited to complete the Austrian loess stratigraphy for the upper part of MIS 3 and to refine correlation with neighbouring European loess areas. For the loess overlying the archaeological layers at the Grub-Kranawetberg site a significant difference between IRSL and OSL ages still exists which calls for further investigations, e.g. the application of the NCF protocol for quartz OSL (SINGHVI et al. 2011).

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# Impact of atmospheric warming on permafrost degradation and debris flow initiation – a case study from the eastern European Alps

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## Abstract:

The present study demonstrates the importance of recent atmospheric warming for the spatial distribution of debris flow initiation in a central alpine area of the Eastern Alps. In particular, permafrost degradation due to increasing mean annual air temperature (MAAT) since the end of the Little Ice Age (LIA) caused mechanical instabilities of sediments and slopes. In the study area, the Rieserferner-Ahrn Nature Park, it can be shown that almost half of the debris flow initiation zones originate in areas with loose rock that were still stabilized by glacier ice and/or permafrost about 150 years ago.

At present, the permafrost area covers 109 km<sup>2</sup> in the study area, while it covered ~211 km<sup>2</sup> during thermal conditions at the end of the LIA. Since then glacier recession and permafrost degradation exposed extended areas of unconsolidated debris that are uniformly distributed in the Rieserferner-Ahrn Nature Park. In general, present-day debris flows mainly originate from corresponding sediments, where permafrost degraded since the end of the LIA. These sediments are now susceptible to debris flow hazards.

Compared to present conditions the permafrost area would decrease by approximately 72 % by the middle of the 21st century with regard to an increased air temperature of +1 to +2 K. Moreover, glaciers widely disappear in this scenario. Ongoing glacier recession and permafrost degradation increase the amount of instable debris as well as the potential of debris flow detachment zones in the future.

## Der Einfluss der atmosphärischen Erwärmung auf Permafrostdegradation und das Anreißen von Muren – eine Fallstudie aus den Ostalpen

## Kurzfassung:

In der vorliegenden Arbeit wird die Bedeutung der Erwärmung der Kryosphäre seit dem Ende der Kleinen Eiszeit (LIA) für die räumliche Verbreitung von Muranrissen in einem zentralalpiner Gebiet der Ostalpen untersucht. Vor dem Hintergrund der atmosphärischen Erwärmung verursachte hier insbesondere die Degradation von Permafrost bodenmechanische Instabilitäten. Im Untersuchungsgebiet, dem Naturpark Rieserferner-Ahrn in Südtirol lässt sich zeigen, dass mehr als die Hälfte der Muranrisse in Lockergesteinen auftreten, unter anderem in Moränen- und Hangschuttablagerungen, die vor rund 150 Jahren noch durch Gletschereis und Permafrost stabilisiert waren.

Derzeit sind rund 109 km<sup>2</sup> des Untersuchungsgebietes von Permafrost unterlagert, während unter den thermischen Bedingungen der Kleinen Eiszeit diese Fläche noch rund 211 km<sup>2</sup> umfasste. Die seither von Gletscherschwund und/oder Permafrostdegradation betroffenen Areale mit schlecht oder nicht konsolidierten Ablagerungen sind über das untersuchte Gebiet weitgehend gleichmäßig verteilt. Ein Großteil der erfassten Muranrisse tritt in entsprechenden und nun für gefährliche Prozesse disponierten Sedimenten auf.

Unter Berücksichtigung aktueller Klimaszenarien mit einem Anstieg der Lufttemperaturen um +1 bis +2 K bis zur Mitte des 21. Jahrhunderts würde sich das von Permafrost unterlagerte Areal im Untersuchungsgebiet um annähernd 72% gegenüber aktuellen Bedingungen verringern. Gleichzeitig würden die Gletscherflächen weitgehend verschwinden. In Zukunft ist durch anhaltenden Gletscherschwund und Permafrostdegradation daher mit einer Zunahme instabiler Schuttvorkommen und einer räumlichen Zunahme von Muranrissen und Murenprozessen zu rechnen.

## Keywords:

*atmospheric warming, climate change, glacier recession, permafrost degradation, natural hazards, debris flow initiation, Little Ice Age, European Alps*

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## 1 Introduction

Glacier ice and permafrost ice react sensitive to climate variability in the European Alps (cf. BENISTON et al. 1997, DAVIES et al. 2001, ARENSON 2003, DAMM & LANGER 2006, HARRIS et al. 2009). Since the end of the “Little Ice Age” (LIA), about 150 years ago, the temperature increase especially af-

ected the mountain cryosphere (e.g. LEWIN & WARBURTON 1994, HARRIS et al. 2003, STÖTTER et al. 2003, HAEBERLI & GRUBER 2009). High-mountain environments are presumed to response to atmospheric warming with glacier recession, increased permafrost temperatures and increased slope instabilities (STOFFEL & HUGGEL 2012). In turn, these changes affect present and future natural hazards (cf. HAEBERLI et



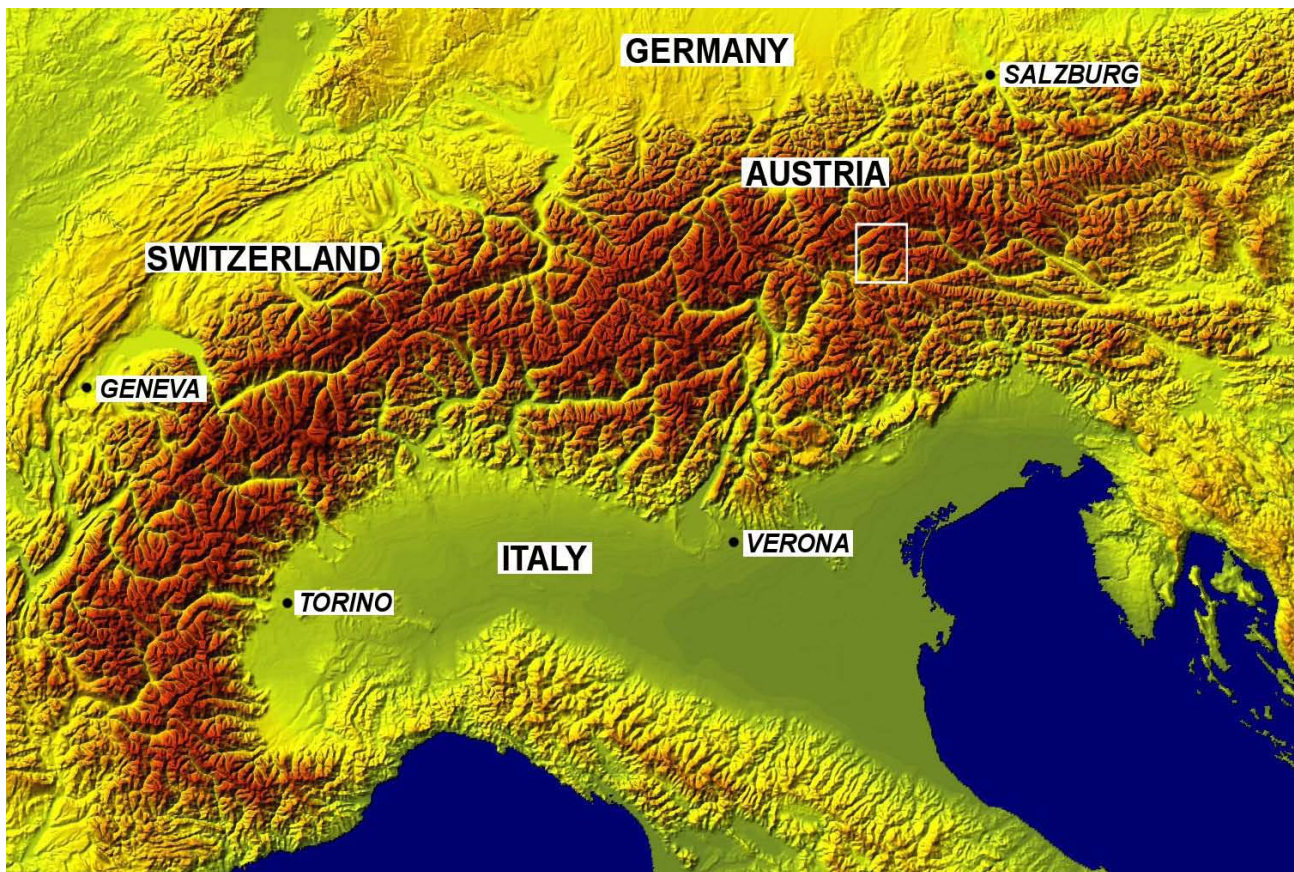


Fig. 1: European Alps with the location of the study area Rieserferner-Ahrn Nature Park (box) and the adjacent areas in South Tyrol (Italy).

Abb. 1: Der Alpenraum mit der Lage des Untersuchungsgebietes Naturpark Rieserferner-Ahrn (Kasten) und angrenzenden Gebieten in Südtirol (Italien).

al. 1999, DAVIES et al. 2001, HARRIS et al., 2001, GRUBER et al. 2004, HARRIS 2005, FISCHER et al. 2006, DAMM & FELDERER 2008, PRÖBSTL & DAMM 2009, KEILER et al. 2010).

The glacier retreat since the end of the LIA is well documented in the European Alps (e.g. DAMM 1998, LAMBRECHT & KUHN 2007, ZEMP et al. 2007, WINKLER 2009, WINKLER et al. 2010). In contrast, knowledge on the occurrence and distribution of permafrost and its spatiotemporal change during the last 150 years is comparatively scarce (cf. VONDER MUEHLL et al. 2007, MAIR et al. 2008). Early investigations estimated an increase of the mean annual air temperature (MAAT) of about +1 to +2 K since the end of the LIA in permafrost areas of the European Alps (cf. KUHN 1990, PATZELT & AELLEN 1990, HAEBERLI 1992). Thus, based on a thermal gradient of 0.5–0.6 K/100 m, the lower limit of permafrost increased by 150–300 m during the last 150 years (e.g. DAMM & LANGER 2006, see also HAEBERLI 1999). Similarly, the prognosticated increase of air temperature of +1 to +2 K in the Eastern Alps by the middle of the 21<sup>st</sup> century (cf. MATULLA et al. 2002, MATULLA 2005) may lead to an additional rise of equilibrium line altitudes (ELA) of alpine glaciers of approximately 150–350 m and a further increase of the lower limits of permafrost of 200–400 m. According to this scenario, numerous glaciers within the eastern Alps would disappear and permafrost diminish to the highest summits. Both processes would affect the stability of mountain slopes and walls (e.g. DAVIES et al. 2001, ZEMP et al. 2006, 2007).

The thawing of mountain permafrost is expected to significantly affect geotechnical properties of perennially frozen and unconsolidated debris and may result in a regional

increase of debris flow frequency and magnitude (cf. ZIMMERMANN et al. 1997, HARRIS 2005, STOFFEL & HUGGEL 2012). Both, the complete melting of the permafrost body and the lowering of the thaw front within permafrost may reduce the shear strength of debris. Instabilities appear most likely to occur in areas near the lower limit of contemporary permafrost distribution, where permafrost bodies are thin and have temperatures close to 0°C (HAEBERLI 1992). Successive failure of the unconsolidated debris destabilized in such a manner can be triggered by the oversaturation of the active layer or debris above water-impermeable rock (e.g. DAMM et al. 2012, SANDMEIER et al. 2012) or by the release of fine grained material formerly fixed in the ice matrix (SATTLER et al. 2011). However, the mechanisms of permafrost degradation and related slope stability are rather complex and many aspects and links remain uncertain to date (STOFFEL & HUGGEL 2012).

The theoretical relation between permafrost degradation in non-creeping slope material and the initiation or increase of debris flow activity has hardly been verified. Field studies of debris flow-triggering mechanisms on thawing slopes remain a challenge given the difficulties to predict where and when slope instabilities occurs (cf. HARRIS 2005). Still, there are some observations of large numbers of debris flows originating in areas presumed to be at the margin of contemporary permafrost distribution supporting the hypothesis mentioned above (e.g. ZIMMERMANN & HAEBERLI 1992, STÖTTER et al. 2003, SATTLER et al. 2011). However, these studies are based on comparatively small study areas and/or short time-frames.

The present study aims to identify and to prove the rela-

tion among atmospheric warming, the spatial variability of glaciers and permafrost and the spatial distribution of debris flows in the study area during the last 150 years. Furthermore, the study aims to predict future spatial development of glacier and permafrost distribution in relation to debris flows as hazardous geomorphic processes using a GIS-based modelling approach. Climatic basis of the scenario is an expected increase of air temperature by +1.5 K until the middle of the 21<sup>st</sup> century (cf. MATULLA et al. 2002, MATULLA 2005, IPCC 2007b). Thus, for the first time this study provides data of the development of hazard potential from debris flow processes in the eastern Alps that exceed the spatial dimension of a catchment area (cf. STÖTTER et al. 2003, KNEISEL et al. 2007, DAMM & FELDERER 2008).

## 2 Regional setting

The study area (fig. 1) extends to 450 km<sup>2</sup> and covers the Rieserferner-Ahrn Nature Park in South Tyrol (Italy) and adjacent areas of the Hohe Tauern National Park (Austria). It comprises the Rieserferner and Durreck Mountains, parts of the Zillertaler Alps south flank, and the outer west parts of the Venediger Mountains. Highest peaks in the study area reach up to 3,500 m a.s.l.

The central parts of the study area are dominated by granodiorite and tonalite bedrock. Additionally, pegmatite, paragneiss, mica schist, and orthogneiss are present. Large areas in the upper valleys are covered by poorly consolidated or unconsolidated sediments, particularly debris cones, talus, till and slide masses (SANDMEIER et al. 2012). These sediments of Late Glacial and Holocene age can reach

thickness of several decametres (DAMM 1996). In general, most of the material is not or only sparsely vegetated. Erosion and debris flow processes can be frequently observed.

In general, the study area is characterized by comparably continental climate. Mean annual precipitation varies throughout the study area and is highest in the northern part of the Rieserferner-Ahrn Nature Park with up to 1030 mm (meteorological station of Prettau, period of 1981–2000), as this region is close to the central Alpine divide. Main precipitation period is the summer with frequent convective rainfall.

The thermal gradient shows significant seasonal variations and ranges from 0.7° C/100 m (summer) to 0.35° C/100 m (winter) based on data of nine meteorological stations between 821 m a.s.l. and 3105 m a.s.l. for the period 1994–2004 (tab. 1, cf. DAMM & LANGER 2006). Using a mean thermal gradient of 0.57 K/100 m (tab. 2), the calculated 0°C-isotherm is at 2323 m a.s.l.

Since the end of the Little Ice Age about AD 1860 the glacier area decreased by 50 to 65% in the Rieserferner-Ahrn Nature Park. The loss of glacier surface was marginally above-average compared to surrounding mountain groups of the Eastern Alps (cf. DAMM 1998, RÖSEN 2005). Strong glacier recession was recorded during the last three decades when the ELA partially rose up to 3100 m a.s.l. within the study area. Significant mass losses occurred during this period, which were related to the decrease of the Accumulation Area Ratios (AAR), for example at the Western Rieser glacier (cf. SECCHIERI & VALENTINI 1985, AUTONOME PROVINZ BOZEN 2010, 2011).

Meteorological station	Coordinate	Altitude [m a.s.l.]
Bruneck	11° 55' 53" E / 46° 48' 16" N	821
Mühlen	11° 56' 25" E / 46° 54' 06" N	870
Steinhaus	11° 58' 34" E / 46° 59' 46" N	1080
Antholz - Mittertal	12° 06' 55" E / 46° 51' 30" N	1236
Antholz - Obertal	12° 06' 55" E / 46° 52' 14" N	1320
Prettau	12° 05' 51" E / 47° 02' 13" N	1449
Rein in Taufers	12° 04' 34" E / 46° 56' 46" N	1600
Prettau - Merbalm	12° 07' 21" E / 47° 02' 11" N	2002
Prettau - Lengspitze	12° 07' 52" E / 47° 00' 57" N	3105

Tab. 1: Position and altitude of meteorological stations used for the calculation of thermal gradients of monthly and annual air temperature in the Rieserferner-Ahrn Nature Park and adjacent areas (source: Autonomous Province of Bozen-South Tyrol – Hydrographic Office).

Tab. 1: Lage und Höhe von Wetterstationen zur Berechnung thermischer Gradienten der monatlichen und jährlichen Lufttemperaturen im Naturpark Rieserferner-Ahrn und angrenzenden Gebieten (Quelle: Autonome Provinz Bozen-Südtirol – Hydrographisches Amt).

Tab. 2: Thermal gradients of monthly and annual air temperature and 0°C-isotherm based on data of nine meteorological stations in the Rieserferner-Ahrn Nature Park and adjacent areas for the period 1994–2004 according to DAMM & LANGER (2006).

Tab. 2: Thermische Gradienten der monatlichen und jährlichen Lufttemperatur sowie 0°C-Isotherme auf der Grundlage von neun Wetterstationen im Naturpark Rieserferner-Ahrn und angrenzenden Gebieten für den Zeitraum 1994–2004 nach DAMM & LANGER (2006).

Month	Jan	Feb	Mar	Apr	May	Jun	Jul	Aug	Sep	Oct	Nov	Dec	Year
Thermal gradient [K/100 m]	0,36	0,51	0,64	0,70	0,70	0,68	0,70	0,62	0,59	0,53	0,48	0,35	0,57
0°C - Isotherm [m a.s.l.]	-34	917	1633	2009	2801	3356	3473	3857	3072	2536	1439	280	2323

### 3 Methods and data sets

#### Ground survey

Field work comprises the identification, mapping and assessment of geomorphological, hydrological and physical permafrost indicators, such as active and intact rock glaciers, frozen talus, frozen till and perennial snow patches, as well as measurements of basal temperature of snow cover (BTS) and temperature of meltwater from permafrost springs (DAMM & LANGER 2006).

Rock glaciers (fig. 2) and frozen talus are characterized by flow structures and were mapped, similar as snow patches, by use of aerial photographs in combination with ground-truth data. In contrast, frozen till and ice saturated talus (fig. 3) were identified by field survey. A total of 30 rock glaciers that still containing ice were recorded (cf. DAMM 1996, 1999), indicating a mean lower limit at 2600 m a.s.l. In addition, ice saturated talus and till was mapped at 69 locations in altitudes between 2240 m a.s.l. and 3160 m a.s.l.

Ground surface temperatures (GST) were logged over a 5-years period from 2003 to 2008 using temperature loggers with a resolution of  $\pm 0.2$  K. The ONSET HOBO loggers were positioned at 32 measuring points in altitudes of 2434 to 3340 m a.s.l. on different locations of rock glaciers, frozen talus, frozen till as well as in the surroundings of perennial snow patches. The recorded temperatures of  $-2.9^{\circ}\text{C}$  to  $-9.5^{\circ}\text{C}$  on south oriented slopes (2760 to 3040 m a.s.l.) and  $-2.9^{\circ}\text{C}$  to  $-7.9^{\circ}\text{C}$  on north oriented slopes (2560 to 2680 m a.s.l.) both characterize the long-term thermal conditions in permafrost areas during winter period and below snow cover. The basal temperature of snow cover (e.g. HAEBERLI 1973, ISHIKAWA & HIRAKAWA 2000) was detected with a transportable BTS sensor in February 2006 in the proglacial area of the Rieserferner glacier and on the glacier tongue. On the basis of ground surface temperatures and BTS the lower limit of discontinuous permafrost is expected to be at 2530 m a.s.l. in north orientation and at 2730 m a.s.l. in south orientation, indicating for this limit mean annual air temperatures (MAAT) of  $-1.5^{\circ}\text{C}$  and  $-2.6^{\circ}\text{C}$  respectively (fig. 4).

#### Use of analogues and digital spatial data

Ground survey and mapping were carried out using aerial photographs of 1997 and 1999 with a resolution of 4 m, digital

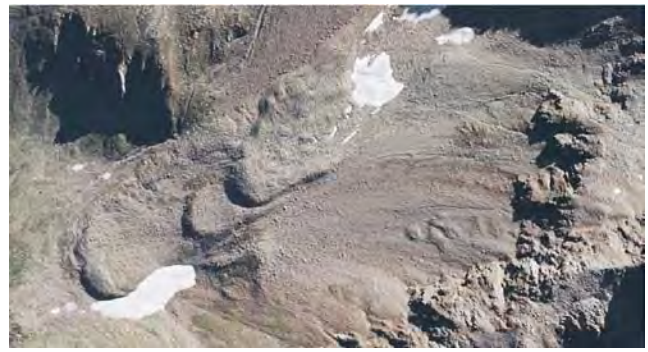


Fig. 2: Rock glaciers at 2500–2650 m a.s.l. in the Rieserferner-Ahrn Nature Park. These geomorphologic phenomena are characterized by flow structures and were mapped by use of aerial photographs and field survey (source: orthophoto of the Autonomous Province of Bozen).

Abb. 2: Durch Fließstrukturen charakterisierter Blockgletscher in 2500–2650 m ü.M. im Naturpark Rieserferner-Ahrn. Blockgletscher wurden für die vorliegende Arbeit auf der Grundlage von Orthophotos und Feldarbeiten kartiert (Quelle: Orthophoto der Autonomen Provinz Bozen-Südtirol).

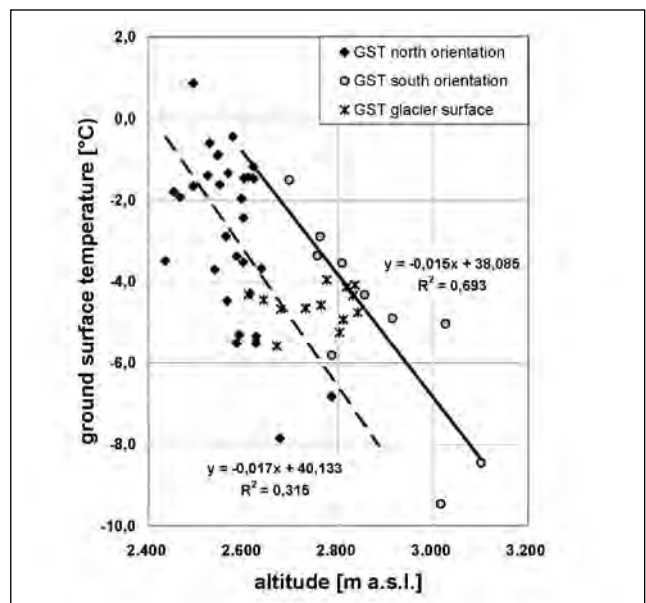


Fig. 4: Relationship between ground surface temperature and altitude derived from temperature logging and BTS measurements in the Rieserferner-Ahrn Nature Park (modified from DAMM & LANGER 2006).

Abb. 4: Beziehung zwischen BTS-Werten aus Temperaturlogger- und Sondenmessungen sowie der Geländeöhe der Messpunkte in unterschiedlichen Expositionen im Naturpark Rieserferner-Ahrn (verändert aus DAMM & LANGER 2006).



Fig. 3: Permafrost degradation in frozen till of a LIA terminal moraine at 2700 m a.s.l. (left hand) and ice saturated talus at 2600 m a.s.l. in the Rieserferner-Ahrn Nature Park (right hand) (photographs: B. Damm).

