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Lake and inland dunes as interconnected Systems: The story of Lake Tresssee and an adjacent dune field (Schleswig-Holstein, North Germany)

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Abstract

The specific aim of the study was to investigate how four adjacent geomorphological systems – a lake, a dune field, a small alluvial fan and a slope system – responded to the same impacts. Lake Tresssee is a shallow lake in the North of Germany (Schleswig-Holstein). During the Holocene, the lake's water surface declined drastically, predominately as a consequence of human impact. The adjacent inland dune field shows several traces of former sand drift events. Using 30 new radiocarbon ages and the results of 16 OSL samples, this study aims to create a new timeline tracing the interaction between lake and dunes, as well, as how both the lake and the dunes reacted to environmental changes. The water level of the lake is presumed to have peaked during the period before the Younger Dryas (YD; start at 10.73 ka BC). After the Boreal period (OSL age 8050 ± 690 BC) the level must have undergone fluctuations triggered by climatic events and the first human influences. The last demonstrable high water level was during the Late Bronze Age (1003–844 cal. BC). The first to the 9th century AD saw slightly shrinking water levels, and more significant ones thereafter. In the 19th century, the lake area was artificially reduced to a minimum by the human population. In the dunes, a total of seven different phases of sand drift were demonstrated for the last 13,000 years. It is one of the most precisely dated inland-dune chronologies of Central Europe. The small alluvial fan took shape mainly between the 13th and 17th centuries AD. After 1700 cal. BC (Middle Bronze Age), and again during the sixth and seventh centuries AD, we find enhanced slope activity with the formation of Holocene colluvia.

Keywords

aeolian chronology, inland dunes, interactions between geomorphological systems, lakes, North Germany, OSL dating, shrinking water levels

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Introduction

Climatic and human impacts on landscapes affect various geomorphologic systems in different ways. Sediment regularly travels from one location to another, triggered by changes in the vegetation cover and soil generated by activities such as forest clearing, agriculture and overgrazing, as well as climatic shifts (with the latter mainly affecting vegetation cover). Not only water but also wind can act as a transport medium for these sediment shifts, causing erosion and – as existing surfaces and soil, profiles are covered by fresh sediment – sedimentary accumulation (cf. Matthews and Seppälä, 2014; Tolksdorf and Kaiser, 2012).

We carried out our investigations in the inland dune field "Düne am Treßsee" and the adjacent Lake Tresssee in the northwest of the European sandbelt (Schleswig-Holstein, North Germany; Figure 1). In this area, four different geomorphological systems stand in close proximity to each other, offering a unique opportunity to study their different responses to past environmental impacts and their interactions with each other. In this context, we examine the following geomorphological landforms, which are located directly next to each other:

1. The inland dune field "Düne am Treßsee," with distinctly visible young sand covers above older podzol profiles (Polensky, 1982; Zölitz, 1989)
2. The directly adjacent Lake Tresssee is a strongly eutrophic lake extending over about 4.5 ha and located within an elongated basin of about 125 ha, which itself is covered by former lake sediments. Over time, the lake has drastically shrunk, presumably by a factor of 26 (Packschies, 1982; Stolz et al., 2020; Wiethold, 1998)
3. The alluvial fan of a small stream that runs along the edge of the dunes and flows into the lake basin
4. A slope beside the outflow of river Treene from the lake with clearly visible Holocene colluvial sediments on its foot.

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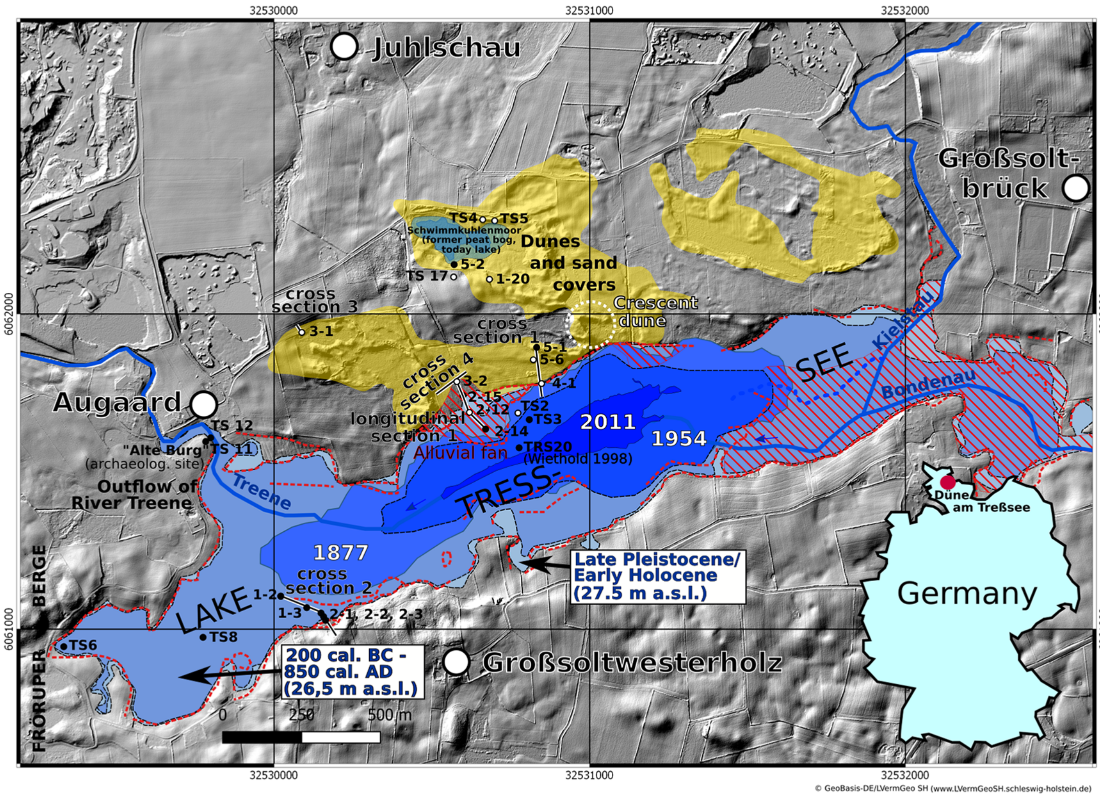

Figure 1. Digital elevation Model (DEM) of the lake and dune area and its location in Germany. The different historic lake levels are highlighted in blue (Landesamt für Vermessung und Geoinformation Schleswig-Holstein, 2011; Landesvermessungsamt Schleswig-Holstein, 1954; Preußische Landesaufnahme, 1877; yellow areas are dune sands after GÜK 250 (LLUR, 2012); alluvial fans are highlighted in red. Cross sections are marked with black lines, single cores mentioned in the text with black dots and soil pits with black and white dots.