Abb. 3: Degradation von Permafrost in einer Stirnmoräne des neuzeitlichen Gletscherhochstandes um 1850 in rund 2700 m Höhe (links) und eisgesättigter Hangschutt in 2600 m Höhe im Naturpark Rieserferner-Ahrn (rechts) (Aufnahmen: B. Damm).



orthophotos with a resolution of 1 m, land cover maps and a digital elevation model (DEM 20x20 m grid) allocated by the government of the Autonomous Province of South Tyrol. Land cover maps are based on aerial photo interpretation and were available in a scale of 1:10,000 (ZANVETTOR et al. 2006).

### Mapping of debris flow initiation areas

Debris flow initiation areas were identified using digital orthofotographs of the summer/autumn 1999 with a resolution of one meter and mapped as point data within ArcGIS 9.1 using the DEM of the Autonomous Province of South Tyrol. In general, initiation areas were defined as areas on which material is released due to erosion or land slide and from which the material is subsequently transported by flow processes in channelized and open-slope debris flows. The mapped debris flow source areas were classified based on spatial distribution, elevation and slope and intersected with the grid information of the DEM.

### Assessment and simulation of permafrost distribution and glacier recession

The calculation of the present mountain permafrost distribution is based on the spatial distribution of perennial snow patches, which can be detected easily with remote sensing techniques. Widespread remnants of snow cover at the end of the late summer suggest low or negative ground temperatures underneath. Hence, the frequent occurrence of perennial snow patches may imply the existence of permafrost in the immediate vicinity (cf. FURRER & FITZE 1970, ROLSHOVEN 1982).

The spatial distribution of 3000 perennial snow patches (total area: ca. 390 ha) between 2100 m a.s.l. and 3450 m a.s.l. was proved and mapped using aerial photographs of 1985, 1997, and 1999. Taking topography, terrain orientation and altitude into account, the assessment and simulation of permafrost distribution was operated by CRYOSNOW (cf. DAMM & LANGER 2006, LANGER & DAMM 2008). CRYOSNOW represents a statistical approach for the simulation of mountain permafrost and identifies to what extent the spatial frequency of perennial snow patches correspond to the spatial distribution of permafrost. In addition to the analysis of snow patches, the simulation also tested the measured and recorded characteristics of hydrological, physical and geomorphologic permafrost indicators (see above). Measured and approximated permafrost areas are significantly correlated. Thus, these parameters allow the determination of the lower limit of discontinuous permafrost areas in the study area (cf. DAMM & LANGER 2006).

Data about the recession of the glaciers in the Rieserferner-Ahrn Nature Park were derived from studies of DAMM (1998), RÖSEN (2005) and supplementary analysis of aerial photographs and orthofoto images. Assuming a future increase of air temperature (MATULLA et al. 2002, MATULLA 2005, see below), the ELA of glaciers rises, based on a thermal gradient of 0.66° C per 100 m related to summer temperatures.

### Analysis and prediction of hazard susceptibility in relation to a variable cryosphere

The distribution of permafrost areas and glacier extent for colder and warmer climatic conditions in the past and fu-



Fig. 5: 10 m thick section of a debris flow gully in the Klamm bach catchment area. Extended surface area is covered by talus, debris cones and till deposits of Late Glacial and Holocene glaciations (photograph: B. Damm).

Abb. 5: Aufschluss in einem rund 10 Meter hohen Anriss einer Mure im Klamm bach-Einzugsgebiet. Ausgedehnte Flächen im Umfeld sind von Hangschutt, Schuttkegeln und Moränen spät- und postglazialer Gletscher bedeckt (Aufnahme: B. Damm).

ture was calculated to analyse and predict hazard susceptibility in relation to a variable cryosphere (cf. DAMM & LANGER 2006). Referring to this, the end of the LIA is supposed to represent an environmental status of increased geomorphologic stability with more extended permafrost and glacier areas compared to the present situation. In contrast, the future environmental status with an expected increase of air temperature by 1.5 K until the middle of the 21<sup>st</sup> century (cf. MATULLA et al. 2002, MATULLA 2005, IPCC 2007a, 2007b) is presumed to change to more unstable geomorphologic conditions with reduced permafrost area and shrunken glaciers compared to the current state. Thermally controlled permafrost extent is linked to the variability of the lower permafrost limit and was calculated using a thermal gradient of 0.57 K per 100 m in the study area (for general discussion see ISHIKAWA & HIRAKAWA 2000, LEWKOWICZ & EDNIE 2004).

Erosion surfaces and detachment zones of debris flows predominantly occur in unconsolidated debris bare of vegetation in the study area. Hence, current areas with related sediments and absent vegetation cover were located from land cover maps and intersected with areas of permafrost degradation derived from permafrost and glacier modelling (cf. DAMM & LANGER 2006) to identify areas that would be exposed to debris flow processes under warmer climate conditions.

## 4 Results

### 4.1 Cryosphere-hazard-interactions: The Klamm bach example

The Klamm bach is a tributary stream of the Antholz creek and located in the south-eastern part of the Rieserferner Mountains (see fig. 8). The south facing drainage basin comprises an area of 6.5 km<sup>2</sup>. Altitudes reach from 1,310 m to 3,270 m a.s.l. along a horizontal distance of 3,000 m. The catchment area of the Klamm bach is composed of plutonic rock of the Rieserferner-Tonalite. Extended surface area is



Fig. 6: Klammbach catchment area and main debris flow channel (left-hand photo) and extent of the August 2005-debris flow deposits in Antholz-Mittertal village (right-hand photo) (source: Autonomous Province of Bozen, flood protection office).

Abb. 6: Einzugsgebiet des Klammbachs mit dem zentralen Murgraben (links) und Ausdehnung der Murablagerungen vom August 2005 in Antholz-Mittertal (rechts) (Quelle: Autonome Provinz Bozen-Südtirol, Amt für Wasserschutzbauten).

covered by talus, debris cones and till deposits of Late Glacial and Holocene age (fig. 5). In general, in the upper parts of the Klammbach basin, in particular above the timberline, this material is poorly consolidated and sparsely covered by vegetation.

Several debris flows in the past were supplied from the above-mentioned debris sources. In order to evaluate the sediment characteristics in the Klammbach catchment, different sediment bodies were sampled and analysed concerning their particle size distribution. Grain size distribution was analysed from till of moraine bodies, talus of debris cones and frost debris of an ancient rock glacier (cf. tab. 3). Five samples were taken from debris flow initiation areas and another three samples from fluvial reworked sediments (cf. SANDMEIER et al. 2012). In general, sampling was undertaken at natural cuts in the proximity of debris flow gullies and deposits. Particle size of the sediment was established according to DIN 18123 by wet-sieving using the required mesh width (DIN 18123).

Analyses result in contents of 37% to 76% of coarse grain fraction. Fine grain fraction contents high percentage of Sand (70–95%) and partly of silt (4–27%), the percentage of clay (0.3–2%), in contrast, is low (cf. tab. 3). The different sediments show substantial similarity in their composition and the grain size distribution of samples from debris

flow initiation areas is highly similar to samples from debris cones or the rock glacier. According to grain size distribution (BONNET-STAU 1999), existing proportion of fine and coarse grain (IVERSON 2005) and weak consolidation these sediments are potential sources for debris flows and have to be considered to be controlling factors of these processes (cf. ZNAMENSKY & GRAMANI 2000, REMAÎTRE et al. 2005).

In summer 2005 a series of three debris flows occurred in the Klammbach gully. In total, about 140,000 m<sup>3</sup> of debris was deposited on the alluvial cone in Antholz-Mittertal village (fig. 6). Agricultural area was inundated and the Antholz creek was dammed and displaced (DAMM 2005). Run-off events and the successive two debris flows in July 2005 were triggered by comparatively low rainfall intensities of 0.5 to 3 mm per minute, which were measured in the upper part of the drainage basin. In contrast, torrential rainfall of 30 mm during 60 minutes on August 1<sup>st</sup> 2005 triggered the third debris flow with a total volume of about 100,000 m<sup>3</sup> (cf. SANDMEIER et al. 2012). This major event originated from a steep debris cone, sparsely covered by vegetation and located below rock flanks and, on the base of ground survey, debris-covered dead ice. The latter are relics of a glacier that melted continuously during the last 80 years. Overland flow from bedrock and dead ice was rapidly conducted into the sediment. Due to several former but smaller events (e.g.

Tab. 3: Grain size distribution of different sediments in the Klammbach basin (modified from SANDMEIER et al. 2012). \*: sample originates from debris flow initiation area. +: sample originates from fluvial reworked sediment.

Tab. 3: Korngrößenverteilung unterschiedlicher Sedimente im Einzugsgebiet des Klammbachs (verändert nach SANDMEIER et al. 2012). \*: Probe aus Muranrissbereich. +: Probe aus fluvial umgelagertem Sediment.

Source / Grain size class [mm]	Glacial sediment */+	Glacial sediment *	Glacial sediment	Glacial sediment +	Glacial sediment	Debris cone, talus *	Debris cone, talus *	Debris cone, talus *	Debris cone, talus	Debris cone, talus	Debris cone, talus +	Rock glacier, protalus
Coarse grain [> 2]	47.35	72.28	52.62	36.71	62.55	67.10	55.79	53.72	53.45	75.79	56.75	55.26
Fine grain [< 2]	52.65	27.72	47.38	63.26	37.45	32.90	44.21	46.28	46.55	24.21	43.25	44.74
Sand [0,02-1]	79.59	71.61	84.14	82.84	87.08	74.23	83.55	83.92	85.57	95.30	69.87	77.70
Silt [0,002-0,02]	18.54	26.55	15.10	15.83	12.00	23.74	14.99	13.07	13.73	4.41	27.20	20.32
Clay [<0,002]	1.88	1.84	0.76	1.33	0.92	2.03	1.46	3.01	0.70	0.29	2.93	1.98

1882, 1947, 1979) the Klammbach drainage system was already known to be an active debris-flow system (AMT FÜR WASSERSCHUTZBAUTEN 2005). Hence, a storage basin with total retention capacity of 100,000 m<sup>3</sup> was constructed on the alluvial cone as a response to the 2005-events.

The following parameters favoured and/or triggered runoff and debris flow processes in particular (AMT FÜR WASSERSCHUTZBAUTEN 2005, DAMM, 2005, 2008):

- On the base of ground survey, extended areas of low or not consolidated talus and till deposits occur in the sparse or not vegetated areas above 2,200 m a.s.l. They enclose an area of about 25% of the total drainage basin (6.5 km<sup>2</sup>).
- The source area of the August-2005-debris-flow was located in a more than 10 m thick talus cone in contact to bedrock. As result of the GIS-based simulations, this area was most likely underlain with permafrost and stable related to soil-mechanics until the middle of the 19<sup>th</sup> century. Ground survey indicates that meanwhile permafrost has completely melted.
- Since 1940 mean annual air temperature rose by 1 K in the Antholz valley (meteorological station of Antholz-Mittertal). The increase of MAAT caused a rise of the lower permafrost limit of most likely 100–180 m in the Klammbach drainage basin, using a mean thermal gradient of 0.57 K/100 m.
- Terrain surface properties controlled overland flow. The existence of extended debris covered dead ice without water retention capacity and steeply sloping rock flanks (approx. 60°) may have intensified runoff quantity and velocity in the rainfall area. Field study subsequent to the event indicates that runoff infiltrated into the talus cone and increased the soil water pressure.

On the base of ground survey and permafrost modelling, the debris flow processes in the Klammbach basin supposed to be substantially controlled by permafrost degradation in the study area. The detachment area of the main debris

flow mentioned above is south exposed and located at an altitude of 2,600 m a.s.l. Until a few decades ago mountain permafrost still occurred in altitudes and expositions like these throughout the Rieserferner-Ahrn Nature Park (e.g. DAMM & LANGER 2006).

The degradation of permafrost due to the increase of MAAT caused soil mechanical instabilities of sediments and slopes. Similar to the Klammbach example, our study during the last two decades could frequently verify cryosphere-hazard-interactions in other areas of the Rieserferner-Ahrn Natural Park and the surroundings. This information is recorded in the natural hazard database for the study area and partly in the ED-30 hazard database of the South Tyrol state (ZISCHG et al. 2007).

#### 4.2 Thermally-controlled development of the cryosphere in the Rieserferner-Ahrn Nature Park

##### *Development of glaciation*

The glacier retreat since the end of the LIA is comparably well documented in the Rieserferner-Ahrn Nature Park. The glacier surface decreased by 50% in the Rieserferner Mountains and by 65% in the upper Ahrntal Mountains over the past 150 years (DAMM 1998, RÖSEN 2005). In general a reduction of glacier area is observed for all glaciers of the eastern European Alps since the end of the LIA (ZEMP et al. 2007) however it was slightly above-average in the study area compared to other mountain ranges. Glacier volume in total decreased by 70% in the Rieserferner-Ahrn Nature Park, even though changes of individual glaciers show a wide variability, depending on their size, physiographic setting and altitude (cf. LAMBRECHT & KUHN 2007). For this reason, maximum losses generally occurred in ablation zones, but numerous small glaciers completely disappeared.

At present glaciers of the Rieserferner-Ahrn Nature Park are retreating with increasing equilibrium line altitudes up to 3,000 to 3,100 m a.s.l. since the latest advance during

Tab. 4: Equilibrium line altitudes (ELA) of glaciers in the Rieserferner-Ahrn Nature Park related to climatic conditions at the end of LIA (ca. "1850"), the present day situation (basis "2000") and compared with the scenario +1.5 K. Assuming that alpine summer temperature increases by +1.5 K, the ELA of glaciers rises by 230 m in the study area, based on a thermal gradient of 0.66° C per 100 m related.

Tab. 4: Mittlere Gleichgewichtslinie (ELA) auf Gletschern im Naturpark Rieserferner-Ahrn für Temperaturbedingungen zum Ende der „Kleinen Eiszeit“ (um „1850“), für aktuelle Bedingungen (um „2000“) und für ein Temperaturszenario +1.5 K. Auf der Grundlage eines thermischen Gradienten von 0.66° C /100m ist bei Zunahme der Sommertemperatur um 1.5 K mit einem Anstieg der Gleichgewichtslinie um rund 230 m zu rechnen.

Glacier area	ELA „1850“ [m]	ELA, rise [m]	ELA „2000“ [m]	ELA +1.5 K [m]
Rieserferner Mountains	2747	99	2846	3077
Ahrntal Mountains	2586	107	2693	2920
Study area, mean	2667	105	2772	2999

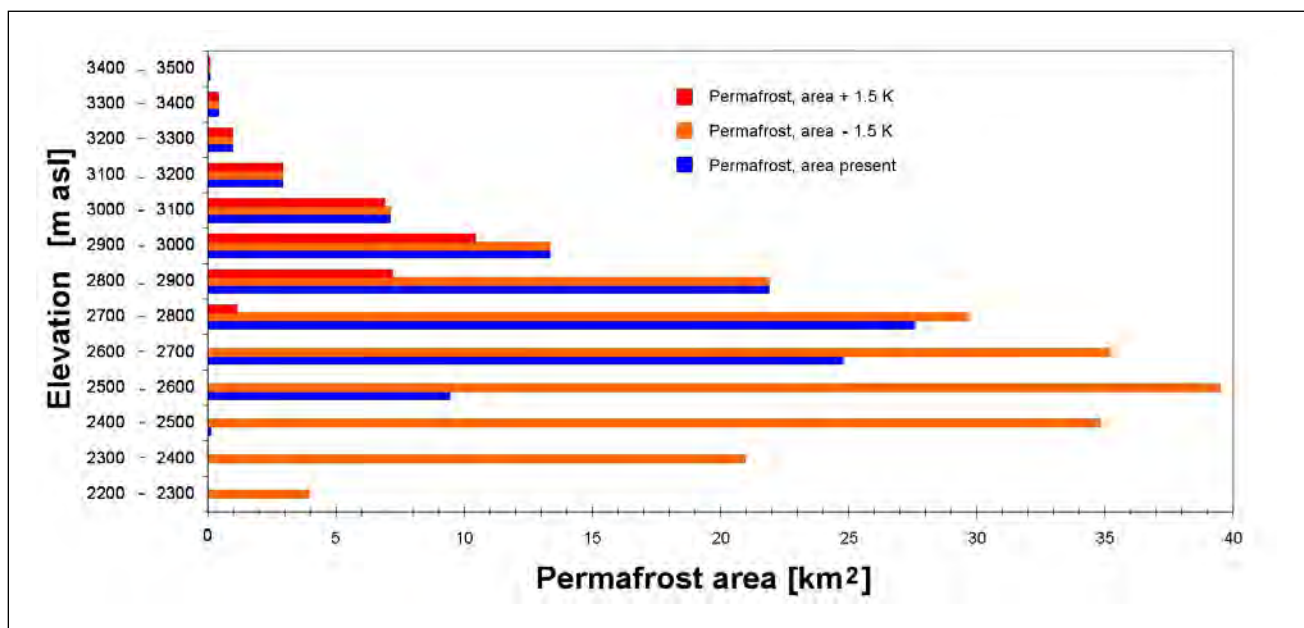


Fig. 7: Present-day extent of permafrost area in the Rieserferner-Ahrn Nature Park related to altitude and different thermal scenarios with lower (-1.5 K) and higher (+1.5 K) mean annual air temperature.

Abb. 7: Von Permafrost unterlagerte Flächen im Naturpark Rieserferner-Ahrn unter aktuellen Klimabedingungen sowie bei um 1.5 K höheren bzw. tieferen Temperaturen, differenziert nach Höhenstufen.

the 1970ies and 1980ies. Air temperature of the ablation period showed a rise of 0.3 to 0.5 K since 1981 and precipitation increased by 14%, as data of the meteorological station Antholz-Mittertal (1236 m a.s.l.) indicate. Based on reports of glaciological survey (COMITATO GLACIOLOGICO ITALIANO 1977–2011), the development of the ablation area from about <50% of the glacier surface at the end of the 1970s to about 20% in 2010 (cf. SECCHIERI & VALENTINI 1985, AUTONOME PROVINZ BOZEN 2010, 2011) is connected with the rise of ELA of 150–300 m, which seems to be the principle reason of the latest glacier recession in the study area (cf. ESCHER-VETTER & SIEBERS 2007).

Assuming the scenario of MATULLA et al. (2002) and MATULLA (2005) that forecast the increase of alpine summer temperature by +1.5 K during the next 40 years, the ELA of glaciers can rise by 230 m in the study area (cf. chapter 3, tab. 4). Hence, numerous glaciers continue to retreat and can be affected by significant loss of volume and extent. The total glacier area decreases to 1.1 km<sup>2</sup> in the Rieserferner Mountains and to 0.7 km<sup>2</sup> in the upper Ahrntal Mountains.

#### Development of permafrost distribution

Different scenarios of the spatial variation of mountain permafrost in the Rieserferner-Ahrn Nature Park (DAMM & LANGER 2006, DAMM 2008) are interrelated with lower (-1.5 K) and higher (+1.5 K) MAAT compared to the present situation. The lower temperature represents the environmental situation before significant warming at the end of LIA, while the simulation with the higher temperature may describe possible future developments according to predicted climatic scenarios. The variability of the lower permafrost limit was calculated using a thermal gradient of 0.57 K per 100 m in the study area.

On the basis of field evidence and permafrost modeling, the present permafrost area in the Rieserferner-Ahrn Nature Park and that related to the thermal conditions at the end of LIA is calculated to be 109 km<sup>2</sup> and 211 km<sup>2</sup>, respectively. Similarly, permafrost area would decrease by approximately 72% compared to present conditions with increasing MAAT of +1 to +2 K (fig. 7). The permafrost areas calculated for the present conditions and different sce-

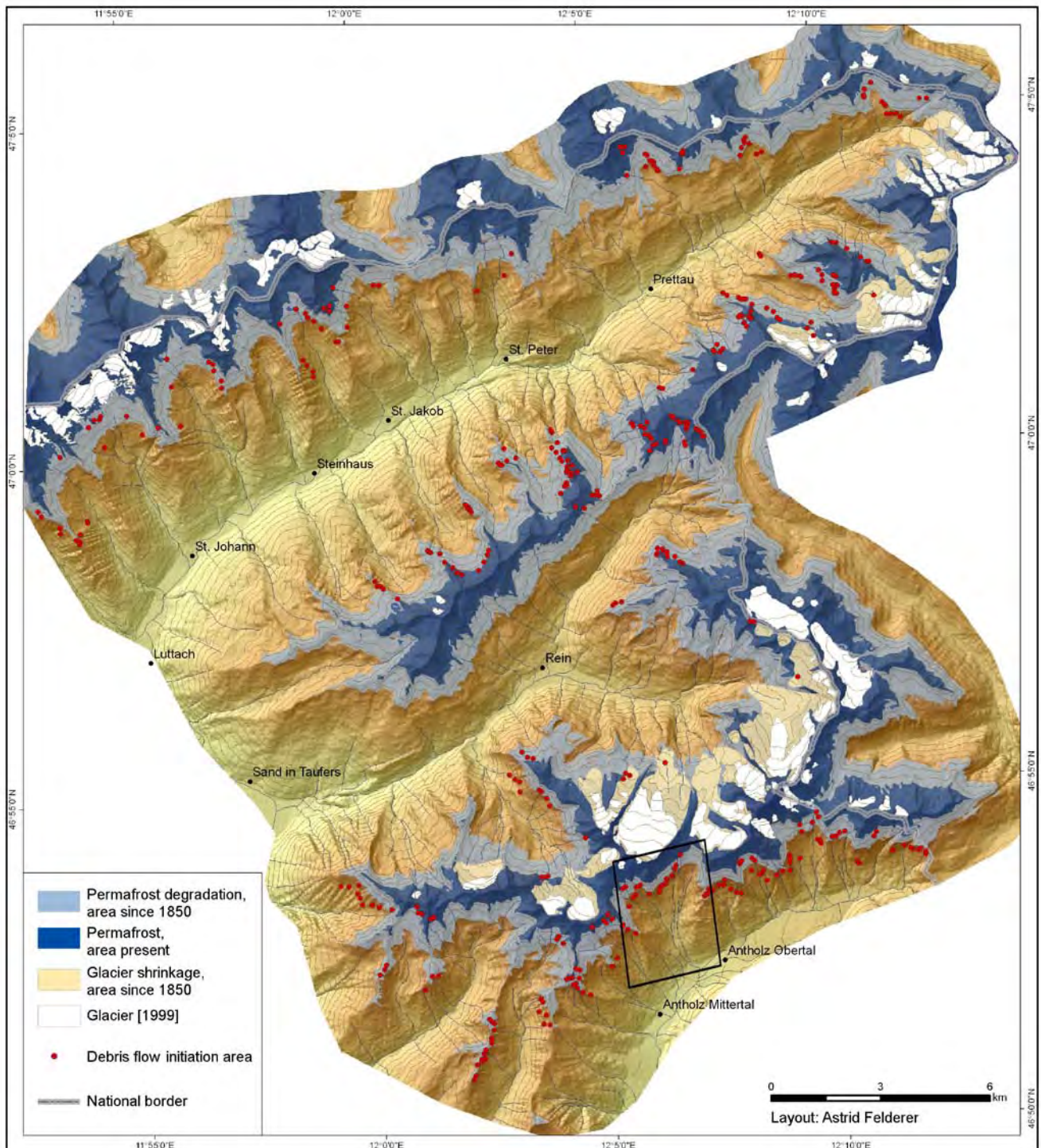


Fig. 8: General map of glacier extent, permafrost distribution and debris flow initiation areas in the Rieserferner-Ahrn Nature Park. In total, 408 debris flow initiation zones were mapped. The Klammbach drainage area is located in the south-eastern part of the study area (box).

Abb. 8: Gletscherflächen, Verbreitung von Permafrost und Lage von Muranrissen im Naturpark Rieserferner-Ahrn. Insgesamt konnten 408 Muranrisse kartiert werden. Das Einzugsgebiet des Klammbachs liegt im südöstlichen Teil des Untersuchungsgebiets (Kasten).

narios include all types of permafrost landforms and materials, such as loose rock, talus, rock glaciers and bedrock. Based on these assumptions present and future debris flow hazard potential is estimated as follows.

#### 4.3 Debris flow initiation zones and debris flow hazards

##### *Characterization and distribution of debris flow initiation zones*

In total, 408 debris flow initiation zones were identified and

mapped in the study area (fig. 8). In general, these zones occur throughout the study area and are concentrated in drainage basins of torrents and in various debris accumulations. Debris flow initiation zones are located to almost 50% in areas, where discontinuous permafrost degraded since the end of the LIA and to approximate 20% within the present area of probable permafrost. In contrast, only few detachment areas were identified in positions, where discontinuous permafrost is supposed to outlast the increase of temperature by +1.5 K.



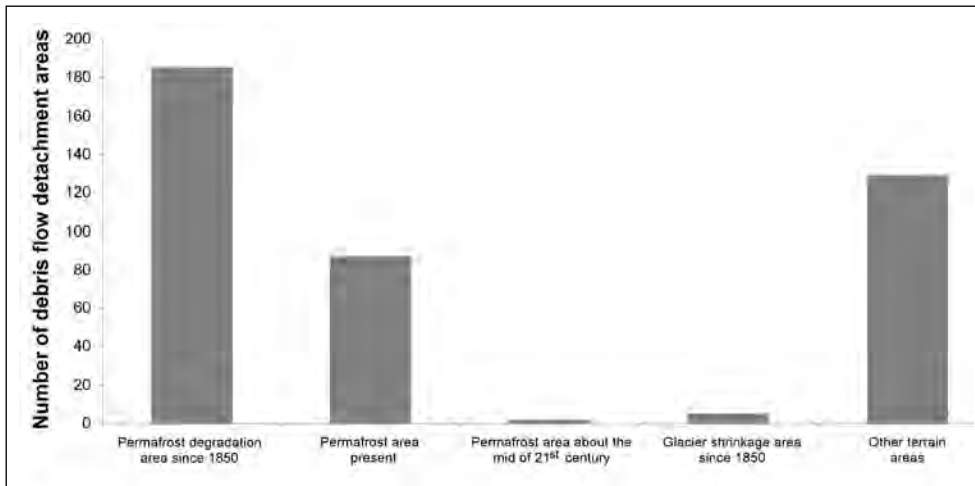


Fig. 9: Debris flow detachment areas in the Rieserferner-Ahrn Nature Park related to temporal variability of permafrost distribution and glacier extent.

Abb. 9: Häufigkeit von Muranrissen im Naturpark Rieserferner-Ahrn differenziert nach zeitlicher Variabilität von Permafrostverbreitung und Gletscherausdehnung.

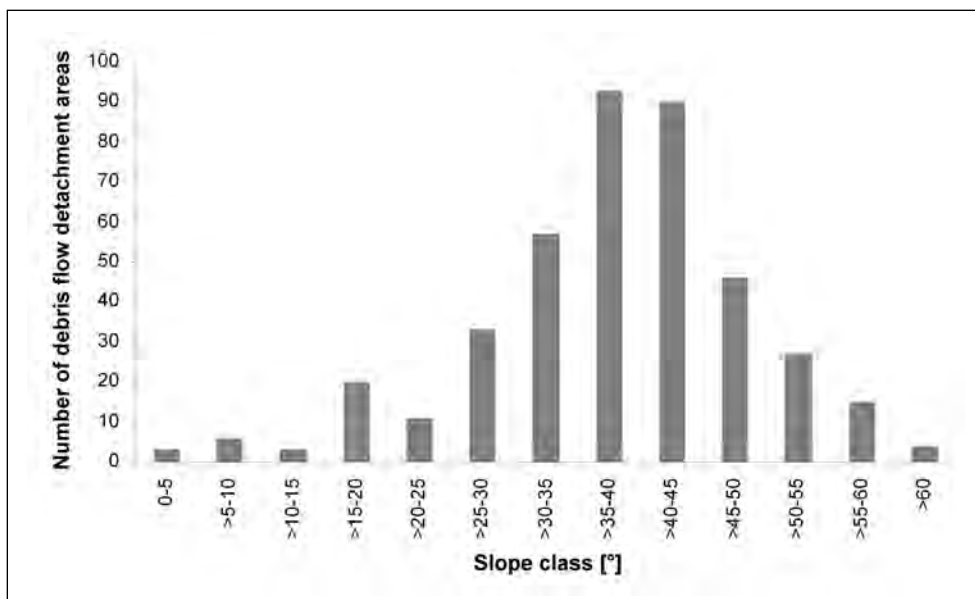


Fig. 10: Debris flow initiation areas in the Rieserferner-Ahrn Nature Park related to slope gradient. Two thirds occur at slopes of more than 35°.

Abb. 10: Verteilung von Muranrissen im Naturpark Rieserferner-Ahrn differenziert nach der Geländeneigung. Zwei Drittel der Prozesse ereignen sich oberhalb von 35°.

### Debris flow source areas in reference to altitude, surface characteristics and slope

Debris flow detachment areas occur between 1,800 and 3,000 m a.s.l. in the Rieserferner-Ahrn Nature Park. The major part of initiation areas is located in altitudes between 2,100 and 2,800 m a.s.l. with a maximum between 2,300 and 2,700 m a.s.l., which coincide with the distribution of present discontinuous permafrost calculated by DAMM & LANGER (2006). Below 1,800 m a.s.l. close vegetation cover consisting of krummholz and forest as well as relief properties reduce the potential of debris flow sources.

Almost 47% of the mapped debris flows originated in loose rock, such as thick talus and till deposits, 37% in areas of alpine meadows and another 13% in shallow detritus covering bedrock (fig. 9). In general, source areas appear at slopes up to 66° with a maximum occurrence between 35–40° in the study area (fig. 10). Two thirds of these zones occurred at a slope steeper than 35°.

### Permafrost degradation and stability of loose rock deposits

The distribution of debris flow initiation zones coincides significantly with the spatial occurrence of unconsolidated debris in the study area. Thus, areas are susceptible to debris flow processes, where permafrost ice is degrading and talus

or till deposits exist in (steep) slopes, where soil mechanical stability is just near the limit.

In general, present-day debris flows mainly originate from areas covered with loose rock, where vegetation cover is absent and where permafrost degraded since the end of the LIA. Corresponding sediments cover about 38 km<sup>2</sup> and are uniformly distributed in the Rieserferner-Ahrn Nature Park. Furthermore, approximately 15 km<sup>2</sup> (39%) of the permafrost degradation area has a critical slope of 25–45° (fig. 10) and is therefore susceptible to debris flow processes with increasing MAAT.

According to field evidence and GIS-based permafrost modelling, the total area of loose rock underlain by permafrost amounts to 21.7 km<sup>2</sup> in the Rieserferner-Ahrn Nature Park (cf. DAMM & LANGER 2006). Increasing permafrost degradation enhances instability of talus and till deposits and generates possible debris flow initiation areas in the future. According to topographic and environmental conditions of the study area the complete degradation of permafrost in areas of loose rock is expected due to a long-term increase of MAAT by +1.5 K. In such a scenario, permafrost in rock faces solely remains above 3,000 m a.s.l. and cover 1.8 km<sup>2</sup>. Overall, the area of debris sources and possible debris flow initiation zones increases to 20.1 km<sup>2</sup> in the study area.

## 5 Discussion

Climate variability mainly controls the cryosphere by changing air temperature and influences the mass balance of glaciers and the thickness and spatial extent of permafrost. As a consequence, glacier recession and permafrost degradation affect slope stability and enhance the susceptibility to mass movement and debris flow processes, as it is demonstrated in several studies (e.g. HAEBERLI 1999, HAEBERLI et al. 1999, GRUBER et al. 2004, DAMM & FELDERER 2008).

In the context of debris flow processes, it is fundamental that permafrost degradation influences debris supply to talus and torrential systems, for instance by rock fall from destabilized rock walls and/or by permafrost creep processes (DAVIES et al. 2001, GRUBER et al. 2004, DAMM 2007). In addition, the lowering of the permafrost table may increase the susceptibility of slopes for instabilities and consequently the initiation of debris flows. In this case, the thickening of the active layer can increase sediment availability in potential debris flow initiation zones and may also reduce the shear strength of debris (for this discussion see SATTLER et al. 2012).

The permafrost table acts as aquiclude and as potential failure plane during periods of elevated pore pressure, such as thaw periods or after strong summer rainfall (cf. LARSSON 1982). Laboratory experiments of RIST (2007) showed that the active-layer instability was much stronger affected by the release of fine-grained material formerly fixed in the ice matrix than by the oversaturation of material at the base of the active layer. But for all, the most important factor seems to be the loss of internal ice in loose rock by permafrost melting, a process that was frequently observed in the study area (fig. 11, DAMM & LANGER 2006, DAMM 2008, SANDMEIER et al. 2012). The melting of interstitial ice leads to the loss of cohesion of previously frozen debris, which becomes available for mobilization by surface processes. The development of enhanced and irregular pore space and incomplete consolidation following permafrost thaw influences the hydraulic conductivity and reduces the overall mechanical slope stability (cf. ZIMMERMANN & HAEBERLI 1992).

The hypothesized connection between permafrost degradation in non-creeping slope material and enhanced debris flow activity has not been proven yet, but observations of debris flows originating in areas presumed to be at the margin of contemporary permafrost distribution support this hypothesis (e.g. ZIMMERMANN and HAEBERLI 1992; STÖTTER 1994; DAMM and FELDERER 2008, SATTLER et al. 2012). Furthermore, it is supposed that slopes currently underlain by degrading permafrost become less stable with ongoing climate change, even with increasing altitude (HARRIS et al. 2001, 2009, STOFFEL & HUGGEL 2012).

The cryosphere development in mountain regions has particular importance for issues of future research on climate impact (IPCC 2007a, 2007b). According to the „Special Report“ of the IPCC from 2011 the number of cold days and nights will decrease, while the number of warm days and nights will increase in Europe (IPCC 2012). Corresponding changes are supposed to affect the alpine cryosphere directly by intensifying the continuous deglaciation and permafrost degradation.

In the present study the assessment of future natural hazard potential in the Rieserferner-Ahrn Nature Park is based



Fig. 11: Internal ice of talus and till deposits uncovered by sliding at the Hochgall peak north face in July 2011. Thickening of the seasonally unfrozen active layer and the loss of internal ice by permafrost degradation destabilize loose rock and favour the detachment of debris (photograph: B. Damm).

Abb. 11: Durch Rutschung freigelegtes Eis in Hangschutt- und Moränenablagerungen in der Nordflanke des Hochgalls im Juli 2011. Zunehmende Mächtigkeit der saisonalen Auftauschicht und tiefgreifende Permafrostdegradation destabilisieren Lockergesteine und begünstigen deren Abgleiten (Aufnahme: B. Damm).

on a scenario that forecasts an increase of air temperature by 1.5 K until the middle of the 21<sup>st</sup> century. Even though the world climate report (IPCC 2007b) forecasts a global rise of mean air temperature by 1.1 to 6.4 K by the end of the 21<sup>st</sup> century, in our study a more detailed differentiation seemed to be necessary for assessing climate impact. The climate change scenario assumes that the MAAT increases by +2 K within the next 30 to 40 years in the Eastern Alps showing a pronounced warming during summer time with +2.5 K (cf. MATULLA 2005, OCCO 2007, FORMAYER et al. 2008). In this scenario, the mean annual precipitation hardly varies, while precipitation exhibits a significant seasonal variability. Overall, an increase of summer temperature (summer-JJA-season) by +1.5 to +2.5 K and winter temperature (winter-DJF-season) by +2 to +3 K is assumed for the study area and the surroundings. Hence, this study is based on a restrained assessment for the degradation of cryosphere and the resulting natural hazard potential by using a moderate climatic scenario of +1.5 K.

The warming of the cryosphere in the study area exposed extended sources of destabilized debris in till deposits, creeping permafrost and talus during the past 150 years. Such sediments are susceptible to debris flow processes now. About 68% of the mapped debris flows originated in areas of destabilized debris. Furthermore, 74% of the detachment zones are located between 2,300 and 2,700 m a.s.l. Taking into account the climatic conditions of differing exposure this altitude range coincides with the present lower limit of permafrost in the Rieserferner-Ahrn Nature Park, which rose due to increased air temperature since the end of the LIA. In this perspective, the observable evidence of the recent past suggests a spatial expansion of the debris flow initiation area under warming conditions. In the Rieserferner-Ahrn Nature Park, the potential debris flow detachment area is calculated to increase by 52%, if long-term MAAT rises by +1.5 K.

Similarly to the results of our study, significant correlation between debris flow source areas and permafrost degradation areas were found amongst others by ZIMMERMANN & HAEBERLI (1992) in the Swiss Alps and by STÖTTER et al. (2003, 2012) in the Suldental in northern Italy. In contrast, the case study of SATTLER et al. (2012) documents that distinct changes in the spatial position of debris flow initiation zones mainly occur at elevations above the current permafrost degrading areas in the Schnals valley (Ötztal Alps / Italy). The study was based on comparing debris flow activity since 1983 with modelled contemporary permafrost distribution. The authors conclude that the changes detected in debris flow activity were probably not influenced by atmospheric warming-induced permafrost degradation, but were connected with the thickening of the active layer. In accordance to the study of SATTLER et al. (2012), about 20% of debris flow detachment area in the Rieserferner-Ahrn Nature Park occurred above the simulated lower limit of permafrost. On the one hand, this may be due to uncertainties of the modelling of permafrost distribution. On the other hand, it seems to be probable that numerous debris flows are related to the thickening of the seasonally unfrozen active layer and were triggered by increased ground-water circulation and pressure.

About 40% of debris flow detachment areas are located at slopes between 26–45° in the Rieserferner-Ahrn Nature Park. Depending on lithologic characteristics slope inclinations like that are principally assumed to be susceptible to debris flow processes (e.g. JOHNSON & RAHN 1970, LEWIN & WARBURTON 1994). Thus, relevant terrain needs to be especially considered when assessing potential debris flow hazards, when permafrost continues to thaw with warming conditions.

The present study has tried to shed light on the effects of climate change on the high mountain cryosphere and the response of morphodynamics. With focus on the identification of present causalities and the prediction of potential future processes, the investigation has addressed the impacts of glacier recession and permafrost degradation on the spatial occurrence and distribution of debris flow initiation areas. Despite of the recent progress in climate impact research, the study illustrated that some questions have to be treated in greater detail, so as to bridge the gaps in knowledge, which still exist.

## 6 Conclusions

Atmospheric warming in high mountain environments causes a range of impacts, including glacier recession, reduction of permafrost extent and distribution as well as changes in thermal permafrost properties. Furthermore, it is likely that climate change affects the occurrence of natural hazards, like shallow landslides and debris flows, as their initiation is related to the degradation of the cryosphere. However, changes in mass movement activity can hardly be detected so far.

The present study indicates the importance of atmospheric warming for the occurrence and susceptibility of debris flow processes in the eastern Alps. There is first evidence that it is possible to quantify the regional debris flow hazard potential on the basis of field survey, in-situ

measurements, climate data analyses, and GIS-based simulations for different climate scenarios and variable geomorphic stability.

The presented scenarios of the future distribution of unstable areas susceptible to debris flow hazards are based on the assumption that MAAT increases by 1.5 K by the middle of the 21<sup>st</sup> century. This may be presumed to be a moderate increase of temperature in relation to the predicted climate development of the IPCC. The quantitative results of the present study can provide a basis for spatial planning and risk assessment.

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# Geomorphological and geophysical analyses in a landslide area near Ebermannstadt, Northern Bavaria

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**Abstract:**

The region close to Ebermannstadt is highly prone to landslide activity. Geomorphological mapping, geophysical surveys (electric resistivity and seismic refraction) and substrate analyses were applied to investigate a landslide which was triggered in 1957. The study aims to reconstruct observations of the 1957 event and to compare the earlier situation with the recent surface and subsurface conditions.

It was possible to trace the former observations in the relevant slope area and furthermore, to classify different slide masses and processes. The study also shows area-wide occurrence of ancient slide masses, which were incorporated during the landslide event of 1957. A remobilization of these accumulations cannot be excluded.

**Geomorphologische und geophysikalische Untersuchungen auf einer Rutschung bei Ebermannstadt, Nordbayern****Kurzfassung:**

Die Region um Ebermannstadt ist stark anfällig für Massenbewegungen. Eine Rutschung aus dem Jahr 1957 wurde geomorphologisch kartiert und mit geophysikalischen Messungen (Gleichstromgeoelektrik sowie Refraktionsseismik) untersucht. Die Ziele der Untersuchungen waren eine Rekonstruktion der Beobachtungen von 1957 sowie deren Vergleich mit der heutigen Oberfläche und dem oberflächennahen Untergrund.

Die Ergebnisse zeigen, dass die früheren Beobachtungen auch heute noch Gültigkeit besitzen. Darüber hinaus ist es möglich, einzelne Rutschungselemente und Prozesse zu differenzieren. Die Studie zeigt zudem die Verbreitung von alten Rutschmassen, welche in die Rutschung von 1957 eingebunden waren. Eine erneute Remobilisierung dieser Ablagerungen kann nicht ausgeschlossen werden.

**Keywords:**

*Cuesta landscape, Franconian Alb, landslide, earth flow, geomorphological mapping, Electric resistivity tomography, Seismic refraction tomography*

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## 1 Introduction

Cuesta scarp slopes are highly prone to slide activity due to susceptible geological and geomorphological conditions. The development of cuesta landscapes like the Franconian Alb is controlled by slightly inclined permeable and non-permeable bedrock. The geological setting results in the characteristic cuesta landforms of almost flat backslopes and steeper front slopes (SCHMIDT & BEYER 2001). The latter usually consist of steep, sometimes cliff-like upper slopes, formed in the resistant, permeable bedrock and a less inclined lower slope in non-resistant, impermeable layers. This bipartite structure leads to a strong disposition for mass movements. Water can be easily drained through fissures in the upper bedrock – in particular in karstic structures of limestones. In the middle and lower slope areas, impermeable clay- and marlstones lead to an accumulation of water, decreased consolidation and the formation of slip surfaces.

Although mass movements are a common hazard in low mountain areas like the Franconian Alb, only devastat-

ing events, endangering buildings, roads or infrastructure are noticed in public (HÜTTEROTH 1994). Various scientific landslide studies at the Franconian Alb investigate the distribution of landslides for selected regions. In general, the focus is set on internal (lithological, structural and internal morphometric) controlling factors (VON DER HEYDEN et al. 1993, HÜTTEROTH 1994, VON DER HEYDEN 2004), comparable to studies from other German low mountain areas (e.g. WENZEL 1994, TERHORST 2001, BEYER 2002, SCHMIDT & BEYER 2001). Geological studies (e.g. ZÜRL 1980, HAMMER 1984) primarily comprise soil and rock mechanisms or deal with hydraulic patterns within slide masses.

One of the first works on landslide distribution in the Wiesent River valley was published by DORN (1920). After the respective movements FREYBERG (1957, 1961), MÜLLER (1957) and HEGENBERGER (1961) described visible processes, triggers and morphological landslide elements of the major slides close to Ebermannstadt and provided first information on the slide events. The present study is based on the observations of MÜLLER (1957) at the Hasenberg landslide area.

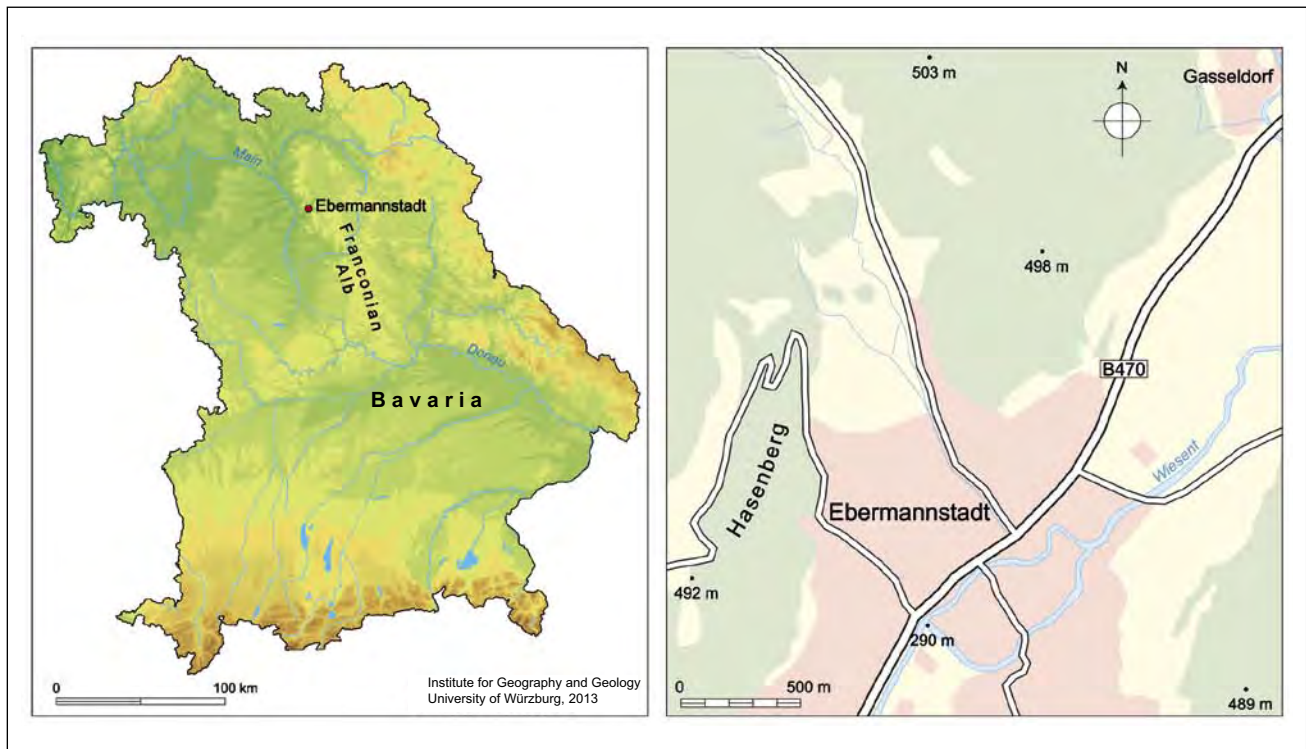


Fig. 1: Location map of the study area.

Abb. 1: Übersichtskarte des Arbeitsgebiets.

In order to analyze controlling factors as well as past processes in the landslide mass, an accurate inventory of the slope is necessary (MCKEAN & ROERING 2004). High resolution Digital Terrain Models (DTM) derived from airborne laser altimetry (LiDAR) provide detailed information of the earth's surface (VAN WESTEN et al. 2002) and therefore enable more detailed mapping of relief units, which include land system, landform and elementary form (MINAR & EVANS 2008). As surface morphology provides indications for determining the type of a landslide (HUTCHINSON 1995), high resolution datasets are commonly used for field mappings on any scale (e.g. MONTGOMERY et al. 2000, LEE 2001, CROSTA & AGLIARDI 2002). The combination of these geo-information tools with additional deterministic input data, such as shallow geophysics, provides a profound method for analyzing landslide hazard and vulnerability as well as risk assessment (BRUNO & MARTILLIER 2000, VAN WESTEN et al. 2002).