With regard to the dunes, we wanted to investigate when they were created, what environmental conditions triggered the processes of blowing up of aeolian sand, as well as when and why the subsequent sand-drift events occurred. In the boundary area between the lake and the dunes, intermeshing effects could be expected.

Concerning the lake, we mainly aimed to understand the reasons behind its historical shrinkage, given that it was clearly much larger in the past and its size at specific points in time. Different types of human impacts are most likely responsible for this extreme retreat. In general, it must be assumed that the input of sediments and nutrients (combined with increased plant growth) from the inflowing streams into the lake must have resulted in increased sedimentation. Another reason could have been the impact of sand drifts from the adjacent dune field (Polensky, 1982). It is also known that during the 18th and 19th centuries attempts were made to artificially reduce the size of the lake in order to gain meadowland (Wiethold, 1998). The history of the vegetation and land use in the surrounding areas as well as the contents of algae and Cladocera remains of the lake sediments have been well investigated using two pollen profiles (Stolz et al., 2020; Wiethold, 1998). Following these first studies, which sought to reconstruct how climatic and anthropogenic factors caused variability in the lake levels and water conditions, there are still some open questions concerning the size of the lake at various points in the past. Our initial aim was to precisely reconstruct the water levels at different periods during the Younger Holocene and find out when Lake Tresseee reached its maximum extension. To achieve this, we focused especially on the rim of the former lake basin. The intermeshing zone of the lake's former northern shoreline: adjacent to the dunes, the former shorelines along the south of the basin, and the outermost western part of the lake basin, where we suspected indications of the highest lake water levels ever. In the northern basin, sand drifts blown out from the dunes contributed to the lake's sedimentation; these drifts are key to our understanding of how land in the surrounding areas was used during different periods in the past. Thus, for the alluvial fan of a small tributary, it made sense to reconstruct how soil erosion by water within its catchment and historical soil erosion processes on a slope beside the former lake basin contributed to the fan's growth.

Regional setting

Lake Tresssee (German: *Treßsee*; Danish: *Træsø*, from Danish *træ*=wood, meaning "the lake in the forest"; 54.701°N, 9.481°E) is located in Northern Germany (state of Schleswig-Holstein, Schleswig area, landscape of Angeln; Figure 1), about 20 km south of the city Flensburg and the border to Denmark. The Tresssee basin (about 3.3 km × 0.9 km), was formed by subglacial meltwater processes. At its southern edge, it is bordered by ground moraines (LLUR, 2012; Riedel and Polensky, 1987). The remaining part of Lake Tresssee is located in the center of the basin. Today, it is about 0.81 km × 0.18 km wide and up to 1.45 m deep. It is strongly eutrophic (Grudzinski, 2007). The water surface area is less than 5 ha and is widely covered by reeds – mostly *Phragmites australis* – and other water plant associations; Schröder (1985). The catchment of the lake is located in the historical landscape of Angeln and is about 133 km² in size. In 1877 and 1954, its water surface would still have been about 57 and 30 km², respectively; however, the surface decreased to only 9.5 km² in 1970 after the destruction of a weir (Hamer, 1995; Wiethold, 1998). The lake has five tributaries: two rivers in the east (the Bondenau and the Kielstau) and three quite small streams in the north, the northeast, and the south of the lake basin (Figure 1). The lake is the origin of the Treene River, which is the largest river in this landscape, flowing out from the lake basin in the southwest and draining with the Eider River into the North Sea.

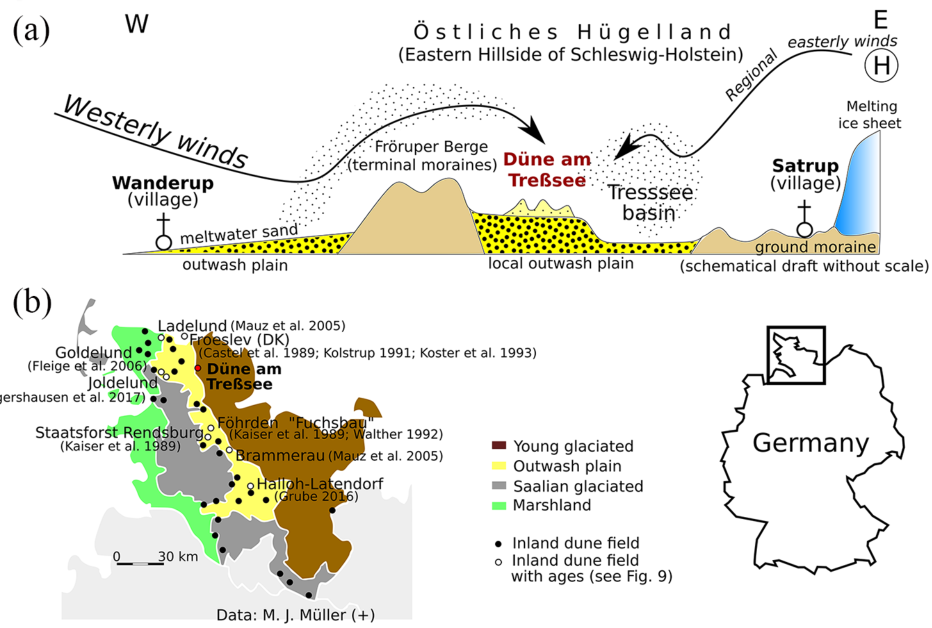

Figure 2. (a) Schematic diagram about the origin of the dune sands. (b) The location of the inland dune fields and the geomorphological units of Schleswig-Holstein (dunes with age determination are marked with black and white dots; e.g. Figure 9).

The adjacent dune field "Düne am Treßsee" in the lake's northern basin is located near the western edge of the young glaciated area of the Jutland Peninsula. It is a special situation, because most of the inland dune fields in this area are to be found on the Weichselian outwash plain and not within ground moraine area (Müller, 1999; Figure 2). The dune field (about 1.1 km × 0.7 km) is situated on a local outwash plain in the east of the Weichselian ice margin, an area characterized by gravel and mostly coarse sand. The dunes, most of which have well-developed podzols or visibly degradated podzol profiles and partly younger sand covers above, are generally more than 10 m thick; in some cases, crescent-shaped dune forms are visible (Zölitz, 1989), while in other locations there are only aeolian cover sands. Between the dunes, dune valleys and wet, drainless depressions (probably death-ice forms) exist. Traces of previous agricultural use in the dune field (field boundaries in the form of earth works, so-called *Knicks*, drainage ditches from the dune valleys, and plowing horizons) are also evident (Bronger and Polensky, 1985; Meynke, 1985; Packschies, 1982, 1984; Polensky, 1982; Rickert, 2013; Stolz et al., 2020; Wiethold, 1998). In several locations, the dune sands are interstratified with sediments from the former lake basin (Polensky, 1982; Wiethold, 1998).