Within the last decades geophysical methods have developed as standard methods for the minimal-invasive investigation of the shallow subsurface in the fields of environmental research (e.g. TELFORD et al. 1990; SCHROTT et al. 2003; REYNOLDS 2011), and likewise in the studies of landslides (i.e. McCANN & FORSTER 1990; HACK 2000; BICHLER et al. 2004; FRIEDEL et al. 2006; JONGMANS & GARAMBOIS 2007; SCHROTT & SASS 2008; SOCCO et al. 2010). Recently, the development of 2D and 3D techniques can be regarded as a major advance for imaging the often complex and heterogeneous subsurface. A comprehensive review on the geophysical investigation of landslides is given by JONGMANS & GARAMBOIS (2007). So far, no geophysical studies were applied on landslides in the region of Ebermannstadt.

As MÜLLER (1957) described slide processes by the use

of geomorphological observations, this study obtains slide processes using surface and subsurface data: A Digital Terrain Model (DTM) was primarily employed for geomorphological mapping, focusing on reconstructing formerly described processes on the surface. Based on these data designated geophysical surveys aimed at obtaining new insights into the subsurface. Besides, the focus was set on defining the type of movement by applying the concepts of VARNES (1978), HUTCHINSON (1988) and CRUDEN & VARNES (1996). A further aim is to assess the susceptibility of the slopes for future movements.

With the towns' expansion already reaching the foot of the landslide, detailed studies of the slide mass are essential for the Hasenberg area.

## 2 Regional characteristics and study area

The study area is situated in the Franconian Alb, approximately 40 km north of Nuremberg (Fig. 1). The Franconian Alb is part of the south German cuesta landscape, formed by Mesozoic rocks from the upper Triassic, Jurassic and Cretaceous. Related to the occurrence of mass movements, the cuesta scarps of the Rhaethian/Hettangian (Upper Triassic/Lower Jurassic), the Aalenium (Middle Jurassic) and the Oxfordian (Upper Jurassic) are of major interest.

In general, the slopes of the cuesta scarps are affected by different types of mass movements, such as topples, slides, lateral spreads and flows, either in single or in combined occurrence. Falls can be observed on vertical cliffs and fronts of slide blocks. Unconsolidated material and/or old slide masses often cause secondary (translational) landslides.

Due to its extension from north to south, the northern Franconian Alb acts as a barrier for weather-effective air

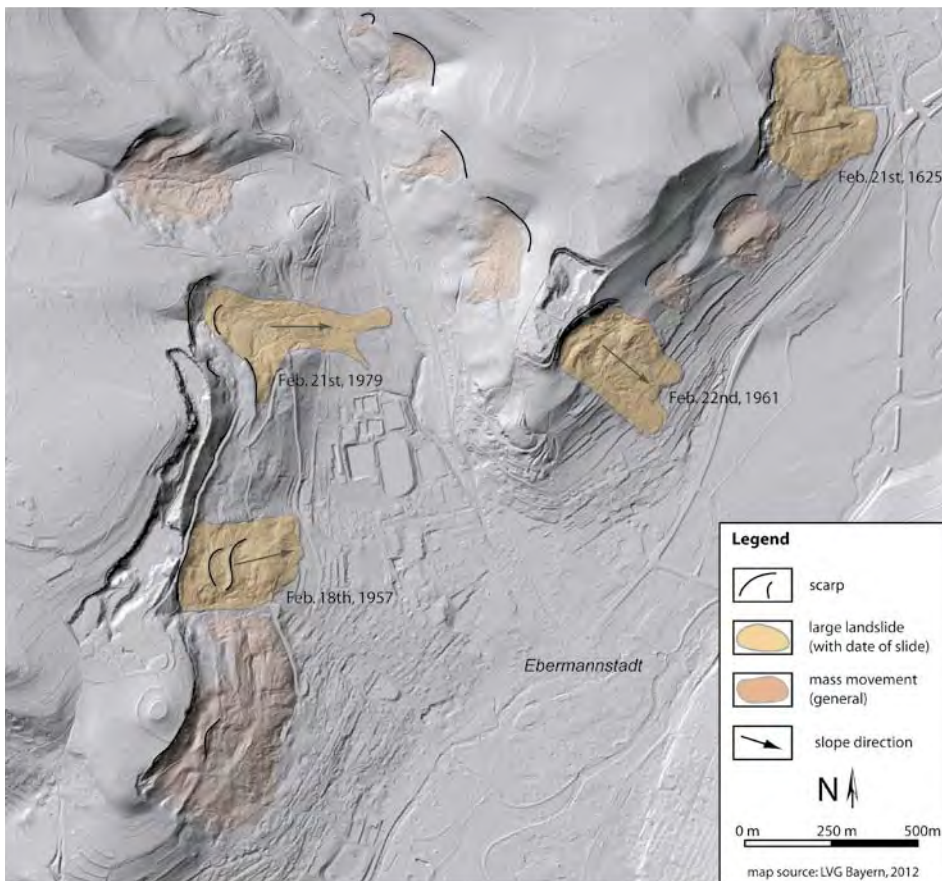


Fig. 2: Mass movements around Ebermannstadt.

Abb. 2: Massenbewegungen im Raum Ebermannstadt.

masses coming in from the north Atlantic. As a result, the wind-facing west side of the mountain ridge, where the study area is situated, receives higher precipitation than the western lowlands (Forchheim 684 mm, Ebermannstadt 898mm) (purchased data from DWD (German Meteorological Service)).

Ebermannstadt is located in the broad valley of the Wiesent River at an altitude of 292 m a.s.l. The adjacent mountains rise up to over 500 m a.s.l. (Wachknock: 517 m a.s.l.) and therefore receive a significant amount of precipitation as snowfall during winter.

The geology in the region of Ebermannstadt can be seen as exemplarily for the structure of the Franconian Alb with permeable and impermeable layers. The flat valley floors of the Wiesent river valley, as well as the lower slopes in the region consist of claystones from Lower Aalenian layers (Opalinuston). Above, Upper Aalenian sandstone (Doggersandstein) forms steeper slopes covered by layers of impervious Callovian Claystones (Ornatenton). Most parts of the area are covered by debris from the overlying Oxfordian Limestones (Malm  $\alpha / \beta$ ). This unit is built up by thick, marly limestones divided by thinner schistose layers. Uppermost parts of the slope are formed by Lower Kimmeridgian Marls and Limestones, the latter of which are mined in the municipal quarry, located above the study area. Measured over a distance of 500 m, all described layers show similar dipping of 8-10 m (= approximately  $1^\circ$ ) towards the east (MÜLLER 1957).

Due to dipping and aspect, several groundwater springs are found on the east-facing Hasenberg. The most important horizons in terms of groundwater springs are the tran-

sitions between Lower Aalenian layers and Upper Aalenian Sandstone, as well as between Callovian layers and Oxfordian Limestones. There are no karstic springs found in the vicinity of Ebermannstadt. The mainly concave slopes of the surrounding mountains are often covered by landslide masses, especially in the Eschlipp valley towards the north of Ebermannstadt and on the eastern slopes of the Wiesent valley (Fig. 2).

In 1625, 1957, 1961 and 1979, four major landslides took place around Ebermannstadt (Fig. 2). All of them occurred in the Callovian Clay layers, which underlie the Oxfordian limestones. These geological settings were already described by HÜTTEROTH (1994), who investigated mass movements at the "Lange Meile" southwest of Ebermannstadt.

The study area is situated at the eastern flank of the Hasenberg, adjacent to the western town limits, which extend onto the foot of Hasenberg. MÜLLER (1957) describes the region at Hasenberg as affected by numerous past landslides, proven by outcrops of landslide derived debris layers. According to MÜLLER (1957), lacking soil formation between single debris layers indicate rather close successions of the different mass movements.

### 3 Methods

#### 3.1 Geomorphological mapping

A detailed mapping of surface morphology and existing failures on a slope is fundamental for determining the landslide type (HUTCHINSON 1995), and for analyzing regional distribution of landslide patterns (GUZZETTI et al. 1999, 2000). Therefore, a detailed geomorphological map was cre-



ated in order to capture this spatial reference of the slopes' structure as well as movement processes and to delineate potential instable areas through localization of active scars and accumulation forms of the 1957-event.

Geomorphological mapping of the present study was based on a 1-m DTM to precisely detect and locate landslide elements. Hence, a reconstruction of processes can be concluded as well as the influence of single geological layers. GIS-derivates, for example hillshades, contour lines, surface curvature, slope gradients and aspect were used during field work and data processing.

The geomorphological mapping was accomplished on a scale of 1:3000. The focus was set on the most important landforms and geomorphological units of the 1957-landslide.

With slight modifications, the symbols for steps, slope gradients, surface and valley forms are based on the mapping keys of TERHORST & KIRSCHHAUSEN (2001), who developed a legend on mass movements. The mentioned mapping key is based on the geomorphological mapping instructions from LESER & STÄBLEIN (1978).

### 3.2 Geophysics

Geophysical methods and sites were selected based on the results of the geomorphological mapping and in order to correlate detected landforms to subsurface structures.

Electric resistivity tomography (ERT) measurements were conducted using a 72-channel multi-electrode resistivity meter from IRIS instruments (Syscal Junior Switch) using 36 electrodes per array. With respect to the aimed penetration depth and resolution, a unit electrode spacing of 3 m has been applied. The array types Wenner (robust; high signal-to noise ratio) and dipole-dipole (more prone to errors, but high sensitivity to horizontal and vertical structures) were used (e.g. TELFORD et al. 1990; BURGER et al. 2006; REYNOLDS 2011).

Three ERT-surveys were conducted, beginning below the debris deposits at 441 m a.s.l., and ending in the lower parts of the landslide mass at 381 m a.s.l. (see figure 4). To achieve best possible data coverage within the subsurface, ERT arrays were measured using a roll-along technique with an overlap distance of 15 m between arrays 1 and 2 and 18 m between arrays 2 and 3.

Tab. 1: Details of the ERT and SRT measurements on Hasenberg study site.

Tab. 1: Details der ERT und SRT Messungen im Arbeitsgebiet Hasenberg.

	ERT	SRT
# of concatenated spreads	3	3
electrode / geophone spacing [m]	3	3
# of electrodes / geophones per spread	36	24
spread length [m] [single spreads]	105	69
overlap distance [m]	15 m / 18 m	3 m / 6 m
overall spread length [m]	282 m	246 m
altitude [m a.s.l.] [begin/end of spread]	381 m / 441 m a.s.l.	391 m / 427 m a.s.l.
# of stacks	Wenner, dipole-dipole	25 / 75
array type / # of shot points [per array/overall]		10
sample interval [ms]		0.25
record length [ms]		128

Datasets were concatenated (LOKE 2010) and inverted within the software package RES2DINV (GEOTOMO) using the robust inversion scheme (L1-norm). To reduce side-block effects an extended model (additional model blocks at both sides and beneath the original data array) was used to calculate resistivity models (LOKE 2010). Areas within the tomograms that are not covered with data points are shaded.

Seismic refraction tomography (SRT)-measurements were conducted using a 24 channel seismograph (Geode) from Geometrics with 24 geophones. A unit electrode spacing of 3 m has been applied, stating a trade-off between aimed penetration depth and resolution of the subsurface. A 5kg sledgehammer was used to generate the seismic signal. To achieve high data coverage, and in consideration of an expected distinct subsurface heterogeneity, shot locations were situated between each geophone pair, as well as in front of and behind the spreads (remote sources), resulting in 25 shot locations per array. As for ERT-measurements, a roll-along technique with an overlap distance of 3 m between spreads 1 and 2 and 6 m between spreads 2 and 3 has been applied. More details on the setups of ERT- and SRT-measurements are given in Table 1.

For data processing and inversion the software package SEISIMAGER/2D (GEOMETRICS) with the modules PICKWIN (picking of first arrivals) and PLOTREFA (concatenation of datasets; tomographic inversion) was used. Areas in the tomograms that are not covered by raytraces are shaded to prevent over-interpretations.

The topography of each ERT and SRT survey line has been recorded as the estimated relative vertical distance between two neighbouring electrodes/geophones, to enable an incorporation of relief data into the inversion process within RES2DINV and PLOTREFA, respectively.

### 3.3 Substrate analyses

Along geophysical surveys, additional analyses of substrates were realized to validate and improve the obtained data from geoelectrical and seismic measurements.

A total of 38 study sites in form of small pits of 20–40 cm in depth were dug over a distance of 380 m. Texture was determined by finger testing according to the Ger-

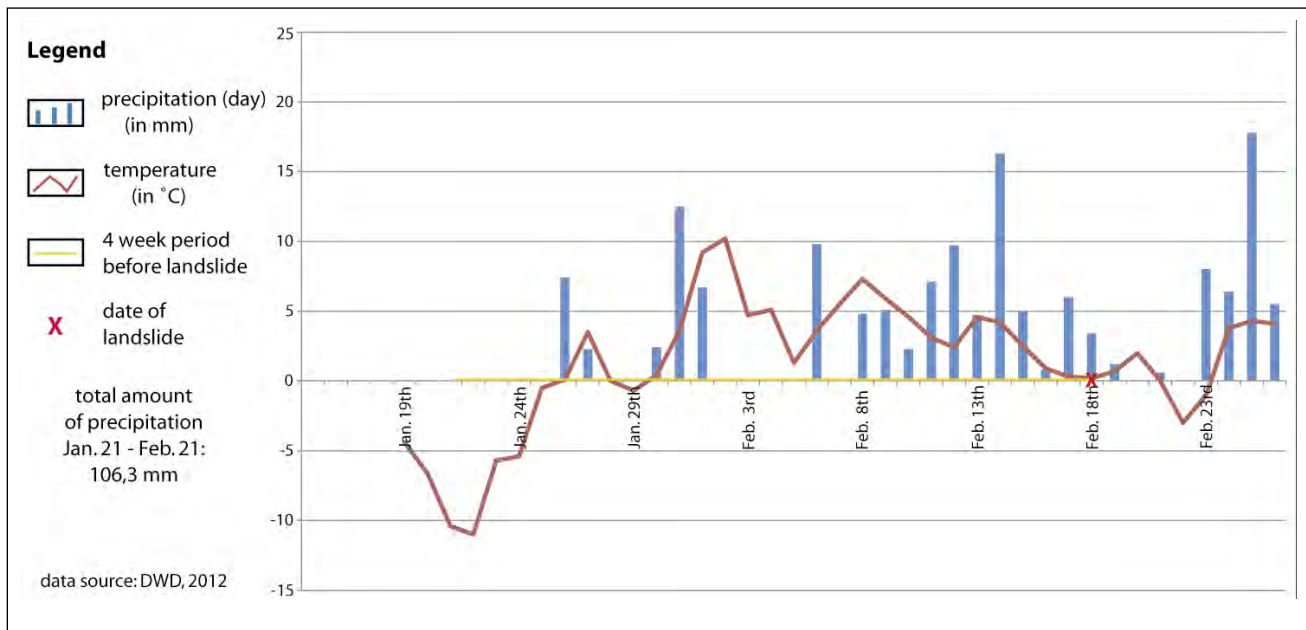


Fig. 3: Precipitation and temperature trend before the landslide of 1957.

Abb. 3: Niederschlags- und Temperaturverlauf vor der Rutschung 1957.

man Field Book for Soil Survey (AD-HOC-ARBEITSGRUPPE BODEN 2005).

One focus was set on the occurrence of limestone debris skeleton, which may indicate a slide mass for the middle and lower slope areas. Furthermore, it was important to quantify the amount of fine grained, i.e. clayey material.

#### 4 History of the landslide at Hasenberg 1957

The studied landslide at the slope of Hasenberg occurred in late winter of 1957. According to descriptions of MÜLLER (1957), slope failure became apparent on February 17<sup>th</sup>, when several cracks opened in the mid slope area of the Callovian Clays covered by Oxfordian Limestone debris. These processes continued for at least twelve hours and initiated further slide movements uphill and flow processes in the lower slope areas. Overall, the landslide covered an area of 225 m (width) x 350 m (length) with an estimated volume of approximately 625.000 m<sup>3</sup>. Average velocity during the major movements was approx. 10 cm per minute, with local maxima up to 1 m per minute (MÜLLER 1957).

The landslide mass mainly comprises Oxfordian Limestone debris mixed with clay material from Callovian layers. Most parts of the slide mass did not contain fresh limestone fragments but remobilized debris from older slide masses. Furthermore, in the lower parts of the accumulation, upper Aalenian Sandstone fragments were intermingled into the slide mass (MÜLLER 1957). The upper parts of the landslide area in the Callovian layers were dominated by cracks and slide surfaces, while tongues with an irregular, hummocky surface were formed in the lower areas, underlain by Aalenian layers. Vegetation remained intact on most parts of the landslide, except of the foot. Here, it was destroyed due to the more turbulent flow pattern and the steep front (MÜLLER 1957).

According to MÜLLER (1957), the landslide in 1957 was presumably triggered by an exceptionally high amount

of water input. Meteorological data for late January and February 1957 (see Fig. 3), showed several days of (heavy) rainfalls, summing up to an amount of 107 mm for the period of 4 weeks before the landslide event. Compared to further datasets between 1931 and 2012, this sum is definitely above the long-term average (65.9 mm).

On the other hand, precipitations in January of 1957 (77.8 mm) and December, 1956 (55.4 mm) were below average. Furthermore, some years (e.g., 1946 and 1970) show even higher precipitations in February which did not cause landslides in the study area.

Temperature records of early 1957 display values increasing to degrees around zero and above, four weeks before the landslide event. This is supposed to have caused intensive snow melting processes in that time span which lead to a significant increase of the soil moisture (the presence of a snow cover was reported by contemporary witnesses [see MÜLLER 1957]). Due to this additional water input, the actual water intake at the Hasenberg slope was much higher than precipitation data proposes. This example illustrates that soil moisture is influenced by different factors (e.g. type of precipitation, snow cover, snow melting, subsurface freezing and others) over a varying time span. Due to its complexity, this factor cannot be discussed in detail in this paper.

## 5 Results

### 5.1 Geomorphological Map

The first geomorphological map of the slide area after the event of 1957 was published by MÜLLER (1957). The map is more of a rough sketch which records the observed processes (e.g., the formation of shear planes), basic geomorphological information (e.g. land surface forms, fissures, scarps) as well as geological layers. The linking of morphological forms to slide processes was limited to a minimal extent.

Figure 4 shows a detailed geomorphological map of the landslide area based on high resolution DTM. Moreover, a colorized, GIS-processed hillshade combined with slope gradient information provides a precise visualization of the landforms in the study area. The hillshade accurately localizes the geomorphological forms without the problem of GPS-offsets.

The uppermost slope sections are dominated by thick debris deposits. Geologically, this area represents the transition from Oxfordian Limestones (above) to Callovian Clays (below). The area is completely covered by limestone debris and blocks with diameters from a few centimeters to approximately 1 m.

North of the footpath, a large limestone block has been detached from the wall. The rock formation of the block is still intact and can be directly linked to the adjacent rock-wall behind. This indicates a relatively constant and slow drifting, without major destruction probably moving on underlying clay layers.

In the upper area of the Callovian Clay (approx. 425–435 m a.s.l.), slopes are characterized by lower inclinations. Although initial soil formation is present in form of thin litter layers, the area is still dominated by limestone debris. Numerous parallel fissures (orientated rectangular to the slope) are present primarily in the center of the slide area. Most fissures provide widths of approximately 2 m, depths of 1–2 m and lengths between 5 and 25 m. The proper depth could not be measured exactly as all fissures are partially filled with soil material, debris and/or vegetation. Apart from the widespread mixed forest, vegetation in these areas shows ferns and mosses, which is an indicator for in-

creased moisture, probably due to the clayey conditions.

Slide blocks become apparent in lower parts of the Callovian Clay (approx. 405–425 m a.s.l.). They are characterized by converse uphill slopes, with heights of 1–3 m and oversteepened downhill slopes. Small areas with a hummocky surface are present in front of larger blocks, due to secondary movements in form of creeping.

The largest block is situated in the center of the landslide area (approx. 410 m a.s.l.). It forms a significant ridge with a height of 3 m (view from uphill). The front slope represents a main scarp of the 1957 landslide (total height 12 m), with a ruptured surface, indicating intensified movements here. While the upper sections of the front slope are covered by a comparably thin layer of debris, lower areas are buried under large sediment masses. Due to the curved shape of the scarp, a rotational slide can be assumed.

The region downslope (east) of the forest road (approx. at 400 m a.s.l.) represents the accumulation area of 1957. Most parts of this area are dominated by a wavy pattern, forming relatively smooth surfaces. In the central part a significant “V-shaped” valley, with a maximum depth of 3 m, cuts through the rather flat slide mass. At its lower extensions, remnants of an old pavement become visible, identifying the valley as a former defile (see MÜLLER 1957). Remnants of the undisturbed trail allow the reconstruction of its uphill course. It also identifies stagnant and displaced landslide segments: Gaps between single trail segments reveal rates of deposition between 20 to 30 m.

Lower sections of the slide area (around and below the second forest trail in the east, approx. 390m a.s.l.) are again characterized by a wavy surface. At two minor rims, slope

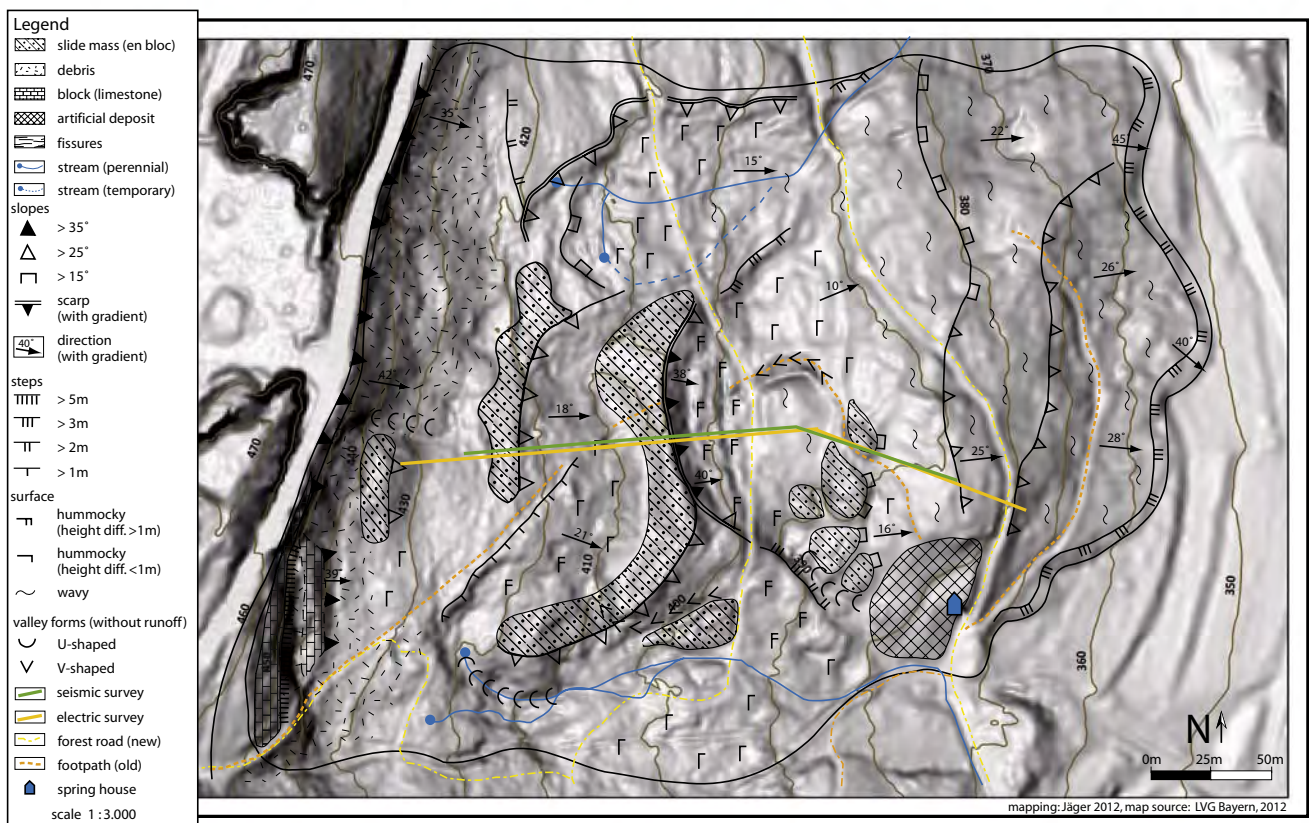


Fig. 4: Geomorph map of the landslide at Hasenberg. The background map is showing slope gradients.