The climate in northern Schleswig-Holstein is characterized by mild oceanic winters and cool summers (mean annual air temperature: 8.2°C; precipitation sum: 903 mm; measurement station of Flensburg 1961–1990; Mühr, 2007), with a westerly wind more than 55% of the time.

Development of Lake Tresssee

Reconstruction of the lake's history is mainly based on studies from lake sediment sequences and palynological studies by Stolz et al. (2020) and Wiethold (1998). By approximately 6000 BP, Lake Tresssee was already a eutrophic lake with a distinct macrophyte zone. Based on Cladocera and pollen results, we have proven that its largest expansion – to a size that filled nearly the entire basin – took place between 3635 and 3500 cal. BC, at the end of the Atlantic period and the start of the Subboreal period (Stolz et al., 2020). Thereafter, the evidence suggests that water levels tended to decrease, probably due to the cooler climate and lower precipitation trends during the Early Subboreal period (3521–3366 cal. BC). The subsequent drop in the water level of the lake may be related to the impact of the first humans on the lake area and on its outflow to the Treene River. From the Young Neolithic period (3881–3800 cal. BC) onward, the organic carbon content of the lake sediment decreased as the volume of clastic sediments in the lake increased. The advanced cultivation of the river flood plain must have led to lower lake levels. After the Bronze Age (approx. 2200 cal. BC), the lake became more eutrophic and the lake level appears to have decreased, according to Cladocera results. From the Late Iron Age to early medieval times (0–900 cal. AD), higher values of heavy metals are detectable in the lake sediments, presumably in consequence of early iron smelting in the surroundings (Stolz et al., 2020). Starting in the early medieval and Viking Age (800 cal. AD), the lake level declined continuously. In a former study, we documented the history of sedimentation in the outflowing upper Treene River and an increased sedimentation rate on the flood plain during the late medieval and early modern periods (14th–16th century; Stolz et al. (2016)). Dreibrodt and Bork (2005) showed the same result for Lake Belau (100 km to the southeast). During the 20th century, the lake became extremely eutrophic due to agriculture and wastewaters in its catchment, and the water level declined to a minimum. Since the 1990s, it has been part of the "Obere Treenelandschaft" conservation area.

Vegetation and landuse history

Two pollen diagrams are available from Lake Tresssee (TS 3 from Stolz et al. (2020) and TRS 20 from Wiethold, 1998) and a further two are available from two small mires in the neighboring terminal moraines of Froeruper Berge (Rickert, 2006). An overview of the vegetation history of Schleswig-Holstein was given by Nelle and Dörfler (2008).

Röschmann (1963) reported on several Mesolithic finds from the southern bank of the lake and in the east of the lake basin. From the Young Neolithic (3881–3800 cal. BC), the first indicator plants for open landscapes and grazing are visible in the most recent pollen core of the lake (TS 3; Stolz et al., 2020). From the Northern Bronze Age (approx. 1800–500 BC), there are fewer finds, mainly graves (Willroth, 1992). About 500 BC, there was widespread heath vegetation in the surroundings of the lake. Indicators of forest clearings, like a rise of Poaceae pollen, are visible in profile TS 3 (Stolz et al., 2020). Due to several settlements, graves, and single

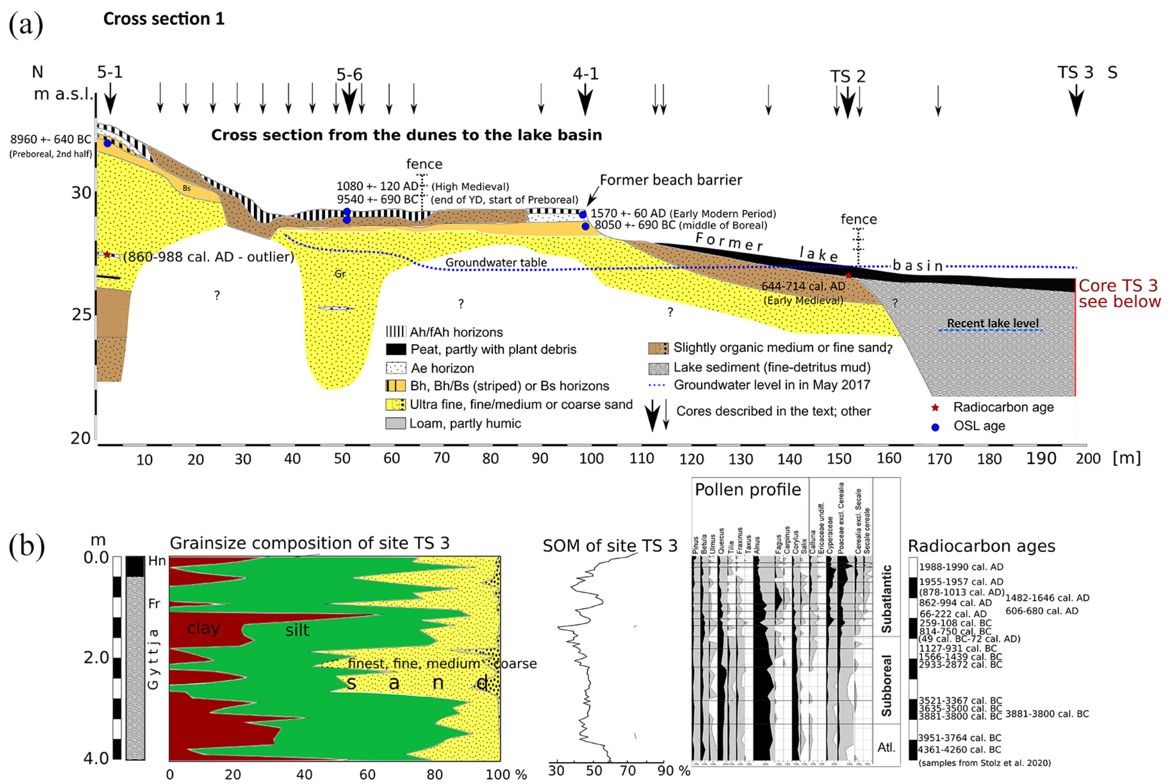

Figure 3. (a) Cross-section 1 from the dunes to the lake basin. Core TS 3 is located on the northern lakeshore (e.g. Stolz et al., 2020). (b) Detailed data of site TS 3, with a simplified pollen diagram (after Stolz et al., 2020; counted by Irena A. Pidek).