Abb. 4: Geomorphologische Karte der Rutschung am Hasenberg. Die Hintergrundkarte zeigt die Hangneigungen.

gradients increase about 5°–8°. The toe of the slide mass is formed by a steep step (max. slope gradients 45°) with a maximum height between 3–4 m. At this clear boundary the landslide mass can easily be differentiated against the undisturbed surroundings. In the northwestern part, the tip is only 5 meters west of a new road, which has been built after the landslide of 1957.

## 5.2 Substrate Analyses

Figure 5a outlines the changes of substrate at the surface. The profile is coherent with the geoelectric surveys but extends to the slide's boundaries. In total, substrates were analyzed over a length of 380 m.

Substrate in the steep uphill region is clearly dominated by limestone debris and blocks with maximum diameters of 80–100 cm and an average of around 20–50 cm. Deposition on the surface is loose due to slope parallel orientation of the fragments' x-axis, with fine-grained material only partially filling the interspaces. Below the depth of 20 cm, initial fillings tend to stabilize the deposit; however, several unfilled cavities are still left. The amount of large debris fragments and blocks decreases downslope in favor of finer material.

The section between the upper ridge and main scarp (horizontal length 100–160 m) offers varying amounts of

fine grained material and debris/coarse grains. A composition of limestone fragments embedded in a dense/loamy fine grained matrix characterizes the ridges (such as the upper ridge or main scarp). In contrast, straight slope sections are dominated by clayey material with small amount of coarse grains. Mostly behind (upslope) the ridges, shallow organic horizons (5–10 cm) could accumulate.

Below the main scarp, a composition dominated by a fine grained, mostly clayey matrix was found. The amounts of limestone debris and coarse grains average approximately 50 %. In two small sectors, the amount of clay and fine grains increase to rates of approximately 70–80 %. In any case, clayey material is less dense and slightly less moist compared to the area above the main scarp.

In summary, three different substrate areas could be determined: The uppermost area is dominated by debris and coarse grains while clayey material dominates the mid-slope sections of the Callovian layers. In the lower parts of the study area, the slide mass contains a composition of both materials.

## 5.3 Geophysics

Figure 5 b & c presents the electrical resistivity- and seismic refraction tomograms in comparison to substrate anal-

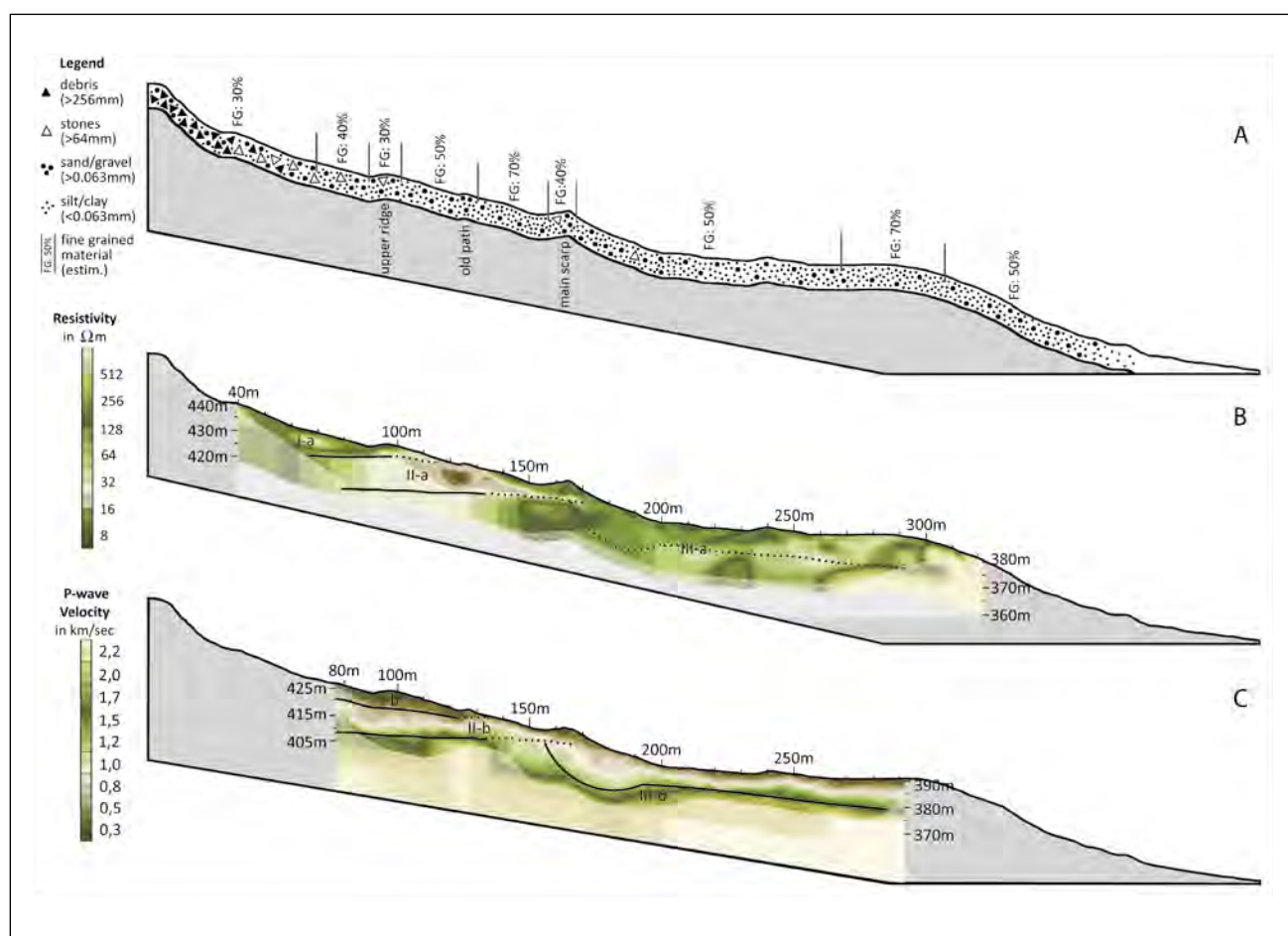


Fig. 5 a, b, c: Profiles showing results drawn from substrate analyses (a), geoelectric surveys (b) and seismic surveys (c). See Fig. 4 for location. The shaded areas could not be interpreted as they are not sufficiently covered by data.

Abb. 5 a, b, c: Profilschnitte mit den Ergebnissen aus Substratanalyse (a), geoelektrischen Sondierungen (b) und seismischen Sondierungen (c). Lagebeschreibung in Abbildung 4. Schattierte Flächen sind nicht ausreichend mit Daten aufgelöst und daher nicht interpretiert.

yses. Results from ERT and SRT allow for defining three differentiated subsurface sections (see markings in the tomograms) that will be analyzed in the following section.

The subsurface resistivity distribution is recorded over a total profile length of 282 m. The presented tomogram exhibits three major units with varying resistivity characteristics. High resistivities ( $\geq 200 \Omega\text{m}$ ) were obtained in the uppermost region (I-a) and in single spots downslope. Adjacent to this area a unit of very low values (8 - 32  $\Omega\text{m}$ ) was measured in the slope section above the main scarp (II-a). The lower half of the profile (III-a) is characterized by mostly intermediate values around 64–128  $\Omega\text{m}$ , with several lenses of higher or lower resistivities.

In the uppermost sections (I-a), a relatively homogeneous area with high resistivities ( $>200 \Omega\text{m}$ ) is detected, reaching from the surface as deep as 10 m (deeper layers cannot be taken into account due to survey geometry). Tapering out downslope, the body covers the top of the small ridge (at 225 m survey length), before it disappears completely. The transition between sections I-a and II-a is characterized by a sharp boundary and a strong contrast in resistivity values.

Resistivity values in unit II-a are mainly between 24–32  $\Omega\text{m}$ , primarily detected for rather homogeneous sections below a depth of 9 m. Within the shallow subsurface (0–9 m depth) an anomaly with distinct lower resistivity values is detected. Most probably this structure can be ascribed to higher moisture contents there rather than to differing substrate properties. This interpretation is confirmed by substrate analyses that show increased amounts of clayey material. In a depth of approx. 5 m, this unit extends close to the main scarp downslope (horizontal distances 130–160 m). At the surface (0–5 m depth) the unit is covered by a layer of higher resistivity values (max. 100  $\Omega\text{m}$ )

From the main scarp to the lower extensions of the survey a heterogeneous resistivity distribution is apparent. The section records an intensively disturbed composition as deep as 15 m. The unit (III-a) can be subdivided into the steep slope section below the main scarp (horizontal distances 165–200 m) with resistivity values between 64–128  $\Omega\text{m}$  and the gently inclined section between horizontal distances 200–300 m. The resistivity distribution in the latter section indicates a vertical differentiation of the subsurface into three layers, with a conductive layer (5–1.5m depth, 24–32  $\Omega\text{m}$ ) embedded between two layers with higher resistivity values ( $>64 \Omega\text{m}$ ). The conductive layer is horizontally differentiated into three lenses, which are cropping out to the surface between horizontal distances 260–290 m. Upslope (horizontal distances 260–200 m) this layer is superimposed by a heterogeneous layer of up to 5 m in thickness with rather high resistivity values between 64–128  $\Omega\text{m}$ . The area below 15m depth is characterized by a homogeneous resistivity distribution. However, data coverage at this depth is rather poor. In general, the transitions between all parts are relatively smooth.

Seismic surveys cover a total profile length of 215 m. Corresponding to ERT data, three major units can be identified over the course of the profile: (I-b) very low velocities between 0.3–0.5 km/sec, found at the surface and as deep as 4 m in the upper regions, as well as at the ridge of the main scarp. This unit is underlain by a section with medium ve-

locities (II-b) constantly increasing from 0.5 to 1.2 km/sec, which also tends to reach the surface in some areas. Unit III-b is characterized by rather fast velocities between 1.2 and 1.7 km/sec. Acceleration within this unit mostly occurs over short vertical distances.

In the uppermost section, the unit of low velocities (I-b) is dominant at the surface. Below the small upper ridge, velocity slowly increases to 1.1 km/sec. The only refractor is detected in a depth of 15 m, with velocities rapidly increasing from 1.2 to 1.6 km/sec on a vertical distance of approximately 5 m.

The step of the old footpath (horizontal distance 195 m) limits the downslope extent of the low velocities at the surface. Here, velocities  $> 0.9$  km/sec were obtained at the surface (unit II-b), constantly increasing with depth. Below a depth of 10 m, values increase more rapidly from 1.2 to max. 2.2 km/sec.

This pattern significantly differs from the situation underneath the main scarp. Again, low velocities were found in the surface near layers, similar to the small upper ridge. With depth, they moderately increase forming a pocket of medium resistivities (0.5–0.9 km/sec) in the steep area downslope of the main scarp. The pocket is underlain by a refractor as deep as 20 m forcing velocities to rapidly increase from 1.2 km/sec to 1.8 km/sec.

The gently inclined section between horizontal profile lengths of 200–290 m presents all three units arranged in a uniform subsurface set-up. From the surface as deep as 3 m, low velocities (0.3 km/sec) are detected. Velocities constantly increase over a vertical distance of 15 m to max. 1.8 km/sec. A refractor can be presumed in a depth of 18 m due to further velocity acceleration ( $> 2.0$  km/sec).

## 6 Interpretation and discussion

The region around Ebermannstadt provides three factors favoring landslides: (I) a bipartite geologic structure with permeable and impermeable layers, (II) increased precipitation resulting from conditions at the windward site of the Franconian Alb and (III) dipping towards the direction of the hillslope (in case of east-facing slopes). This leads to the occurrence of several mass movements in the area. As geophysical surveys show, all processes are linked closely to local geology, indicating an important influence of this slope internal factor. The major part of the slope movements occur on the boundary between (Callovian) Clay and debris cover. Obviously, slope parallel dipping of the geological layers additionally favors the development of slip surfaces. As a result, east facing slopes are affected by numerous landslides (MÜLLER 1957) as described by DORN (1928).

### 6.1 Subsurface structure and formation

In order to create a generalized model of the 1957 landslide at Hasenberg, geophysical surveys and supporting substrate analyses are combined.

The uppermost slope sections are characterized by debris-rich deposits on the surface. As substrate analyses reveal, a loose accumulation is prevalent. Parts of the interspaces are filled with fine-grained (weathered) mate-

rial. High resistivity data (unit I-a) and low velocity values (unit I-b) indicate the presence of debris (KNÖDEL et al. 2003, REYNOLDS 2011) not only at the surface, but at least as deep as 10 m. Fissures (see Fig. 4) cause reduced interspace-fillings between horizontal distances of 60 - 80 m, which is represented by a lens of increased resistivities in this segment.

From top of the survey downslope, the amount and size of limestone blocks and fragments by trend decreases. However, the upper ridge contains a higher amount of debris, as substrate analyses reveal. Again, this is supported by resistivity data and seismic velocities. Both surveys lead to the assumption of comparably solid, debris dominated cover (TELFORD 1990, KNÖDEL et al. 2003, REYNOLDS 2011), approximately 3–5 m in thickness, building up the ridge. This layer is underlain by unit II-a and II-b values.

Below the ridge (horizontal length 95–125 m) in downslope direction, a denser (packed) and/or moister subsurface is present, due to reduced debris and increased amounts of fine grained material. Seismic velocities at the surface are still low (unit I-b: 0.3–0.6 km/sec) as deep as 4 m. Below, they constantly increase to unit II-b values. The ERT displays considerably lower resistivities, compared to the debris dominated areas upslope. Both, the SRT and ERT values, can be interpreted as a compound of fine grained (clayey) material with a large amount of limestone debris (> 50%) included (KNÖDEL et al. 2003, BICHLER 2004, REYNOLDS 2011). Moist conditions, indicated by mosses and ferns and also obtained in substrate analyses, cause an additional decrease of resistivities in this area.

Between the small forest trail and the main ridge (horizontal lengths of 130–150 m), substrate analyses revealed considerably higher amounts of clayey material close to the surface. From uphill directions, the course of the old footpath is interrupted in this area, as a gap of approximately 20 m in length was detected during geomorphological mapping (see Fig. 4 for details). The path can be rediscovered at the main ridge, indicating a displacement in this area. With this in mind, a clayey subsurface appropriate to the formation of slide surfaces is expected. The intermediate surface velocities (0.5–0.9 km/sec) exhibited in the SRT clearly support this assumption. Resistivity values between 64–128  $\Omega\text{m}$  (unit III-a) indicate comparably dry composition of clayey material (TELFORD 1990, KNÖDEL et al. 2003, REYNOLDS 2011) and limestone fragments adjacent to the ridge.

To summarize, for the region between 95–165 m horizontal length, both ERT and SRT assume an extension of the in situ Callovian Clay to altitudes of approximately 422 m a.s.l. to 408 m a.s.l. This corresponds with the data of the geological map, which records a thickness of 16 m (MÜLLER 1959) for this layer.

At the ridge of the main scarp, resistivity data (unit III-a) and velocity values (unit I-b) imply a composition of higher amounts of limestone fragments and less fine grains as deep as 5 m (KNÖDEL et al. 2003, REYNOLDS 2011). The block (with the main scarp) is situated at the lower boundary of Callovian Clays covering Aalenian Sandstone. Both features are assumed to be exhibited in the ERT: A stripe of low resistivities (clay) covers a unit

of high resistivities (sandstone) obtained in depths below 10 m. Due to hidden layer effects (see REYNOLDS 2011 for details), SRT is unable to detect the layer of soft clayey material. Constantly increasing velocities reveal no distinguishable refractor, which leads to the assumption of a heavily weathered and shattered structure of the Upper Aalenian Sandstone, assumed as deep as 15–20 m. These characteristics were also stated by MÜLLER (1957).

Downslope of the main scarp, a composition of coarse and fine grained (50/50 %) material forms small ridges (e.g. at the steep slope areas below the scarp and at 110 m and 80 m survey length) at the surface. Higher amounts (>50%) of fine grained, mostly clayey material are found in the gently inclined sections of the accumulation area. As described above, the obtained resistivity data displays varying values in this unit III-a. In ERT, no in situ sandstone could be obtained as deep as 15 m. Poor data coverage prevents the detection of reliable data in greater depths (TELFORD 1990, KNÖDEL et al. 2003, REYNOLDS 2011). The SRT shows very slow increasing velocities as deep as 12 m, assuming no significant changes in the fine grained slide mass substrate (KNÖDEL et al. 2003, GLADE et al. 2005). For depths of 12–19 m, curved layer structures are clearly visible in the SRT. This structure leads to the assumption of a rotational slip surface present in this area. Rapid increasing velocities in depths below 19 m mark the surface of rupture and the assumed transition to underlying sandstone.

In the gently inclined slope areas, the SRT exhibits increasing velocities over a vertical distance of 18 m. In combination with the varying resistivity data values, an unconsolidated slide mass, containing stones and debris, embedded in a matrix of fine grained material can be assumed (KNÖDEL et al. 2003, BICHLER et al. 2004, GLADE et al. 2005). Below depths of 15 m, few high resistivity units suggest the presence of fragmented bedrock (Aalenian sandstone) (KNÖDEL et al. 2003). In a corresponding location, the SRT displays no clear refractor, but constantly increasing values. Therefore, this again may be interpreted as an intensively fissured surface of the underlying bedrock (Aalenian sandstone).

## 6.2 Morphology, landslide processes and type

According to MÜLLER (1957), the landslide was initiated by movement of water saturated debris in the lower areas of the slide area. Disappearance of the abutment led to further movement of limestone debris and blocks in the upslope regions. MÜLLER (1957) defined the overall movement as an earth flow. This designation was used again by HEGENBERGER (1961), who described the landslide of 1961 at Ebermannstadt Einbühl, which was very similar to the 1957 event. MOSER & RENTSCHLER (1999) supported this assumption when they investigated types of movement in clay layers of the Franconian Alb. However, as our investigations show, the 1957 event contained heterogeneous material and displayed more processes than a single earth flow.

Detachment of a block from the bedrock took place on the uphill slopes in the southwestern part. A combination of topple (VARNES 1978, DIKAU et al. 1996) and creeping (SCHMIDT & BEYER 2001) processes occur, as the detached

block is slowly dipping forward while moving downslope on the underlying clay. RADBRUCH-HALL (1978) defines this movement as “creep of blocks over soft rocks”. The soft bottom layer also enables rotation of the blocks (BEYER & SCHMIDT 1999).

Geophysical surveys revealed a shallow translational slab slide (HUTCHINSON 1988, DIKAU et al. 1996) of debris in the clay dominated area uphill of the main scarp. Slab slides frequently occur in debris mixed with fine material (HUTCHINSON 1988). According to DIKAU et al. (1996), a shear zone close to the surface is characteristic, as well as movements in slightly weathered bedrock. Both characteristics were obtained at Hasenberg.

The undestroyed part of the footpath on the ridge of the main scarp proves an en bloc movement of this part as suggested by resistivity and seismic data, rather than involvement of buckling, wedging or heaving processes which may also occur at slab slides (HUTCHINSON 1988). The ridge of the main scarp supposably contains a core of (relatively) solid rock. This may be a relic of limestone bedrock or sintered limestone debris, which were observed during field work. Besides, their occurrence on Hasenberg was reported by MÜLLER (1957). Movement of the block was disrupted at the transition of the Callovian Clay to the underlying Aalenian Sandstones with a clayey slip surface tapering out.

Parts of the block presumably exceeded the step of the underlying Aalenian Sandstone.

At the step, the upper layers of the sandstone are shaped by intensive weathering (MÜLLER 1957) and/or prior movements (DORN 1928). Internal weaknesses lead to the formation of a curved slip surface (see SRT) and a small (debris) rotational slide (VARNES 1978, DIKAU et al. 1996). The slide also affected the underlying sandstone, as substrate analyses obtained sandstone particles in the slide mass of the main body (see also MÜLLER 1957).

Earth flow processes (MÜLLER 1957) are characteristic for the lower parts of the slide area and have to be specified as they do not comply with the definitions of VARNES (1978), HUTCHINSON (1988) or CRUDEN & VARNES (1996). In terms of involved material (“*earth*”), resistivity data and substrate analyses at Hasenberg assume a significant amount of debris embedded in a matrix of fine material. According to the authors mentioned before, the limit of coarse material in an earth flow is 20 % (amounts exceeding 20 % = debris). On the other hand, a review of flow type processes by HUNGR et al. (2001) determines clay contents of only 10–70 % for earth flows. Therefore, a distinct definition of the material properties is difficult to apply.

This is also the case in terms of movement mechanism (“*flow*”). VARNES (1978) classifies all slope movements with internal distortion involved as “*flows*”. However, HUNGR et al. (2001, p. 222) state, it is “often difficult to determine whether internal distortion or boundary sliding is dominant in a given case”. Furthermore, investigations proved that several “*flows*” moved predominantly by sliding along a shear surface, rather than an irregular flow-pattern (HUTCHINSON 1970, BRUNSDEN 1984). A coinciding characteristic for flows is a high content of involved water. According to a study of VON DER HEYDEN (2004) the accumulation of a landslide mass on permeable layers (like the Aalenian Sandstone) represents a rather stable condition,

as water is able to percolate through debris and subsurface layers. Therefore, a saturation of debris, which may lead to a flow process, is rather unlikely. However, the recent classifications of flows by HUNGR et al. (2001) reveal water contents near the plastic limit as sufficient for a slow moving earth flow.

In the case of Hasenberg, the old road may solve the problem of a designation. Although being displaced with different rates of movement, the old road mainly remained connected (MÜLLER 1957). This indicates the existence of one or more relatively coherent masses/blocks with fine grained material involved (see also ERT).

According to the presented literature, movement in the lower parts of the landslide can basically be defined as “*earth flow*” sensu HUNGR et al. (2001), with the constraint of a coherent mass instead of individually moving particles as mentioned by DIKAU et al. (1996).

Concerning the entire landslide, the term “*earth flow*” is not wrong but rather misleading as it does not reflect the complexity and disparity neither of occurring processes nor of the involved materials.

An enhancement to “*complex earth flow*” is considered to be necessary. However, as involved material mainly comprises debris and the movements represents typical slide processes (see DIKAU et al. 1996, CRUDEN & VARNES 1996) the classification of the studied landslide as “*complex debris slide*” or generalized as “*complex landslide*” (CRUDEN & VARNES 1996) seems to be most appropriate.

### 6.3 Future susceptibility

In order to stabilize the slope some adjustments were achieved after the 1957 landslide. The planting of cottonwood with its shallow roots was meant to stabilize layers close to the surface. However, movements in deeper stages, as detected by geophysical data, cannot be prevented by vegetation. A second regulation was the tapping of springs in order to prevent uncontrolled infiltration of water into the subsurface. This is one of the most important adjustments especially if impermeable layers, e.g. clay, are involved (HAMMER 1984).

Until today, no further landslide has occurred. This is in favor of the interpretation of the Hasenberg as a stabilized landslide area. However, cracks and damages in the road indicate contemporary movements.

Additionally, recent climate change leads to modified precipitation patterns. Most notably, winter precipitation rises (PARRY et al. 2007). Although reliable predictions for this most important trigger factor are not possible yet, a decreasing influence is not assumed. Ongoing investigations especially have to clarify the trend of precipitation along with effects concerning infiltration and consistency boundaries.

Considering all factors, processes and circumstances obtained in the study area are not restricted to the Hasenberg landslide – as illustrated by the event at Ebermannstadt Einbühl (1961) showing similar forms and processes. Also, slopes adjacent to the landslide area provide similar conditions. Assuming certain requirements (intensive snow melting, high precipitation rates), these facts indicate an endangered susceptible slope area.

## 7 Conclusions

The landslide on Hasenberg from 1957 combined a variation of movement processes like slab slide, rotational slide and earth flow. As a result, the whole movement should be defined as a “complex landslide” or, more detailed, as “complex debris slide”.

Slide material is a composition containing varying amounts of limestone debris, clayey material and – in the downslope areas of the study site – fragments of Aalenian Sandstone. In 1957 only little amounts of fresh debris were produced, the majority corresponded to debris which resulted from weathering and/or older slope movements. With exception of the uppermost slope sections, debris contains all grain sizes between clay, silt to stones and debris. The thickness of debris accumulation differs from a few decimeters (covering the slip surface) to more than 10 m (downslope main scarp).

The obtained processes assume a very close relation to the internal factors of local geology (bipartite structure: permeable and impermeable layers) as well as dipping (analog to exposition). External triggers are (winter) precipitation combined with additional snow melting (and therefore temperature).

After the landslide, necessary adjustments for slope stabilization were achieved. Until today, no further landslide occurred in the study area. However, no guarantee can be given, especially since roadway damage indicates certain movements. Even if the 1957 slide area should be stabilized, similar movements may occur on adjacent slopes.

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# Ice Age geomorphological Ahorn Valley and Ailsbach River terrace evolution – and its importance for the cave use possibilities by cave bears, top predators [hyenas, wolves and lions] and humans [Neanderthals, Late Palaeolithics] in the Frankonian Karst

*Case studies in the Sophie's Cave near Kirchahorn, Bavaria*

Cajus Diedrich

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## Abstract:

The Sophie's Cave in Upper Franconia, Bavaria (South Germany) eroded into Upper Jurassic reef dolomite and is a perfect model including all three stages of cave development ranging from a 1. ponor cave, to 2. intermediate periodically flooded cave to 3. dry cave. The key position of the cave along the Ahorn Valley, a side valley of the larger Wiesent River Valley, allow a cave genesis and evolution reconstruction which started in the Pliocene. The main refill took place in the Quaternary with Middle to Late Pleistocene river terrace sediments, present as relict sediments. Seven valley genesis stages between Pliocene to final Late Pleistocene can be separated in elevations of 440 to 375 meters a.s.l. The lowering of the Ailsbach River in the Ahorn Valley is important to understand the accessibility of caves for Pleistocene animals and Palaeolithic humans in different valley positions and elevations during different times in Upper Franconia, and the natural erosive opening/closing of cave entrances towards drainage valleys. The Sophie's Cave was used first in the Middle Pleistocene in elevations of 420 m. a.s.l. over small entrances only by small carnivores such as martens and later in the Late Pleistocene (412 m a.s.l.) by cave bears mainly, which denned over generations leaving hereby rich Late Pleistocene bonebeds. Hyenas also used the cave entrances as dens, similar as wolves, but over shorter periods only. The Ice Age top predators and cave bears seem to have been in competition within the fluent geomorphological change or collapse of cave entrances especially along river valleys during the late Middle (Neanderthals) to Late Paleolithic (Aurignacians-Gravettians), which humans did not occupy the present Sophie's Cave entrance. Late Palaeolithics (Gravettians, indirectly dated by reindeer antler: 25.750±130 BP and mammoth pelvic fragment: 24.150 ±130 BP) used in the final Late Pleistocene a deeper part of dry cave only as sanctuary in form of a shamanic deposition of selected mainly shed male reindeer antlers (and possibly two mammoth pelvic remains). This human caused antler accumulation was finally scattered at the end of the Ice Age (Dryas to Alleröd) and was damaged by large dropping ceiling blocks and by dropping waters under the last and main speleothem genesis period. After the disappearance of cave bears and predators within the last maximum glaciation (= LGM), and after glaciers (or larger snow fields) might have been present in some of the Upper Franconian Valleys, such as the Ahorn Valley, there was no sign of cave use by large mammals or humans. After the LGM, during the Dryas to Alleröd periods, not far in the Ahorn Valley at the Rennerfels rock shelter a settlement of similar Late Magdalénian/Epipalaeolithic age is known on the Ailsbach River terrace in elevation about 380 m a.s.l.

**Eiszeitliche geomorphologische Ahorn- und Ailsbach-Flussterrassen Entwicklung – und ihre Bedeutung für die Höhlen-nutzungs-Möglichkeiten durch Höhlenbären, Top-Prädatoren [Hyänen, Wölfe und Löwen] sowie Menschen [Neandertaler, und Spät-Paläolithikum] im Fränkischen Karst – Fallstudien in der Sophienhöhle bei Kirchahorn, Bayern**

## Kurzfassung:

Die Sophienhöhle in Oberfranken, Bayern (Süddeutschland) erodierte in die massiven Ober-Jura Riff-Dolomite und ist eine perfektes Modell inklusive aller drei Höhlengenesen-Stadien von 1. Ponor-Höhle, 2. Intermediate zeitweilig geflutete Höhle, 3. Trocken-Höhle. Die Schlüsselposition entlang des Ahorn-Tales, einem Seitental des größeren Wiesent-Flusstales, erlaubt eine genaue Höhlengenesen und Verfüllungsrekonstruktion, die bereits im Pliozän begann. Die Hauptverfüllung mit Flussterrassen-Relikt-Sedimentserien fand im Mittel- bis Spät-Pleistozän statt. Sieben Höhlen-Genese und Verfüllungs-Stadien zwischen dem Pliozän und ausgehenden Spät-Pleistozän können in den Höhlenlagen zwischen 440 to 375 ü. N.N. unterschieden werden. Die Eintiefungsschritte des Ailsbachs im Ahorn-Tal sind wichtig für das Verständnis der Zugangsmöglichkeit der talnahen Höhlen für Eiszeittiere und paläolithische Menschen in verschiedenen Tälern von Oberfranken zu unterschiedlichen Zeiten, sowie den generellen Höhleneingangs-Öffnungen und -Schließungen während der Entwässerungs-Phasen in den Tälern. Die Sophienhöhle wurde erstmals im Mittel-Pleistozän in der Höhenlage der Terrasse auf 420 m ü. N.N. über einen kleineren Eingang durch Kleinraubtiere wie Marder genutzt. Später, als die Terrasse auf 412 m ü. N.N. abgesunken war kamen kleine erste Höhlenbären-Arten über einen neuen Eingang die im Früh-/Mittel bis Spät-Pleistozän die Höhle als Horst nutzten. Diese hinterließen reichhaltige Knochenlagen. Hyänen nutzten nur gelegentlich und kurzfristig den damaligen Höhleneingang, wie auch Wölfe. Die Eiszeit-Top-Prädatoren (Hyänen, Wölfe) und Höhlenbären scheinen im steten Konflikt im Kampf um die wieder und wieder zusammenbrechenden und sich verändernden Höhleneingänge entlang der Täler gewesen zu sein, besonders im Mittel-Paläolithikum (Neandertaler) oder Spät-Paläolithikum (Aurignacien-Gravettian). Keine dieser Menschengruppen nutzte die Sophienhöhle oder deren heutigen Eingang. Spät-Paläolithiker (Aurignacien/Gravettian, indirect datiert an Rentiergeweihe: 25.750±130 BP und einem Mammut-Beckenrest: 24.150 ±130 BP) nutzten am Ende des mittle-

ren Spät-Pleistozän einen tieferen Bereich der trockenen Höhle nur als rituellen Ort in Form einer schamischen Anreicherung von überwiegend selektierten männlichen Abwurfstangen (und möglicherweise zwei Mammut-Pelvisresten). Diese durch Menschen verursachte Geweihanreicherung wurde letztendlich am Ende der eiszeit (Dryas bis Alleröd) von herab fallenden Deckenblöcken und Tropfwasser der letzten Speleogenese-Periode auseinander gerissen und verfrachtet. Nachdem Höhlenbären und Top-Prädatoren (Hyänen, Löwen) im Hochglazial-Maximum (= LGM) in der Region verschwanden, als tentative Gletscher in einigen Tälern Oberfrankens vorhanden waren, wie im Ahorntal, erschienen Spät-Magdalénian/Epipaläolithikum-Rentierjägergruppen. Ein Jagdlager wurde aus diesen Epochen am Rennerfels-Abri unweit der Sophienhöhle in einer Höhenlage um 380 Metern auf der Ailsbach-Terrasse gefunden.

**Keywords:** *Bavaria, Ice Age, cave, Ahorn Valley, Ailsbach River, terrace evolution, bears, humans, Neanderthals, Late Palaeolithics*

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## 1 Introduction

The Ahorn Valley in Upper Frankonia (Bavaria) connects to the Wiesent Valley (Fig. 1) and has the most dense cave amount in the Frankonian Alb karst whereas most of them are only small clefts or cavities (SCHABDACH 1998). The caves are eroded into the massif Upper Jurassic reef and lagoonal inter-reef dolomites and limestones (MEYER & SCHMIDT-KALER 1992, Fig. 2). The large caves of the region are the König Ludwigs Cave with its large portal including a single large chamber (entrance 385 m a.s.l.). The 1833 discovered higher elevated and herein studied (entrance 411 m a.s.l., Fig. 1) Sophie's Cave (HOLLE 1833, WAGNER 1833) is situated opposite. Other larger caves in higher elevations between 550 to 430 m a.s.l. have also Late Pleistocene Ice Age vertebrate content such as the Große Teufels Cave, Moggaster Cave, Zoolithen Cave; or Geisloch Cave. Smaller and fewer cave bear remains containing caves are the Zahnloch Cave, Neideck Cave, Wunders Cave and Esper Cave (DIEDRICH 2012a, 2013a, Fig. 1).

Already NEISCHL (1904) remarked first, that sediments in caves along Upper Frankonian river valleys are important for the landscape and glacial dewatering system reconstruction. First identifications of river terraces and their possible elevations discussed SPÖCKER (1952) for the Frankonian Pegnitz Valley, but with a coarse model only, whereas "problems of the valley genesis and dating" were reviewed by HABBE (1989). Again, only karst evolution models were presented, especially for the earlier Cretaceous to Tertiary periods (GROISS et al. 1998). A new discussion about river terraces in the valleys of Upper Frankonia appeared with the new sedimentological research of the Zoolithen Cave along the Wiesent Valley, where 140 meters above today's river elevation the entrance must have been flooded postglacially, dated by cave bear tooth morphology and stratigraphy (DIEDRICH 2011, 2013a).

In this study, the Sophie's Cave in the Ahorn Valley is presented, which allows a detailed picture of the valley and cave genesis starting in the Early Pleistocene as a model for the Upper Frankonian geomorphological change, especially with details of river terraces and elevations for the late Middle Pleistocene to final late Pleistocene.

The Upper Frankonian Karst, which formed mainly into the massif Upper Jurassic dolomites and partly also in limestones, can be reconstructed in its geomorphological change between the Plio- and Pleistocene (latter in high resolution). The only data available are cave relict river terrace sequences, which were found on entrance areas of valley-cut caves.

Here it is presented for a case study area of the Ahorn Valley between south of the village Kirchahorn and the Rabenstein Castle, especially in the larger Sophie's Cave of Upper Frankonia (Fig. 4).

Their three main sedimentary series are correlated and dated partly with megafauna remains, whereas repeating floods of the cave and resedimentation destroyed parts of older layers (Figs. 3–4). The Sophie's Cave with autochthonous cave genesis and allochthonous river terrace infills allows the reconstruction of the lowering of one of the Frankonian rivers in the Ahorn Valley starting 5 Mio years ago. The Sophie's Cave is furthermore a perfect model for Frankonian caves in dolomites with three main development stages from a 1. ponor, to 2. transitional and finally 3. dry cave, which much more details can be found in DIEDRICH (2012a-b).

## 2 Material and methods

From January to July of 2011 a first interdisciplinary cave survey of the Sophie's Cave allowed the genesis, refill (see sections Fig. 3) and use by animals and humans to be reconstructed (Fig. 4). The methodologies followed general cave genesis and morphology analyses (BRETZ 1942, JENNINGS 1985), cave sedimentology (DOGWILER & WICKS 2004, SASOWSKY & MYLROIE 2007, WHITE 2007) but also river terrace analyses (KAISER 1961). Finally the recent chronostratigraphy for the Plio-Pleistocene of GIBBARD & COHEN (2009) was used. Sections all over the Sophie's Cave in different positions and elevations were analysed in grain sizes, sediment structures and fossil contents (including reworked Jurassic fossils). Palaeomagnetic studies are still in progress. First and only relative dating was possible following "fluvial river terrace sequence stratigraphy" (BRIDGLAND et al. 2004), and the "evolution of cave bear and especially tooth morphology" (RABEDER 1999, STILLER et al. 2010), but also the "cave megafauna bone assemblages" and "taphonomy" in Late Pleistocene mountainous regions in general (e.g. DIEDRICH 2011, 2012b, 2012a) and its change, which support the climatic interpretations.

The cave was explored geologically, palaeontologically, and archaeologically in January–August 2011. The historically opened sections at different parts of the cave system were prepared and studied being now accessible and protected for further studies and popular scientific presentations within the show cave. New Middle Pleistocene marten tracks were left in-situ, such as a new excavated large bonebed field in the Reindeer Hall, and a small bonebed field in the Bear's Passage. The Late Pleistocene megafauna remains, mainly cave

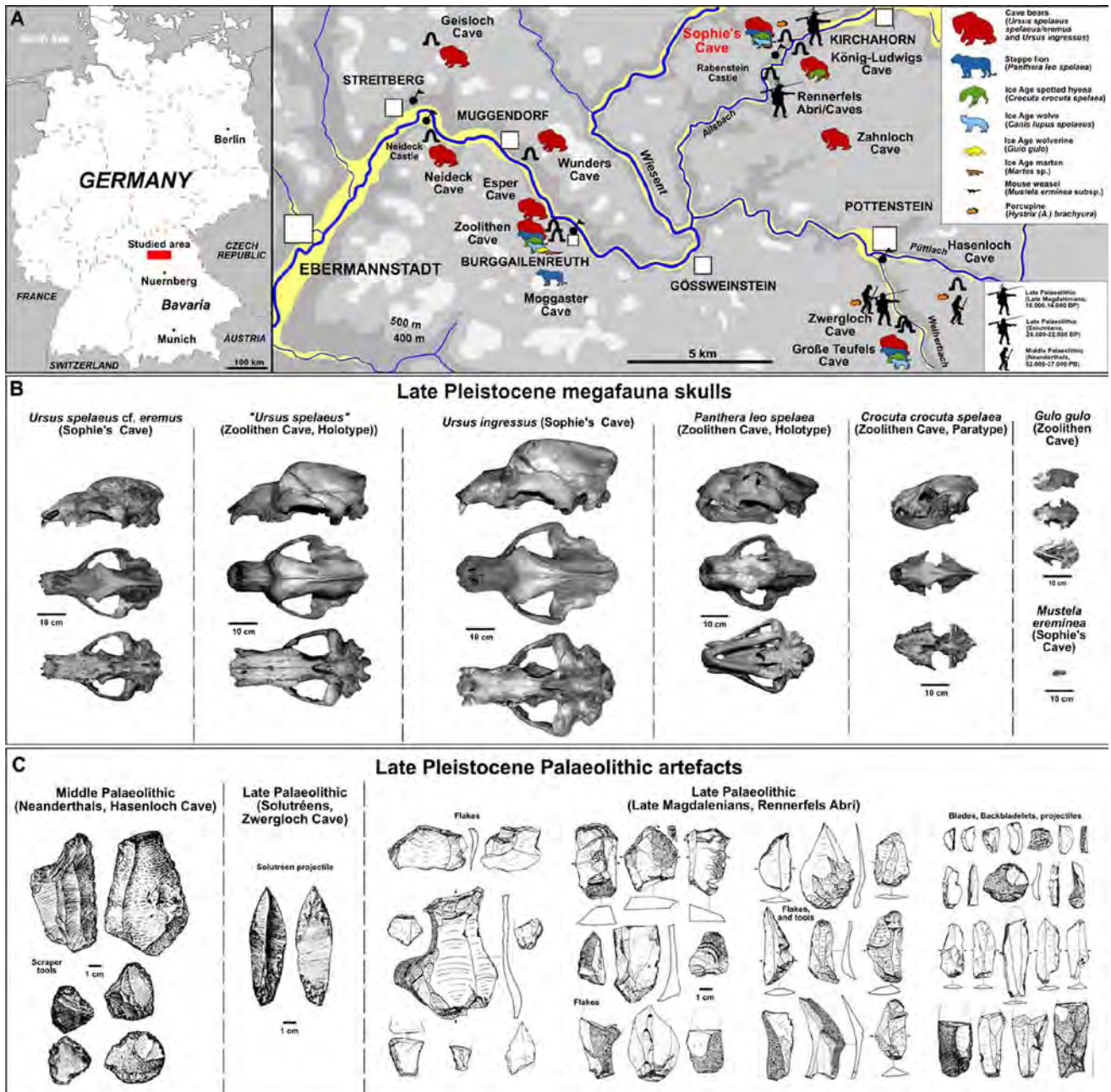


Fig. 1: A. Late Pleistocene cave bear, hyena, wolf, marten and weasel den cave sites and caves/valleys used by humans in the Middle (Neanderthals) and final Late Palaeolithic (Late Magdalénians/Epipalaeolithics). B. Selected Late Pleistocene Ice Age animal skulls (composed from DIEDRICH 2011, 2012a). C. Composed stone artifacts from the Hasenloch Cave (Middle Palaeolithic) and Rennerfels rock shelter (Late Magdalenian/ Epipalaeolithics) (selected from GUMPERT 1981).

Abb. 1: A. Spätpleistozäne Höhlenbären-, Hyänen-, Wolf-, Marder- und Mauswiesel-Höhlen-Horste und Nutzung der Höhlen/Täler durch Mittel-Paläolithiker (Neandertaler) und spätem Spät-Paläolithikum (Spät-Magdalenian/Epipalaeolithikum). B. Ausgewählte spätpleistozäne Eiszeitier-Schädel (zusammengestellt aus DIEDRICH 2011, 2012a). C. Zusammengestellte Steinartefakte von der Hasenloch-Höhle (Mittel-Paläolithikum) und Rennerfels-Abri (Spät-Magdalenian/Epipaläolithikum) (zusammengestellt aus GUMPERT 1981).

bear bones (about 98%), which were dumped historically in two areas of the Bear's Passage and the Bear's Catacombs (more than 1,600), were cleaned, inventoried and studied during this project and are housed now in the museum of the Rabenstein Castle which belongs to the cave ([www.burg-rabenstein.de](http://www.burg-rabenstein.de)). Additionally the so far incorrectly composed "cave bear skeleton" was demounted and analysed. A new small cave bear species skeleton was compiled including all small bones which is presented since 2011 in a show case in the cave lying with the isolated bones in "hibernation bed position".

Finally two C14 datings were made on a reindeer antler and mammoth pelvic (both found both below last speleothem layer in the centre of the Reindeer Hall) by the Beta Analytic Laboratory, Florida, USA ([www.radiocarbon.com](http://www.radiocarbon.com)).