finds and from his own pollen results, Wiethold (1998) presumed a constantly settled landscape from the pre-Roman Iron Age (approx. 500 BC-0) to the Roman Period (55 BC–284 AD), but without complete deforestation. For the pre-Roman Iron Age and the Roman Period, activities of iron smelting in the surroundings are evidenced by finds of slags and iron tools (Hingst, 1974; Jöns, 1992/93; Meier, 2019). In relation to archaeological finds, the Migration Period started between the second third of the 4th century and the middle of the 5th century and lasted at least until the 7th century, with only rare finds (Willroth, 1992). In the radiocarbon-dated pollen profiles, the Migration Period seemed to start later in comparison to Lake Belau in the Holstein area to the south (Dörfler et al., 2012), although it did not show a complete absence of humans in the landscape of Angeln, but nevertheless there was probably a high migration rate. Also, from the early Viking Age (8th/9th century), there is only rare evidence of intensified settlement activities. However, from the Viking Age onward, from about 750 AD, pollen indicators of farming and grazing increased, probably due to an influx of people from Northern Jutland (Wiethold, 1998). The high medieval times started with strongly increased settlement activities in the 10th century. Wiethold (1998) presumes a completely cultivated landscape with small forest and widespread heaths and croplands. In the late medieval period, from about 1300 AD onward, the number of settlements again decreased (e.g. also Lungershausen et al., 2018 for North Friesia). Four deserted settlement sites are presumed for the Tresssee area in the 14th and 15th centuries (Kuhlmann, 1958). During the early modern period, from about 1450 AD onward, the share of forests in the region again declined to a minimum at the beginning of the 19th century (Härdtle, 1996). Today, the Schleswig area still has one of the lowest proportions of forest in Germany (about 5%; Kreis Schleswig-Flensburg, 1999). At the end of the 18th century, cropland increased in consequence of agricultural reforms (the so-called *Verkoppelung*). At the same time, the usage of forests for grazing and pig fattening was prohibited. This led to a high demand for new grasslands (Wiethold, 1998 after Mager, 1930). This may have been the reason for the artificial lowering of the lake level by deepening of the outflow of the Treene River.

Materials and methods

Coring, sampling, and sediment analysis

Between 2017 and 2019, we extracted more than 100 corings within the dune field, in and around the former lake basin. Additionally, several pits were dug in the research area to substantiate the results. The corings on the archaeological site "Alte Burg" took place with the permission of the state office for the preservation of Schleswig-Holstein. For coring, we used a pile corer (diameter: 4 cm), a Puerckhauer corer (diameter: 2 cm), and the Russian Instorf corer (for core TS 2 with pollen sampling; Figure 3). The detailed description of sediment profiles followed the German pedological mapping guidelines (Ad-hoc AG Boden, 2005) and, additionally, the guidelines of the IUSS Working Group WRB (2006). Selected profiles were sub-sampled at regular 5 cm intervals, for detailed laboratory analysis. All samples were analyzed using the following methods: grain size (method after Koehn; cf. Gee et al. (1986); Blume (2000); organic content was removed using H_2O_2), debris content (by mass), soil color (determined by Munsell scale), soil organic matter and carbon (loss on ignition, 430°C), and pH value (electrometric in *Aqua dest.* and in 0.01 M CaCl solution). For profile TS 4 (Figure 4) with dune sands, every 5 cm the roundness of grains ($n \geqslant 100$) in the most common fraction of medium sand (0.2–0.63 mm in diameter) was determined (Figure 4).

Smear slides

Biological components on 23 smear slides from three cores (TS 6, TS 8, 1–2) were analyzed to determine the terrestrial or lacustrine provenance of the sediments (Kelts, 2003). A small volume of fresh, unprocessed material was diluted in ~1 ml of distilled

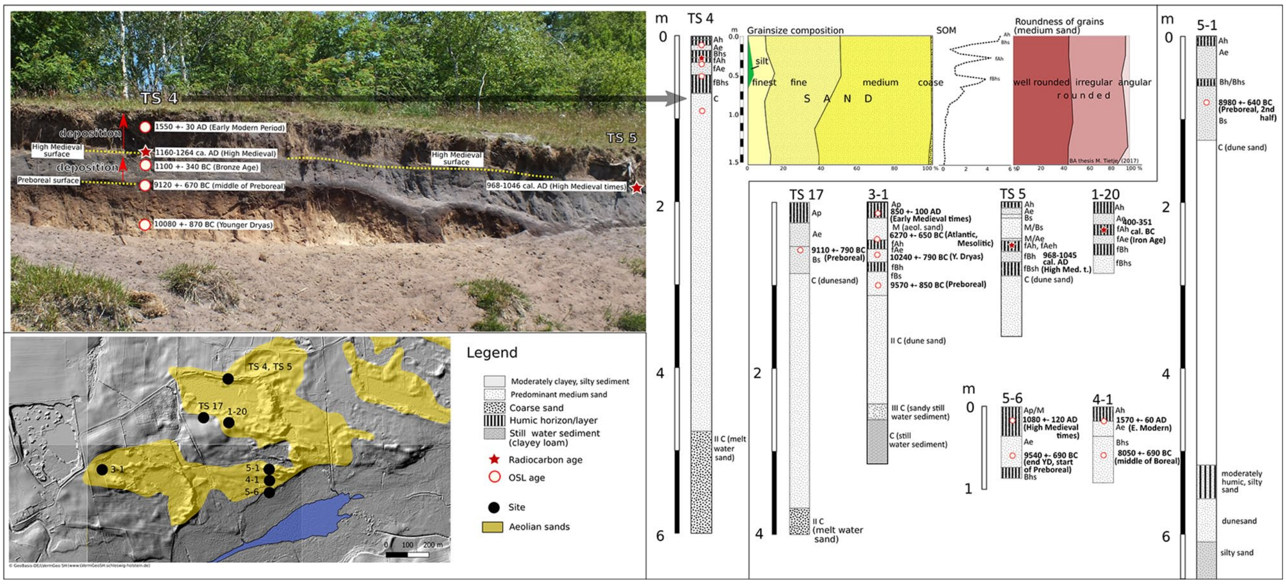

Figure 4. The structure and the age of the dunes and phases of aeolian activity in the past.

water and analyzed under compound light microscope with 100×–600× magnification. At least three slides were prepared for each sub-sample, using 0.1 ml volume of sample for one 24 mm × 60 mm slide. Remains of Cladocera chitinous body parts, *Chironomidae* head capsules, diatom frustules, *Pediastrum* algae, testate amoebae, and selected plant macrofossils were identified and quantified. Based on the overall composition of the sample, the deposition environment was categorized as terrestrial, mixed, or aquatic.

Pollen analysis

For the needs of the present study, the core TS 2 was sampled for pollen analysis. Two pollen samples (1 cm^3 each) were taken at the depth of 100 and 112 cm. The samples were macerated using the standard Erdtman's acetolysis method, after carbonates were removed by means of 10% HCl, and the mineral fraction by means of 38% HF (Berglund, Ralska-Jasiewiczowa, 1986). The frequency of pollen grains was not high, so the pollen sum (AP+NAP) was 150 in the 100 cm sample and 215 in the 112 cm sample.