### 3 Results and discussion

**A. Pliocene-Early Pleistocene:** Below a Pliocene plateau phreatic waters caused the main cave genesis in underground water level period (ponor cave stage, Fig. 5A). The first sediments were deposited fluvial by the underground

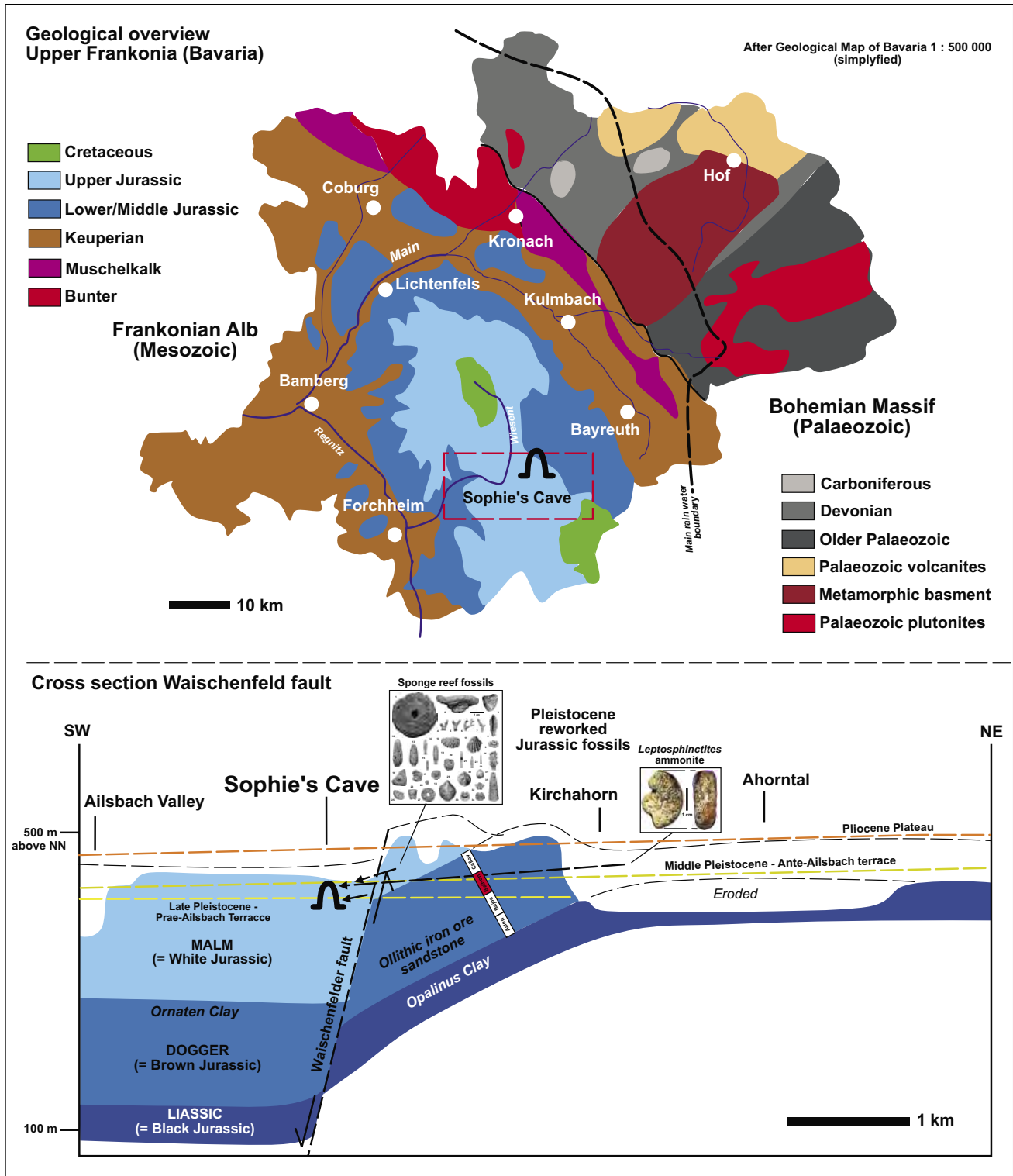


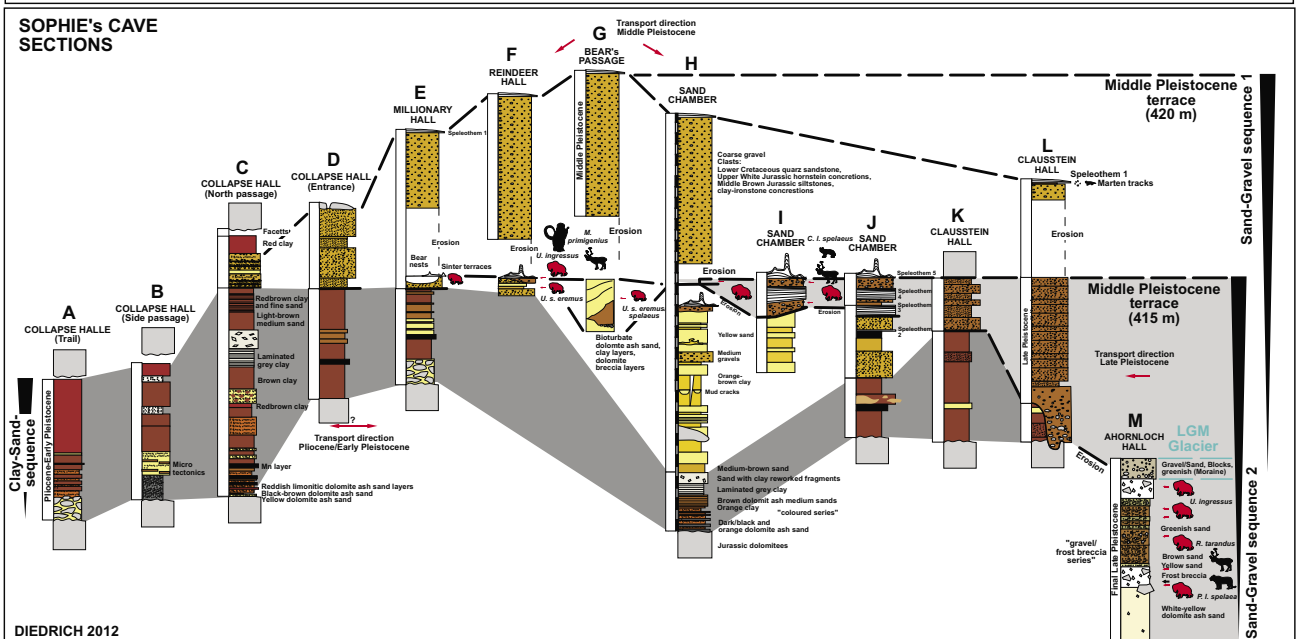
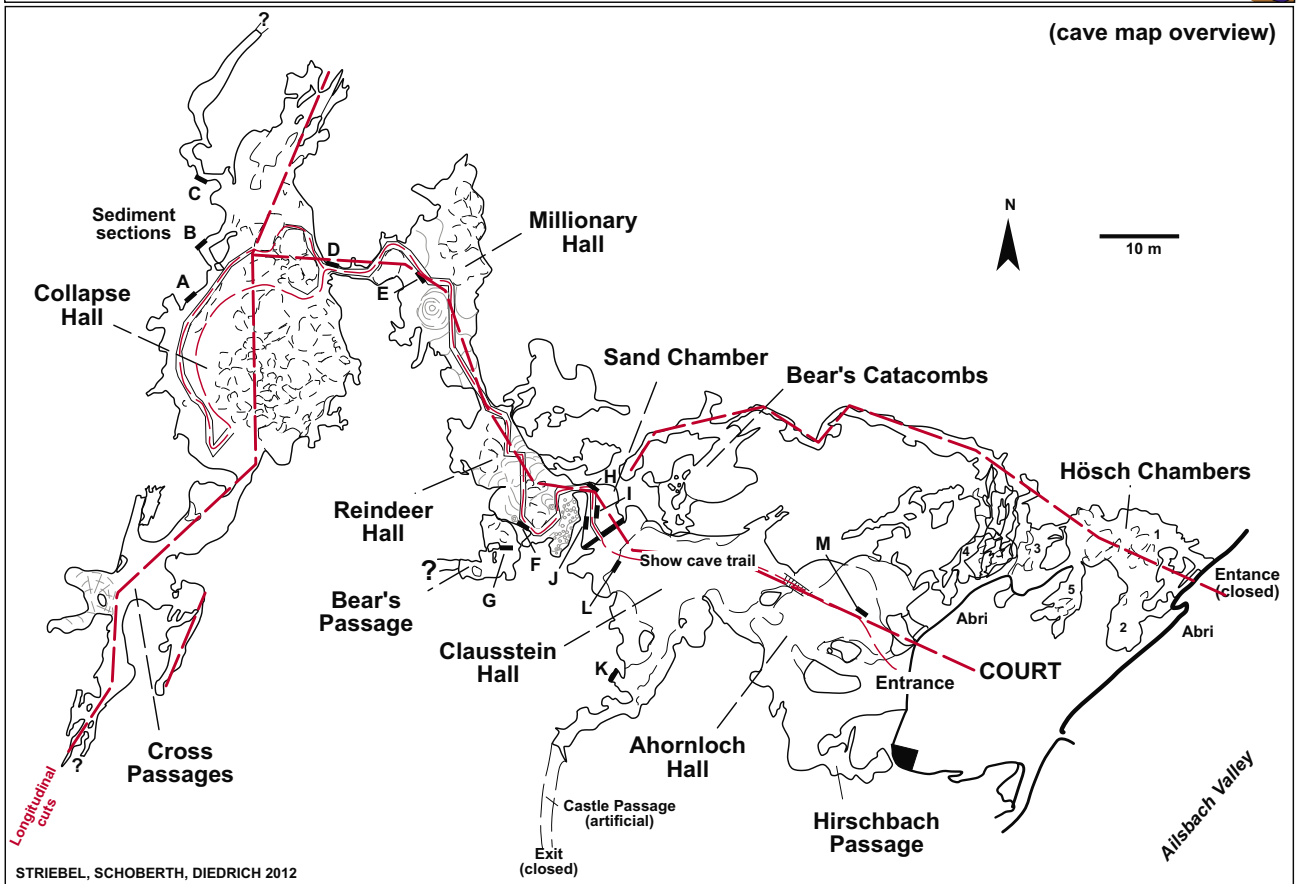
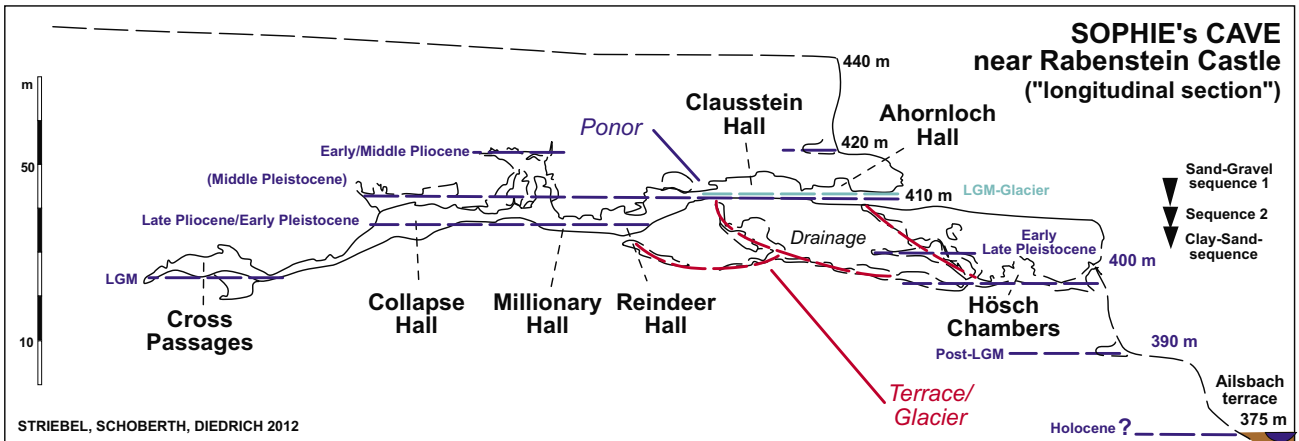
Fig. 2: Geological overview of Upper Franconia and cross section along the Kirchahorn Depression and Ahorn Valley and fault, and former elevations of the river terraces. The fossils found in the Pleistocene sediments of the Sophie's Cave are from the surrounding Brown and White Jurassic, whereas quartz pebbles are from nearby Lower Cretaceous sandstones (geology modified and simplified after MEYER & SCHMIDT-KALER 1992 and Geological Map of Bavaria 1 : 500.000).

Abb. 2: Geologische Übersicht von Oberfranken und Querschnitt entlang der Kirchahorn-Depression und Ahorntal-Störung sowie ehemaligen Flussterrassen-Höhlenlagen. Die Fossilien aus den Pleistozän-Sedimenten der Sophienhöhle stammen aus den umgebenden Braun- und Weißjura-Schichten, wobei Quarz-Kiesel von wenig weiter entfernten Unter-Kreide-Sandsteinen aus der Umgebung herrühren (Geologie verändert umgezeichnet nach MEYER & SCHMIDT-KALER 1992 und Geologische Karte von Bayern 1 : 500.000).

river in the Plio-Early Pleistocene consisting of „coloured series“ of about 1–4 meters thickness: black manganese, red iron (minerals are used for palaeomagnetic analyses) and orange limonite clays, silts and yellow dolomite ash sand layers (Fig. 4). Within the Early Pleistocene, the upper wide extended initial river valley must have developed in elevations

starting below 450 m a.s.l. (with unclear depth of erosion at end of Early Pleistocene) but terrace sediments were impossible at that time to have been washed into the not opened Sophie's Cave branches. The cave was cut by the Ante-Ailsbach River not before the Middle Pleistocene.

**B. Middle Pleistocene:** During the further valley genesis



(continuing wide valley morphology) and river terrace deepening of the Ante-Ailsbach River, parts of the cave fell dry time by time, others (valley close related parts) were flooded and filled (transitional cave stage, Fig. 5B). The oldest Ante-Ailsbach River terrace (elevation about 420 m a.s.l.) left a first larger sedimentary fluvial sequence („yellow series“, Fig. 4) of up to 8 meters thick limonitic, yellow clay/sand (high flood deposits) which end in some meters thick massive coarse gravel layer (terrace maximum high). The water did clearly not wet too deep into the cave system (preserved bonebeds and cave bear nests in the Millionary Hall). The water did not produce a classical underground river system again, most probably due to vadose changed cave morphology and vertical shafts. Waters clearly disappeared vertically in the Reindeer Hall, and the Clausstein-Halls, which have larger vertical shafts to lower elevated systems, in which most probably the water was drained (Fig. 3). Those Middle Pleistocene gravels which maximum thickness in the Reindeer Hall contain Middle/Upper Jurassic non-dolimitic rock pebbles (even Middle Jurassic ammonites: *Leptosphinctites* – Fig. 2, and many metasomatic silicatic replaced and calcitic reef fossils) but also up to 1 cm small well-rounded clear quartz pebbles, which latter were transported twice and originate from surrounding eroded Lower Cretaceous fluvial deposits. The sediments were washed over an open vertical/diagonal shaft above the Clausstein Hall into the cave, and filled up nearly half of the Reindeer Hall and most of the Millionary Hall and were finally covered in some areas by a first undated speleothem layer (first speleothem genesis period, Fig. 3, compared to other Frankonian Caves: cf. NORDHOFF 2005). Below this, in the Clausstein Hall, marten tracks on originally mud-cracked clay surfaces (preserved now as hyporeliefs below speleothem layer) (Fig. 5B) indicate the first use as a marten den of the Clausstein Hall cave part at least.

**C. Early Late Pleistocene:** The terrace lowered only few meters to an elevation of about 415 a.s.l. (Fig. 5C). In the early-middle Late Pleistocene only small cave bears used the cave over a new opened entrance (not today's entrance) even deep for hibernation purposes to protect against carnivore predators (lions, hyenas; and wolves). Many bite damaged and cracked cave bear bones, and scavenged disarticulated carcasses were found deep in the cave (Millionary Hall, Reindeer Hall and Bear's Passage). Wolf (*Canis lupus spelaeus*) and small weasel (*Mustela erminea*) remains were also found there. Porcupines (*Hystrix (A.) brachyura*) are indirectly proven by two chewed (typical 4-5 mm wide rodent bite marks) cave bear cub humeri. Many wolf excrements were found in the Bear's Passage mainly, and few also in the Reindeer Hall and Millionary Hall in which coprolites partly digested cave bear bones are cemented.

**D. Middle Late Pleistocene:** In the middle Late Pleistocene a first slight ceiling collapse and new speleogenesis (second speleothem genesis) changed the cave morphology few which resulted the blocking of the Bear's Passage to the

Reindeer Hall and in general to the deeper hibernation areas. Further 5 meters the river terrace lowered deeper (410 m a.s.l.) (Fig. 3, 5D) possibly during a warmer period (third speleothem genesis period), which finally opened the today's valley-sided large portal entrance, but closed the former one. The youngest Pre-Ailsbach River terrace started with another coarsening up terrace sequence and started to grow again indicating increasing valley terraces (?glaciers) (Fig. 4). This time, quartz-rich sands (brown to greenish colored, warmer period) do not contain clay layers and are intercalated with frost breckia layers (colder period). Large cave bear remains of *U. ingressus* date all those layers of the anterior cave area into the final Middle Late Pleistocene (32.000–24.000 BP, cf. MÜNDEL et al. 2011; Fig. 4), where large cave bear types used the today's entrance and side chambers over many generations (rich bonebeds, and many cub remains). Ice Age spotted hyenas and wolves used this cave part periodically as a den, and scavenged on the bears too. Finally climatic cold interstadial indicators are arctic fox remains from the anterior cave area (Ahornloch Hall).

**E. Final middle Late Pleistocene (Late Palaeolithic Gravettian, around 25.000 BP):** The terraces seem to have been eroded fluvial (or ?by glaciers); the ponor cave was the Hösch Chambers elevation, whereas the deposits in the Clausstein Hall increased about 2 meters in a short time of a cold period (Fig. 5E). This is the time, when Gravettian humans used the Reindeer Hall as sanctuary place only (see cave use by humans), and large *U. ingressus* cave bears the anterior cave parts.

**F. Last Glacial Maximum (= LGM, around 19.000 BP):** Floods of melting large snow field (e.g. large depression area in soft Middle Jurassic sediments around Kirchahorn) or ?glacier surfaces washed in events the side moraine/kames material (unsorted sediments with local dolomitic non-well rounded blocks in the uppermost layers into the Ahornloch Hall (see glauconitic till: Figs. 3-4) and the large cave bear *U. ingressus* bones downwards into the Bear's Catacombs and other side branches (Fig. 5E), which bones are in "non-stratified" dolomite gravel layers (semi-rounded pebbles) and are in non-rounded well conditions. The some meters higher elevated gravels in the Clausstein Hall, which replaced the bone layers, and contain itself only very few bone remains. Those sediments seem to represent LGM side moraine/kames reworked gravels, when the ?terrace increased on elevation of about 410 a.s.l. (if not transported from glacier surface waters).

**G. Final Late Pleistocene – Meiendorf, Alleröd to early Holocene:** When the snow field (or ?Valley glacier) was melting post LGM lowering from 412 m a.s.l. to the ground (?ground moraine) the former river terrace eroded from an unknown level to 385 m a.s.l. (or even deeper, Fig. 5F). Again braided rivers and its coarse gravels must have formed the steep margins of the narrow valley morphology, where no terrace sediments were left on the wall margins before (?due to the glaciers); which lateral moraine sediments (till) mainly

Fig. 3: Correlation of the sections in the Sophie's Cave, and dating of the two river terrace sequences using mainly cave bear species/subspecies (cave map according to SCHOBERTH et al. 1997, and longitudinal cuts composed by Dr. T. STRIEBEL).

Abb. 3: Korrelationen der Profile in der Sophienhöhle und Datierung der beiden Flussterrassen-Sequenzen mit Hilfe der unterschiedlichen Höhlenbären-Arten/Unterarten (Höhlenplangrundlage nach SCHOBERTH et al. 1997, Längsschnitte zusammengestellt von Dr. T. STRIEBEL).

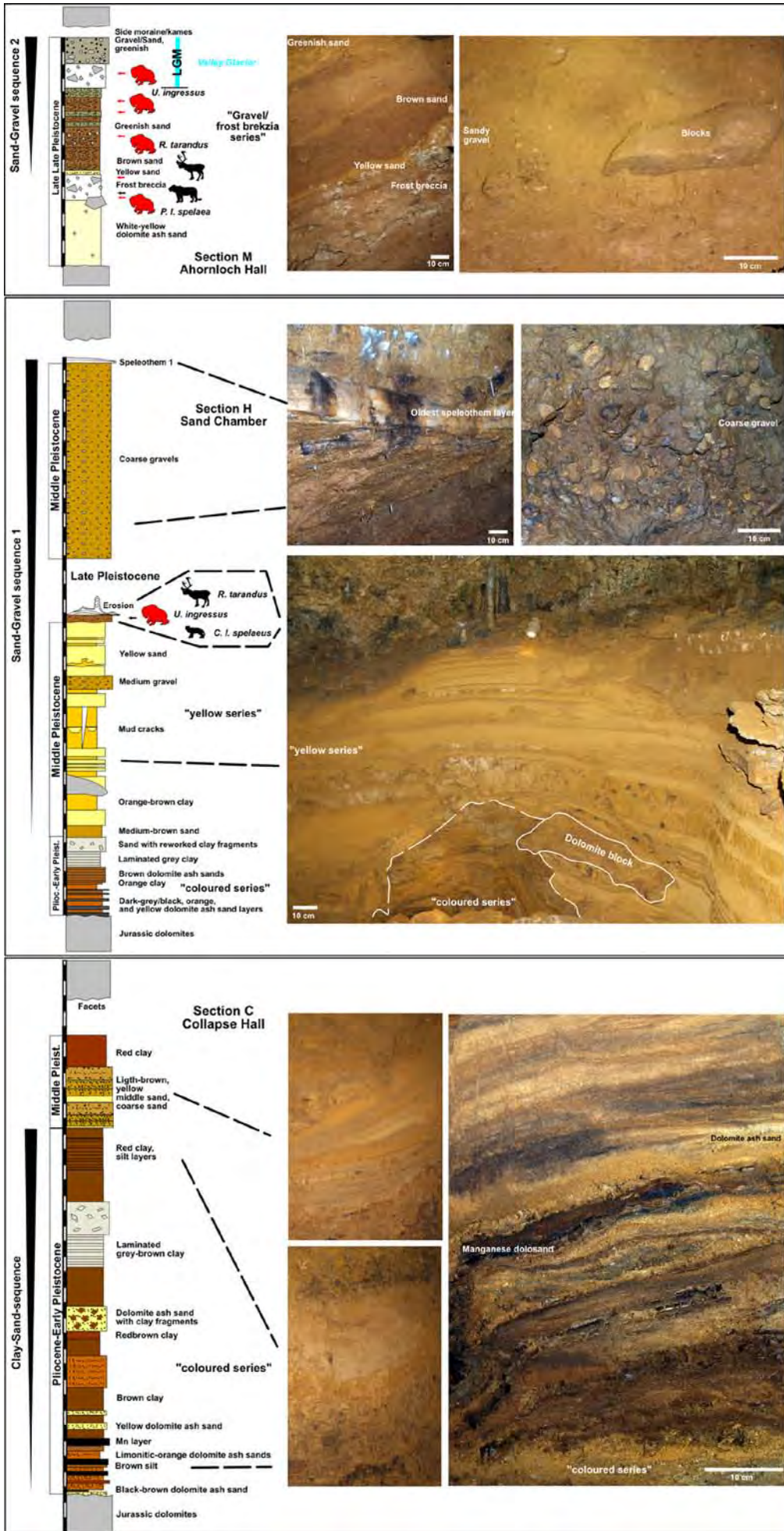


Fig. 4: Sophie's Cave main sedimentary series types and composed generalized Pliocene to Late Pleistocene section.

Abb. 4: Hauptsedimentär-Serientypen der Sophienhöhle und zusammengestelltes generalisiertes Pliozän-Spät-Pleistozän-Profil



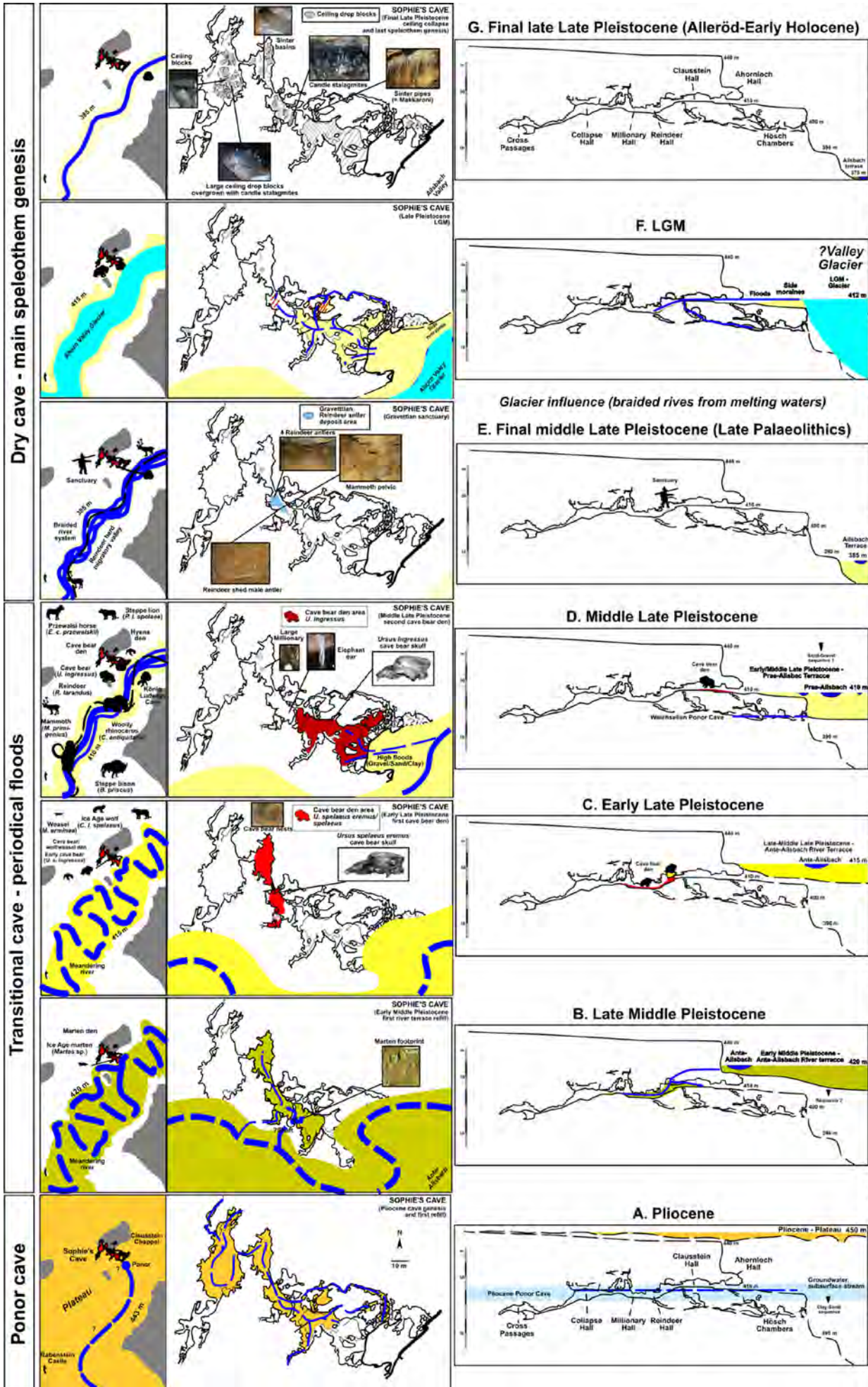
were left only in caves or cavities, such as found in the upper layers of the Sophie's Cave (Fig. 4). In the cave changing conditions caused massive event-ceiling collapse changed the cave morphology drastically, which blocks are mainly in the chambers covering their floors (Fig. 5F). Their age can be dated indirectly by two facts, first the reindeer antler remains partly are "below" those dropped blocks, whereas the last speleothem layer and candle stalagmites grew all over those large blocks, which are comparable in age to absolute dated candle stalagmites from other Frankonian caves (NORDHOFF 2005: dated around 10.513 and 10.227 cal. a B.P.). During this again humid interglacial period the terrace still lowered some meters (today 375 m a.s.l., Fig. 5G) under more humid climates, that also caused the fourth and final speleothem genesis (continuing into the Holocene). Most of the speleothems such as typical "candle stalagmites" also developed on the large ceiling drop-blocks. At this ending Ice Age time, finally the access to the Reindeer Hall over the Sand Chamber was closed by a speleothem layer (that was opened again in 1833 when the cave was "discovered" twice – first by Gravettians).

### The Ahorn Valley terrace stages and dating

The 550 meters elevated Moggaster Cave of Upper Frankonia was refilled with sediments during the Early to Middle Palaeogene (Palaeocen-Eocene: GROISS 2000). The 455 meters elevated Zoolithen Cave (close to the Moggaster Cave, Fig. 1) was refilled very few with sediments (clay layers) not before the Neogene (most probably Miocene: DIEDRICH 2013a). The Sophie's Cave in elevation of 410 meters must have been eroded and refilled correlating the cave level elevations only (south few higher as in the north, see Fig. 1) in younger Tertiary times, which Pliocene age is expected, not being confirmed by fauna or absolute dating. Typical first refills not only in the Sophie's Cave are the dolomite ash sands (yellow or dark-manganese or reddish-limonite layers = coloured series, Fig. 4), products of dolomite weathering (cf. BURGER 1989). Whereas the Zoolithen Cave represents a cave system in elevation between 460–430 meters in about three different levels (DIEDRICH 2013a) with early to maximum middle Middle Pleistocene refill, and finally latest Late Pleistocene flood event and terrace infill with medium to coarse dolomite gravels, the Sophie's Cave is more north and in lower elevation of 415–385 meters and has mainly a Middle to Late Pleistocene terrace refill (Figs. 4–5). The Ahorn Valley was eroded between the Pliocene and late Middle Pleistocene between 450–420 meters (Fig. 5). In the upper first 10 meters, the valley was wide, and started about 440 meters to become narrower. In the late Middle Pleistocene, the Ante-Ailsbach was already one third deep eroded of the today's Ahorn Valley elevation (375 m) on 420 meters. Possibly in this stage and first speleothem genesis time that covered the first terrace sedimentary sequence in the Sophie's Cave correlates more or less with the first speleothem genesis in the Zoolithen Cave (cf. DIEDRICH 2013a). There are few speleothem data available from the Franconian Alb caves, which can be used for correlations of the Sophie's Cave speleothem ages preliminary. Oldest speleothem data delivered the Zoolithen Cave 342.050 ± a 71.400 BP (Middle Pleistocene, Holsteinian interglacial, MIS 9) with its largest and oldest speleothems (KEMPE et al. 2002). Those speleothems in the Zoolithen Cave

developed on an elevation of about 445–435 meters in some parts of the cave system only (DIEDRICH 2013a). At Hunas Cave ruin the first and older speleothem is represented by a dated flowstone sample (HU-i) bearing a single age of 270.593 ± 24379 BP (Middle Pleistocene, Middle Saalian, MIS 7, NORDHOFF 2005). The younger Hunas Cave ruin stalagmite has been dated between 79.373 ± 8237 and 76.872 ± 9686 BP (Late Pleistocene, MIS 5d) during the warm interstadial around 79 ka BP (NORDHOFF 2005). This seems to be the time period, where the larger older speleothems (Large and Small Millionary) of the Sophie's Cave also started to grow. The Mühlbach Cave candle stalagmite reveal some analogy to the candle stalagmites of the Sophie's Cave including the transitions from the post-LGM (Boelling/Alleroed Interstadial and Younger Dryas cold phase) to the Early Holocene between 10.513 and 10.227 BP (NORDHOFF 2005). During the early/middle Late Pleistocene small cave bears (*Ursus spelaeus eremus/spelaeus*) date an unknown river terrace elevation at that time below 410 meters, and indicate dry cave conditions (no more flooded). The final Late Pleistocene and LGM has at least two deposit levels or series in the cave on elevations of 410 and 412 meters, dated by *Ursus ingressus* cave bear teeth which appeared in general in central Europe between 32.000–24.000 BP (STILLER et al. 2010). The Pre-Ailsbach River terrace grew first slowly and finally postglacially after the LGM at least two meters. The post-LGM replacement of bonebeds in the Sophie's Cave is similar to the replacements of bone layers in the Zoolithen Cave or Große Teufels Cave, which is also dated into the post-LGM (DIEDRICH 2013a). Although so far there is no geomorphological and for hard rock areas (here those are soft sedimentary rocks) typical evidence of a (valley) glaciation in the Franconian Alb during the LGM a minor glaciation of some elevated valleys in the study area could be a matter of debate. In this case sedimentary remnants of the glaciers have been eroded or remained undetected so far in the valleys but may have been preserved in caves (see Fig. 5 F and Diedrich 2013b). After 19.000 BP, the terraces must have eroded (probably even by ?glaciers) in some Frankonian valleys (Wiesent Valley, Ahorn Valley) whereas massive melting waters (which originate from larger snow fields or ?glaciers) replaced the "bonebed layers" of the anterior cave areas and mixed them with non-rounded and reworked tilt material. A questionable glaciation in 500 a.s.l. elevations and valley glacier development would have had massive impact of the Palaeolithic settlement of Upper Frankonia, because the natural quick landscape change explain the absence of Late Palaeolithics such as Late Solutréans to Middle Magdalénians (= around LGM time). This is similar as recently described for the Harz Mountain Range and its caves in northern Germany (DIEDRICH 2013b).

Finally, there is the discussion about the extreme high elevation of the Zoolithen Cave gravel relicts of LGM times compared to the today's Wiesent elevation (140 meters deeper), which dolomite gravels are found also in other caves around Muggendorf in elevations between 455–435 meters (DIEDRICH 2013a) which can be explained at the moment only by the presence of large snow fields in depressions or ?glaciers (ice caps or filled depressions) at least on the highest Frankonian areas and branching valleys (over 500 a.s.l.), especially around Muggendorf (DIEDRICH 2013a). The directions of the postglacial drainage (or LGM-valley gla-



riers) reconstructions is still vague and a first model (DIEDRICH 2013a) under further study. This theme has to be discussed much further (e.g. at Große Teufels Cave), but much more cave relict sediments have to be studied and interdisciplinary cave research in Upper Franconia are of need. However, those LGM sediments in the Zoolithen Cave or Sophie's Cave such as Große Teufels Cave seem to be re-worked moraine tilt sediments (including glauconite or loess sands). Glaciers would have been over 80 meters thick in the Muggendorf Wiesent Valley region (after present estimations of the Zoolithen Cave sediments: DIEDRICH 2013a), but only 30–40 meters in the Ahorn Valley near the Sophie's Cave. Glacial signs outside caves in lowlands or glacier valleys (e.g. moraines, drumlins, ice scratch marks on dolomite rocks) would not have survived outside the caves, because dolomites weathered quickly (to dolomite ash sands), also the post-LGM fluvial valley erosion was that rapid caused by climatic changes in the Meiendorf-Younger Dryas stadials/interstadials and must have destroyed those typical glacial signs within the valleys. Possibly relicts and signs are present, but have not been described or identified yet. This is different in "hard rock" middle high mountain regions of central Europe such as the Krkonoše Mountains (CZ, Pl: ENGEL et al. 2010), the Bohemian Forest Mountains (MENTLÍK et al. 2010), or northern German Harz Mountain Range (DIEDRICH 2013b), where all typical glacial structures and deposit types are still preserved in several valleys or also only as relic sediments in caves (cf. DIEDRICH 2013b). At least the deep valleys would fit to a glacier landscape. The early Postglacial was the main "hazard time" in Upper Frankonia, when caves collapsed due to climate changes in the Bölling/Alleröd times (see last speleothem genesis phases of Frankonian Alb in NORDHOFF 2005) when the valley formed quickly (?and glaciers melt rapidly).

### Cave use by mammals

Ice Age mammals were able to use the Sophie's Cave when it was cut during the valley genesis not before the late Middle Pleistocene.

**Middle Pleistocene:** After the first river terrace material was washed into the cave, when it fell dry before the first speleothem generation grew, martens used a part of the cave (Clausstein Hall) as den (only track records, Fig. 5B).

**Early-Middle Late Pleistocene:** The further erosion opened an entrance to the Bear's Passage which was used in the early and middle Late Pleistocene by small cave bear subspecies mainly with still primitive cave bear dentition and enamel morphology (P4 are three-coned, dated after methods of RABEDER 1999), which went deep into the cave for hibernation. About nine cave bear nests are still preserved in the Millionary Hall (DIEDRICH 2012a). Also weasels (*Mustela erminea*) used the same cave area as a den (also Zoolithen Cave, DIEDRICH 2013a), whereas porcupines (*Hystrix (Atherurus) brachyura*) went into the cave via the former entrance (into Bear's Passage) for bone chewing (possibly also

short den use there). In Upper Frankonia such Ice Age porcupines are not only indirectly proven in the Sophie's Cave (DIEDRICH 2012a), those were already found in the Hasenloch Cave, and the Fuchsloch Cave in the Franconian Alb with original bones, and additionally with typical rodent-chewed bones (RANKE 1879, NEHRING 1891, BRUNNER 1954, HELLER 1955). Also lions must have penetrated the cave for bear hunting (especially cubs) similar as demonstrated not only for the Zoolithen Cave (DIEDRICH 2011, 2012a). Lion remains are not directly present yet of the Sophie's Cave from the early-middle late Pleistocene, but from the final Late Pleistocene (DIEDRICH 2013a). Wolves also consumed the bear carcasses in the Sophie's Cave similar as documented for the Zoolithen Cave (DIEDRICH 2011, 2012b, 2013a), especially in the Bear's Passage somehow during the middle Late Pleistocene, when there was a block (large dropped ceiling block) between the Bear's Passage and the Reindeer Hall, a time where cave bears had to hibernate at the end of the Bear's Passage way too close to the entrance. There the bones show strongest carnivore damage caused by of all four, lions, hyenas, wolves and porcupines (DIEDRICH 2011, 2012a-b, 2013a).

**Latest Middle Late Pleistocene:** In this time, the largest cave bears species *Ursus ingressus* replaced the smaller and older cave bear subspecies (*U. spelaeus eremus/spelaeus*; cf. cave bear evolution and dating: RABEDER 1999, STILLER et al. 2010) which used the new opened today's entrance and anterior Sophie's Cave areas (Ahornloch; and Clausstein Halls, or Sand Chamber). A hyena den was present (early *U. ingressus* time) not only at the opposite König-Ludwigs Cave (Fig. 1), which pioneer work there started with the beginning of the "hyena den cave research" by BUCKLAND (1823). He worked against ESPER's (1774) biblical flood theories, explaining at least "non-cave bear bone assemblages in caves" to be of Ice Age hyena origin. Hyenas imported also in the Sophie's Cave few mammoth steppe animal prey remains (*Mammuthus primigenius*, *Coelodonta antiquitatis*, *Equus caballus przewalskii*, *Rangifer tarandus*, even *Panthera leo spelaea* juvenile remains, DIEDRICH 2014) into the short-term used den. Also similar few mammoth steppe prey was imported into the large hyena den of the Zoolithen Cave entrance, which is typical in boreal forest mountainous regions, and explains the "cave bear scavenging and hunt specialization" in those habitats (DIEDRICH 2011, 2013a).

**LGM:** Already before the last glaciation maximum in the final Late Pleistocene cave bears (last large species) and top predators (steppe lions, Ice Age spotted hyenas, Ice Age wolves) disappeared/became extinct with most species of the "mammoth steppe megafauna" from northern Germany and also from Upper Franconia (e.g. DIEDRICH 2013a).

### Neanderthal camp sites

Dense populations of hyena clans, wolf packs and cave bear families in Upper Frankonia might have been a good reason, why Middle Palaeolithic (Neanderthals) humans did not oc-

Fig. 5: Sophie's Cave genesis during the Pliocene to final late Pleistocene, river terrace relicts and cave use by animals (marten, hyena, wolf, cave bear dens) and by humans (Late Palaeolithic, Gravettian).

Abb. 5: Genese der Sophienhöhle während des Pliozäns bis zum ausgehenden Spät-Pleistozän, Flussterrassen-relikte und Höhlennutzung durch Höhlenbären und andere Tieren (Marder-, Hyänen-, Wolfs-, Höhlenbären-Horste) und durch Menschen (Spät-Paläolithikum, Grevettian).

cupy the large portal cave entrance of the Sophie's Cave or other caves nearby (similar at e.g. Zoolithen Cave or Große Teufels Cave). The killing of cave bears by Neanderthals is not proven yet in Europe, and the only site where a camp site was possibly present is the Hasenloch cavity near Potenstein.

#### **Cave use by Gravettians-Solutréens (middle Late Palaeolithic)**

In the final middle Late Pleistocene Gravettians used the Reindeer Hall only for shamanic purposes. They left there a "reindeer antler depot", similar as reported for the Magdalénian culture layers of the Oeger Cave (Sauerland Karst, Westphalia: BLEICHER 1993), or at the unclear dated Westeregeln open air site (DIEDRICH 2012c). The origin of the Sophie's Cave antler accumulation will be discussed with comparisons of other caves in Northern Germany (Sauerland Karst) in future in more detail to be of human origin. The antlers (estimated about 100 using the new documented finds and historically mentioned ones in STERNBERG 1835) are nearly all "shed male antlers", and are clearly not of "carnivore - hyena, wolf" selective origin (DIEDRICH 2012a-b, 2014). Additionally, those are found only limited in the Reindeer Hall, close to the "Elephant Ear" and "Bee Basket" speleothems (map in DIEDRICH 2014). After comparisons to other caves (Zoolithen Cave, Mühlbach Cave), the speleothems Elephant Ear and Bee Basket have not existed that large within the Late Palaeolithic Gravettian period (compared to dated ones in Mühlbach Cave: NORDHOFF 2005). The only large reindeer antlers, of which one shed male antler was dated C14 to 25.750±130 BP, were found in a "sanctuary hall", the Reindeer Hall of the Sophie's Cave. A different shamanic use is found in the Mäander Cave, which is also in Upper Frankonia. There only engravings on speleothems (abstract feminine symbols) were left by Late Magdalénians (BOSINSKI 2011). The absence of artifacts at both shamanic used cave sites in the caves is not unusual and also typical at sites with "cave paintings and engravings" of SW-France (e.g. LUMLEY ET AL. 1984). At both Upper Franconian cave sites, the Late Palaeolithic ritual places are deeper in the caves, where "camp site rubbish" is anyway not expected.

#### **Cave use by Late Magdalénians/Epipalaeolithics (final Late Palaeolithic)**

Late Magdalénians had a settlement rock shelter site very close to the Sophie's Cave situated close to the Neumühle (between Kirchahorn and Oberailsfeld) at the historically discovered and excavated Rennerfels rock shelter which includes some cavities (GUMPERT 1931). There, the Late Palaeolithic (deeper layer, "Madeleine" = herein Late Magdalénian VI) and Early Mesolithic (upper layer, "Tardenoisian" = herein: Epipalaeolithic and not Mesolithic) material was excavated (GUMPERT 1981) being housed at the Museum in Tüchersfeld. The Late Palaeolithic artefacts can be identified herein as belonging to the today's so-called Late Magdalénian VI and Epipalaeolithic based on two typical projectiles figured by GUMPERT (1981). This hunting camp rock shelter site in nearly today's river terrace elevation includes a fire place and artifacts and less bones (GUMPERT 1981).

The König-Ludwigs Cave (opposite to the Sophie's Cave) would have been a perfect Late Magdalénian/Epipalaeolithic

settlement or hunting camp cave site, too. It was also accessible for the humans at that time (14.000–12.000 BP) after (terrace elevation was below the entrance, Fig. 5F). It is known there, the archaeological layers have been removed already during the 19<sup>th</sup> century (SOMMER 2006) and were moved in front of the cave (still dumped there), similar as at the Sophie's Cave entrance area, where also the upper layers of the first halls have been scooped in front of the cave (or Ahornloch Hall branching areas). The final Late Paleolithic reindeer hunter groups (Late Magdalénians/Epipalaeolithics) were possibly present at further rock shelters similar to the Rennerfels (GUMPERT 1981) and not in the caves. At that time entrances were already collapsed or closed by speleothems (e.g. Sophie's Cave) or were not used (no art in caves anymore) as known from other cave-rich regions of Europe). The rock shelter cavities instead were used along the Upper Franconian valleys, because seasonally reindeer herd migration took place and allowed the hunt on those. It seems that the absence of further archaeological sites of those cultures is only a lack of research knowledge and cave history damage or non prosecutions of rock shelters. This is demonstrated at the Sophie's Cave and Mäander Cave with their presence of at least shamanic sanctuaries. Their camp sites along the river valleys (on the plateau margins or caves) must be prospected systematically in future. A further Epipalaeolithic site was described also not far in Upper Frankonia at the site Plankenfels (SCHÖNWEISS & STICHT 1968).

#### **Conclusions**

The Sophie's Cave can be used as a model for the cave evolution from a ponor cave, to transitional cave with floods and refills of river terrace and glacial (?valley glacier) sediments, which became a dry cave that collapsed and built finally its main speleothems at the end of the Ice Age and during the Holocene. Animals started to use the cave as a den within the Middle Pleistocene (marten den). In the Late Pleistocene mainly small cave bear subspecies (*Ursus spelaeus eremus/spelaeus*) denned in the cave, but those had to change the entrances and areas of denning due to lowering of the river terraces, and collapse of the older entrance (transitional cave stage). Those were hunted by lions and scavenged by hyenas and wolves also deeper in the cave. This is the reason why several of their hibernation nests (9) are still preserved in the deepest reachable part, the Millionary Hall. The closing of the Bear's Passage did not allow bears to hibernate there furthermore. Later, at the late Middle Pleistocene (32.000–24.000 BP) new larger *Ursus ingressus* cave bears appeared, which were able to use the new opened today's entrance. *U. ingressus* populations inhabited several caves along the Wiesent and branching valleys, and also still hyenas, wolves and lions which all disappeared already before the LGM (= before 19.000 BP). Hyenas and wolves also denned periodically in the Sophie's Cave entrance area. At this time of final Middle Late Pleistocene Late Palaeolithic Gravettians left a reindeer antler accumulation within the Reindeer Hall of the Sophie's cave. This depot counts more than 100 mainly shed antlers, which are all or dominated male antlers of which one was C14 dated with 25.750±130 BP. Possibly to this shamanic antler/bone deposition also two mammoth pelvis remains belong to, one being dated with an C14 age of 24.150

±130 BP. With few increase of the terrace, and tentative presence of a LGM valley glacier, all their bones were washed post-LGM into deeper and side-branched Sophie's Cave areas, mainly into the Bear's Catacombs. Possible glaciation models with larger snow fields (or ? small glaciers) of the Upper Franconian region could explain both, the absence of humans around the LGM (22.000–16.000 BP: Late Solutréens to Middle Magdalénians), and the strong and deep erosion of the river valleys, and even extremely high elevated young river terraces, such as tilt or glauconite sand relict sediments in valley sided caves. The massive erosion of the steep valleys must have happened directly post-LGM and may tentatively also be explained by smaller glaciation forms in which form however remains unclear with first models being under construction. First new human appearance in this study area are the Late Magdalénians within the end of the Ice Age when the climate allowed the resettlement of the middle high mountainous region. Whereas in the Ahorn Valley at the Rennerfels rock shelter a camp site is known of those Late Magdalénians/Epipalaeolithics. Whereas in the Mäander Cave feminine engravings are present (Magdalénian), in the Sophie's Cave an antler depot was left (Gravettian) – both indicating a shamanic use of medium-deep cave areas within the Upper Frankonian cave-rich region. Late Magdalénians/Epipalaeolithics reindeer hunter groups found the valley already deeply eroded similar as today, which were used by reindeer herds for seasonal migrations to the boreal forests. Epipalaeolithics have left at least one (or two) camp sites (Plankenfels, Rennerfels) in the Upper Frankonian region, with the end of the Ice Age.

The valley and cave genesis of the Upper Frankonian river valleys started mainly in the Pliocene when the region was still a plateau. Most of the caves evolved within the Pliocene to early Pleistocene in the “ground water level” (ponor caves) due to climatic changes which caused the erosion of the plateau – the time of the beginning of the landscape change. Caves in different elevations were then opened time by time by rivers along the valleys. Those were partly refilled and contain different elevated river terrace relict sediments between 450 to 375 meters a.s.l. The natural weathering and erosion such as river terrace lowering changed the accessibility of caves (for humans and animals) of the Upper Frankonian valleys. Some entrances were blocked by collapses, other were closed by terrace sediments or even speleothem layers. In general, caves in the study area above 410 meters a.s.l. contain Late Pleistocene to Middle Pleistocene megafauna remains. Below 410 meters, caves have only final Late Pleistocene (Late Magdalénian to Epipalaeolithics) human settlement or shamanic sites, or even Holocene Early Mesolithic (Rennerfels rock shelter) and younger epoch sites. In the Hasenloch Cave near Pottenstein Neanderthals used a smaller cave (also porcupines there) also in higher elevation as camp site, which was at the branching Püttlach Valley of another strongly frequented cave bear, wolf and hyena den cave, the Große Teufels Cave in the branching Weiherbach Valley. Also at this den site no human artifacts are known. At the time of the Neanderthals, the river valleys were about half to two-third deep eroded, as today and much less in its geomorphology relief. It seems, the larger cave systems where large mammal bone amounts were found, Neanderthals were unable to occupy due to dangerous competitions

with denning cave bears and hyena clans or dwelling wolves and lions.

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