The basic sum used for percentage calculations was the sum of AP (arboreal pollen – tree and shrub pollen), and NAP (non-arboreal pollen – dwarf shrub and herb pollen), excluding the pollen of aquatic and reed-swamp plants, Pteridophyta and Bryophyta spores, *Pediastrum* sp., and *Botryococcus* sp. algae. The results of the palynological analysis are presented as a percentage pollen diagram prepared using POLPAL software (Nalepka and Walanus, 2003).

The lake history is mainly based on studies from lake sediment sequences and palynological studies by Stolz et al. (2020) and Wiethold (1998). Both profiles originate from the northern part of the lake basin (Figure 1). The 4 m profile TS 3 from Stolz et al. (2020) contains 19 radiocarbon ages and covers the period from ca. 6300 BP until today. Wiethold's high-resolution pollen profile TRS20 does not contain any radiocarbon ages and starts in the older Subatlantic (younger Pre-Roman Iron Age) and ends in the modern age.

Radiocarbon dating

The 30 radiocarbon ages (AMS method; Table 1) were obtained from different profiles from the edges of the lake basin to detect the maximum lake level, and from the zone intermeshing with the dunes (plant material, wood, peat, gyttja, bulk samples). In all cases, before dating, carbonates were removed. The dating was carried out by Beta Analytic Radiocarbon Dating Laboratory, Miami, US and calibrated using OxCal 4.3 software (Bronk Ramsey, 1995) with the application of the INTCAL13 dataset after Reimer et al. (2013) and the NHZ1 datasets (Hua et al., 2013).

To calculate the sedimentation rate of core TS 3 (Figure 10, line 3), we used the dating results (19 ages) and the age depth model (constructed by use of the P Sequence function) from Stolz et al. (2020).

OSL dating

OSL samples were taken and transported in lightproof metal containers. The dating was carried out at the luminescence dating laboratory of the chair of Sedimentology and Quaternary Research at Freiburg University. OSL dating was performed on the sand-sized quartz fraction of 100–200 µm which was abundant in all samples and which is considered representative for the local sedimentation processes studied here (Figure 4). The OSL sample preparation followed standard lab procedures of the quartz coarse grain technique. After separating the required grain size fraction by wet sieving, carbonates and organic material were removed with 20% hydrochloric acid or 20% hydrogen peroxide. Quartz grains were then extracted by density separation using hetero-polytungstate heavy liquid (LST, 2.7 and 2.62 g/cm³). The following treatment with hydrofluoric acid (40%, 60 min) eliminated any potential feldspar contamination and removed the alpha-irradiated outer layer of the quartz grains. The purified quartz fraction was sieved again (100 µm), and then medium multiple grain aliquots (diameter of 4 mm) containing ca. 600 grains each were prepared. This relatively high grain number per aliquot was chosen due to weak OSL sensitivity.

OSL measurements were performed on a Risø TL-DA 15 reader using a standard single-aliquot regenerative dose (SAR) protocol (Murray and Wintle, 2000, 2003; Wintle and Murray, 2006). Sets of up to 69 aliquots per sample were stimulated with blue LED light ($\lambda = 470 \pm 30$ nm) at 125°C for 40 s. The resulting OSL signals were detected through a Hoya U340 filter ($\lambda = 330 \pm 40$ nm). Dose recovery tests were performed on quartz of all samples to find the most appropriate OSL measurement settings. For this purpose, aliquot sets were bleached and then irradiated with a defined beta dose. Measurement protocols with

Table 1. Radiocarbon ages.

Site	Sediment depth (cm)	Material	Radiocarbon age (BP)	Cal. age (2nd sigma range)	Probability	delta 13C	Laboratory no.	Reference
TS2	73–75	Peat/plant material	1340 ± 30 BP	644–714 cal. AD	84.10%	−27.2	Beta – 506873	This study
				744–765 cal. AD	11.30%			
TS3	30–32	Plant	116.84 ± 0.44 pMC	1988–1990 cal. AD	77.9%	−26.5	Beta – 506860	Stolz et al. (2020)
TS3	54–56	Peat/plant material	103.03 ± 0.38 pMC	1955–1957 cal. AD	95.4%	−26	Beta – 506861	Stolz et al. (2020)
TS3	62–64	Gyttja/sediment	1110 ± 30 BP	878–1013 cal. AD	95.4%	−31.4	Beta – 517917	Stolz et al. (2020)
TS3	82–84	Plant	320 ± 30 BP	1482–1646 cal. AD	95.4%	−26.6	Beta – 506862	Stolz et al. (2020)
TS3	92–94	Gyttja/sediment	1120 ± 30 BP	862–994 cal. AD	91.8%	−33.9	Beta – 506863	Stolz et al. (2020)
TS3	102–104	Gyttja/sediment	1380 ± 30 BP	606–680 cal. AD	95.4%	−28.5	Beta – 517918	Stolz et al. (2020)
TS3	114–116	Gyttja/plant material	1880 ± 30 BP	66–222 cal. AD	95.4%	−31.2	Beta – 506864	Stolz et al. (2020)
TS3	124–126	Gyttja/sediment	2160 ± 30 BP	259–108 cal. BC	55.3%	−33.2	Beta – 506865	Stolz et al. (2020)
				358–279 cal. BC	40.1%			
TS3	134–136	Gyttja/sediment	2580 ± 30 BP	814–750 cal. BC	87.0%	−31.6	Beta-497428	Stolz et al. (2020)
TS3	144–146	Gyttja/sediment	1990 ± 30 BP	49 cal. BC–72 cal. AD	95.4%	−32.5	Beta – 506866	Stolz et al. (2020)
TS3	164–166	Gyttja/sediment	2870 ± 30 BP	1127–931 cal. BC	95.4%	−31.2	Beta – 497429	Stolz et al. (2020)
TS3	192–194	Gyttja/sediment	3240 ± 30 BP	1566–1439 cal. BC	80.3%	32.3	Beta – 506867	Stolz et al. (2020)
TS3	212–214	Gyttja/sediment	4280 ± 30 BP	2933–2872 cal. BC	93.7%	−30	Beta – 517919	Stolz et al. (2020)
TS3	272–274	Gyttja/sediment	4670 ± 30 BP	3521–3367 cal. BC	93.8%	−30.8	Beta – 497430	Stolz et al. (2020)
TS3	282–284	Gyttja/sediment	4740 ± 30 BP	3635–3500 cal. BC	75.3%	−32.1	Beta – 506868	Stolz et al. (2020)
TS3	314–316	Gyttja/sediment	5090 ± 30 B	3881–3800 cal. BC	59.5%	−32.5	Beta – 506869	Stolz et al. (2020)
				3963–3895 cal. BP	35.9%			
TS3	332–334	Gyttja/sediment	5100 ± 30 BP	3881–3800 cal. BC	58.2%	−34.3	Beta – 506870	Stolz et al. (2020)
				3968–3896 cal. BC	37.2%			
TS3	354–356	Gyttja/plant material	5040 ± 30 BP	3951–3764 cal. BC	94.5%	−29.2	Beta – 506871	Stolz et al. (2020)
TS3	372–374	Gyttja/sediment	5470 ± 30 BP	4361–4260 cal. BC	95.4%	−31.6	Beta – 497431	Stolz et al. (2020)
1–2	50–52	Peaty sediment/plant material	840 ± 30 BP	1154–1264 cal. AD	95.10%	−28.5	Beta – 543232	This study
				1059–1062 cal. AD	0.30%			
1–2	87–89	Peaty sediment/plant material	950 ± 30 BP	1024–1155 cal. AD	95.40%	−28.2	Beta – 543233	This study
1–3	51–53	Peaty sediment/plant material	1560 ± 30 BP	420–565 cal. AD	95.40%	−26.4	Beta – 543234	This study
1–3	86–88	Peaty sediment/ plant material	1700 ± 30 BP	313–406 cal. AD	71.90%	−27.5	Beta – 543235	This study
				254–304 cal. AD	23.50%			
1–20	35–48	Buried A horizon/sediment	2270 ± 30 PB	400–351 cal. BC	49.30%	−28.9	Beta – 506874	This study
				304–210 cal. BC	46.10%			
1–2	47	Reed/plant material	1590 ± 30 BP	406–542 cal. AD	95.40%	−30.7	Beta – 543237	This study
2–1	80	Woody material/ wood	1700 ± 30BP	313–406 cal. AD	71.90%	−30.3	Beta – 543236	This study
				254–304 cal. AD	23.50%			
2–1	85–90	Peaty sediment/ plant material	2780 ± 30 BP	1003–844 cal. BC	95.40%	−29.1	Beta – 543227	This study
2–2	62	Woody material/wood	1840 ± 30 BP	86–242 cal. AD	95.40%	−28.1	Beta – 543230	This study
2–3	48–50	Peaty sediment/plant material	1500 ± 30 BP	532–639 cal. AD	85%	−28	Beta – 543229	This study
				432–489 cal. AD	10.40%			
2–3	98–100	Peaty sediment/ plant material	2130 ± 30 BP	210–52 cal. BC	85.10%	−29	Beta – 543228	This study
				350–308 cal. BC	10.30%			
2–12	40–42	Peat/sediment	340 ± 30 BP	1470–1640 cal. AD	95.40%	−27.8	Beta – 506875	This study
2–12	82–91	Peat/sediment	540 +– 30 BP	1388–1437 cal. AD	65.60%	−28.3	Beta – 506876	This study
				1316–1354 cal. AD	29.80%			
2–14	139–141	peat/ sediment	740 ± 30 BP	1224–1291 cal. AD	95.40%	−28.1	Beta – 517916	This study
2–15	264–269	Organic sediment	4370 ± 30 BP	3031–2907 cal. BC	85.70%	−27.6	Beta – 543244	This study
				3089–3055 cal. BC	9.70%			
2–15	535–538	Organic sediment	5440 ± 30 BP	4346–4246 cal. BC	95.40%	−27.6	Beta – 543245	This study
5–1	517–537	Buried A horizon/ sediment	1130 ± 30 BP	860–988 cal. AD	86.60%	−29.7	Beta – 506878	This study
				805–842 cal. AD	5.60%			
				777–791 cal. AD	3.20%			
5–2	81–83	Peat/plant material	102.01 ± 0.38 pMC	1954–1956 cal. AD	95.40%	−28.2	Beta –543231	This study
TS4	31	Buried A horizon/sediment	830 ± 30 PB	1160–1264 cal. AD	95.40%	−30.9	Beta – 506872	This study
TS5	58	Organic sediment	020 +/- 30 BP	968–1046 cal. AD	90.50%	−28.8	Beta – 543243	This study
				1094–1120 cal. AD	4.20%			
				1141–1147 cal. AD	0.70%			
TS6	20	Peat/plant material	1340 ± 30 BP	644–714 cal. AD	83.80%	−28.4	Beta – 543242	This study
				743–766 cal. AD	11.60%			
TS6	50	Gyttja/sediment	2560 ± 30 BP	805–764 cal. BC	66%	−29.8	Beta – 520151	This study
				643–553 cal. BC	21.80%			
				686–666 cal. BC	7.60%			
TS6	100	Gyttja/sediment	2890 ± 30 BP	1133–978 cal. BC	85.80%	−33.1	Beta – 520152	This study

(Continued)

Table 1. (Continued)

Site	Sediment depth (cm)	Material	Radiocarbon age (BP)	Cal. age (2nd sigma range)	Probability	delta 13C	Laboratory no.	Reference
				1196–1142 cal. BC	9.60%			
TS8	78	Peat/plant material	990 ± 30 BP	989–1052 cal. AD	57.20%	−25.7	Beta – 520153	This study
				1080–1152 cal. AD	38.20%			
TS8	100–105	Peat/plant material	1200 ± 30 BP	765–895 cal. AD	87.90%	−26.9	Beta – 520154	This study
				714–744 cal. AD	6.10%			
				928–940 cal. AD	1.50%			
TS8	140–145	Peat/plant material	1200 ± 30 BP	765–895 cal. AD	87.90%	−28.3	Beta – 520155	This study
				714–744 cal. AD	6.10%			
				928–940 cal. AD	1.50%			
TS8	240–245	Gyttja/sediment	2000 ± 30 BP	55 cal. BC–70 ca. AD	94.80%	−29.7	Beta – 520156	This study
				84–80 cal. BC	0.60%			
TS8	340–345	Gyttja/sediment	2010 ± 30 BP	61 cal. BC–65 cal. AD	91.50%	−29.2	Beta – 520157	This study
				91–69 cal. BC	3.90%			
TS11	110–117	Organic sediment	1440 ± 30 BP	566–654 cal. AD	95.40%	−27.4	Beta – 543241	This study
TS12	146–152	Plant material	3370 ± 30 BP	1745–1611 cal. BC	94.70%	−29	Beta – 543246	This study
				1572–1566 cal. BC	0.70%			

different preheat and cut-heat temperatures were then tested. A SAR protocol with a preheat temperature of 200°C (10 s) and a cut-heat temperature of 180°C (0 s) best recovered the known dose. This protocol was applied to determine the unknown paleodose of the natural samples. The measurement files were analyzed using Luminescence Analyst (v4.12) and R Luminescence Package (v0.9.0.110).

High-resolution gamma ray spectrometry was applied to estimate the sediment dose rates arising from the decay of primordial radionuclides. Subsamples were dried and homogenized before they were sent to VKTA Dresden (Niederniveaumesslabor Felsenkeller) to determine the contents of uranium (^{238}U), thorium (^{232}Th), and potassium (^{40}K). The contribution of the cosmic radiation to the total dose rate was estimated as a function of geographic position, burial depth, and altitude. The dose rates and the resulting OSL ages were calculated using the DRAC dose rate and age calculator (v1.2, Durcan et al., 2015).

For dose rate determiniation, an estimated uniform water content of 10% ± 5% was used instead of the measured one to account for seasonal and Holocene variations.

Geographic Information System (GIS) analysis

For an exact leveling of the profiles and to calculate the former expansion of the lake, we used QGIS software with a digital elevation model of 1 m resolution (Landesamt für Vermessung und Geoinformation Schleswig-Holstein, 2017) and our own tachymetric measurements.

Results

OSL dating results

The dated quartz exhibited favorable properties. However, some aliquots showed an IRSL (infrared stimulated luminescence) signal indicating feldspar contamination. Aliquots emitting a significant IRSL signal and yielding a low OSL IR depletion ratio (Duller, 2003) were excluded from further analysis. The samples TS4 (38 cm), 3–1 (15 cm), 3–1 (40 cm) from the upper parts of dune sediment sequences show a positively skewed dose distribution more typical for colluvial sediments or short transport routes. Similar effects could be observed for samples from the alluvial fan (3–2, 32 cm) and from the transition zone between the edge of the dune field and the lake basin (4–1; 20 cm and 5–6; 30 cm). All other OSL ages of this study show mostly narrow symmetric Paleodose distributions with overdisperion values below 15%, proving sufficient daylight exposure during the most recent sedimentation process to be dated. Especially, this is the case for the samples of Late Glacial/Early Holocene dune sands.

In order to minimize the risk of age overestimation, the Minimum Age Model (MAM) according to Galbraith et al. (1999) was applied to samples with stronger scattering or positively skewed dose distributions. The MAM selects the best bleached grain population by giving preference to the lowest dose values, here an overdispersion of sigma $b = 0.15$ was allowed in the MAM. All other Paleodoses were calculated using the Central Age Model (CAM, Galbraith et al., 1999). The results are shown in Table 2.

The structure and age of the dunes

The elevation of the dunes ranges between 28 and 41 m a.s.l. At the site TS 4 (36.4 m a.s.l.; Figure 4), we detected a 4.7 m thickness of pure dune sands over coarser meltwater sands. At the site TS 17 (35 m a.s.l.), this value was 3.7 m, and at site 5–1 (32.4 m a.s.l.), we cored 5.2 m of pure dune sand. Below, there are around 2.5 m of partly humic/loamy sands, and below these, there are pure dune sands again, until 9 m depth or deeper. Therefore, we can estimate a maximum thickness of the highest dune of 18 m (pure dune sands without interruption about 12 m).

We investigated the dune field in seven different locations in detail (TS 4/TS 5, TS 17, 1–20, 3–1, and in the cross section no. 1 with the sites 5–1, 4–1 and 5–6; Figure 4). The composition of dune sands is visible in Figure 4. Medium sand (0.2–0.63 mm) is the main size range (44–56%), followed by fine sand (0.063–0.2 mm; 23–36%). Coarse sand (0.62–2 mm) is rare and mainly present in deeper parts of the dune profiles (up to 2%). The sediment is free of debris and without any soil formations as well as nearly free of silt, clay, and organic matter (SOM). Often, the soils on top of the dunes are podzols with a quite similar characteristic (up to 120 cm thick), but frequently shortened by erosion and partly covered by younger sand layers including at least one younger soil formation (initial podsolization). Within the A and B horizons, the maximum of SOM is 3.8 to 4.7%. The regular color of non-weathered dune sand above the ground water level is 10 YR 7/8. The pH values range between 4.2 and 5.4.

To study the phases of sand drift in the past, we searched purposefully for relevant locations and found two areas with younger sand covers within the dune field. At site TS 4/TS 5 (Figure 4), we found a well visible younger sand cover (26–54 cm thick) with initial podsolization. Zölitz (1989) previously mentioned the open profile beside a path. Below the sand cover, there is a well visible

Table 2. OSL dating results. Uranium, thorium and potassium contents were determined by high resulution gamma ray spectroscopy (HPGe detector). Following daughter products were considered: Uranium: Pb-214 and Bi-214 for Ra-226; Thorium: Ac-228 for Ra-228, Pb-212 and Tl-208 for Th-228. The measured (original) water content is shown, for total dose rate calculation an overall water content of 10% ± 5% was assumed. The Paleodose was determined using the Central Age Model (CAM) or the Minimum Age Model (MAM) respectively. MAM doses are printed in bold (sigma b = 0.15).

Lab reference	Depth (m)	U (ppm)	Th (ppm)	K (%)	Water content (%)	D_{cosmic} (Gy/ka)	D_{total} (Gy/ka)	Paleodose (D_e) (Gy)	Overdispersion (OD)	OSL age (ka)
TS 4	0.14	0.24 ± 0.02	0.72 ± 0.06	0.65 ± 0.07	1.1	0.25 ± 0.03	0.91 ± 0.06	0.43 ± 0.01	13.5	0.47 ± 0.03
TS 4	0.38	0.21 ± 0.02	0.55 ± 0.04	0.44 ± 0.04	3.2	0.22 ± 0.02	0.69 ± 0.04	**2.14 ± 0.19**	48.0	3.12 ± 0.34
TS 4	0.64	0.28 ± 0.03	0.92 ± 0.08	0.76 ± 0.05	–	0.19 ± 0.02	0.99 ± 0.05	11.0 ± 0.27	12.9	11.14 ± 0.67
TS 4	0.9	0.35 ± 0.03	0.89 ± 0.08	0.67 ± 0.07	3.8	0.18 ± 0.02	0.91 ± 0.06	11.01 ± 0.24	11.1	12.1 ± 0.87
TS17	0.55	0.32 ± 0.03	0.93 ± 0.08	0.69 ± 0.07	0.2	0.2 ± 0.02	0.94 ± 0.06	10.47 ± 0.26	11.4	11.13 ± 0.79
3–1	0.15	0.31 ± 0.02	0.72 ± 0.05	0.46 ± 0.05	11.5	0.25 ± 0.03	0.76 ± 0.05	0.88 ± 0.05	34.5	1.17 ± 0.1
3–1	0.4	0.33 ± 0.03	0.77 ± 0.07	0.54 ± 0.06	7.9	0.21 ± 0.02	0.81 ± 0.05	6.72 ± 0.27	22.9	8.29 ± 0.65
3–1	0.65	0.27 ± 0.03	0.85 ± 0.08	0.61 ± 0.05	5.3	0.19 ± 0.02	0.85 ± 0.05	10.28 ± 0.28	13.0	12.26 ± 0.79
3–1	1	0.31 ± 0.03	0.87 ± 0.08	0.66 ± 0.07	3.8	0.18 ± 0.02	0.89 ± 0.06	10.3 ± 0.26	13.5	11.59 ± 0.85
3–2	0.32	0.51 ± 0.04	0.84 ± 0.08	0.57 ± 0.05	0.5	0.22 ± 0.02	0.89 ± 0.05	**1.05 ± 0.11**	46.5	1.19 ± 0.14
3–2	0.8	0.31 ± 0.03	0.82 ± 0.08	0.72 ± 0.05	12.2	0.19 ± 0.02	0.94 ± 0.05	11.17 ± 0.23	11.5	11.83 ± 0.7
4–1	0.2	0.31 ± 0.03	0.89 ± 0.08	0.63 ± 0.07	0.2	0.24 ± 0.02	0.91 ± 0.06	**0.41 ± 0.05**	61.6	0.45 ± 0.06
4–1	0.65	0.34 ± 0.03	0.89 ± 0.07	0.61 ± 0.06	2.7	0.19 ± 0.02	0.86 ± 0.06	8.69 ± 0.06	12.9	10.07 ± 0.69
5–1	0.8	0.29 ± 0.03	0.88 ± 0.08	0.76 ± 0.05	8.1	0.19 ± 0.02	1.01 ± 0.05	11.04 ± 0.24	11.8	10.98 ± 0.64
5–6	0.3	0.25 ± 0.02	0.50 ± 0.05	0.36 ± 0.03	1.4	0.22 ± 0.02	0.63 ± 0.04	0.59 ± 0.07	13.2	0.94 ± 0.12
5–6	0.55	0.22 ± 0.02	0.45 ± 0.05	0.54 ± 0.04	5.2	0.2 ± 0.02	0.76 ± 0.04	8.75 ± 0.16	11.4	11.56 ± 0.69

fossil podzol (80 cm thick) with an old surface. It shows the marks of former tree roots. The dune sand below is quite homogeneous and reaches up to 370 cm. In the lower part, the sands are slightly reduced. Below the dune sands, there are coarser melt-water sands. From the visible old surface, we took two samples for radiocarbon dating (site TS 4, 31 cm; humic sediment; 1160–1264 cal. AD; and site TS 5, 58 cm; humic sediment; ca. 968–1146 AD, High Middle Ages). Furthermore, OSL ages of site TS 4 gave the following results: 14 cm (in the sand cover): 1550 ± 30 AD, early modern period; 38 cm (below the old surface): 1100 ± 340 BC, Bronze Age; 64 cm (sesquioxide horizon of the fossil podzol): 9120 ± 670 BC, middle of Preboreal; 90 cm (non-weathered dune sand): 10,080 ± 870 BC, Younger Dryas).

At the site 1–20 (Figure 4), there is a similar situation with a thin sand cover of 30 cm and a visible old surface. The radiocarbon sample of the covered A horizon gave an age of 400–351 cal. BC (Iron Age).

On a dune next to the lake basin (see also Figure 3), we took from site 5–1 (Figure 4) one OSL sample at 80 cm depth (sesquioxide horizon) with a result of 8980 ± 640 BC (second half of the Preboreal). The profile shows dune sand until 9 m depth, interrupted, however, by slightly humic and silty/loamy sands between 520 and 750 cm depth. We tried to get a radiocarbon age of this layer in 517–337 cm depth; however the result must be regarded as an outlier, because the overlying OSL age is much younger (the reason for this is probably a contamination during coring). At the foot of this dune to the south, we took two samples from a flat-deposited slightly humic colluvial sand layer of site 5–6, 30 cm. It gave an age of 1080 ± 120 AD (High Middle Ages). The sediment from the sesquioxide horizon of the same site gave an age of 9540 ± 690 BC (end of the Younger Dryas / start of the Preboreal). A second site is located in a former erosion edge of the lake, cut in an aeolian sand layer (site 4–1). In 15 cm depth, a slightly humic colluvium gave an age of 1570 ± 60 AD (early modern period) and another sample from 65 cm (dune sand, probably older colluvium) gave an age of 8050 ± 690 BC (middle of Boreal).

In the location TS 17 (Figure 4), there is a gently sloped aeolian sand layer or flat dune with a strongly eroded podzol and a clearly visible plowing horizon. The dune sands are 370 cm thick in this place, and in their lowest part, they intermesh with the coarser meltwater sands below. An OSL sample from 55 cm gave the age of 9110 ± 790 BC (Preboreal).

At site 3–1 (Figure 4) in the southwestern part of the dune field (clearly offsite of the lake basin; Figure 6, cross-section 3), a broken-down edge is visible. Its genesis is unknown. The permanent open profile shows dune sands with a well-developed podzol (110 cm thick). The sands are 240 cm thick and located above resp. beside of clayey-sandy, fine layered and obvious older still-water sediments. We took four OSL samples from the dune sands at site 3–1: 15 cm (850 ± 100 AD, Early Middle Ages/Viking Age), 40 cm (6270 ± 650 BC), 65 cm (10,240 ± 790 BC) and 100 cm (9570 ± 850 BC).

Investigations at the edges of the lake basin concerning the development of lake levels

We extracted more than 50 corings and pits at the edges of the former lake basin to reconstruct former lake levels. This investigations took place (1) at the northern bank (area intermeshing with the dunes), (2) at the southern bank (with ground moraines in the Westerholz forest), and (3) in the outermost west of the lake basin (next to the terminal moraines "Fröruper Berge"). All locations can be seen in Figure 1.

The northern bank of the lake basin (zone adjacent to the dunes; cross-section 1). The southern part of the dune field ends with a rough cover of aeolian sands (Figure 4). The podzols in this location are partly modified by former agriculture (plowing). Onto the lake basin, the sand cover is cut by a former beach barrier/erosion edge (an OSL age of these steps showed the age 8050 ± 690 BC (middle of Boreal); see the text section about the dunes). Below it, there is a slightly inclined former beach zone of the lake. It consists of peaty silting up sediment (SOM 55%–72%; site TS 2, 0–75 cm depth); light brown in the upper part (0–22 cm) and dark brown in the lower part. A sample from the base of this deposit (73–75 cm) gave the radiocarbon age of 644–714 ± cal. AD (Migration Period/Early Middle Ages). The dark sediment becomes thicker in the direction of the lake. Below, there is a dark to grey colored reworked dune sand with more or less organic content (SOM: 1%–4%; 75–115 cm). Furthermore, two palynological samples taken from the sandy bottom sediments of the TS

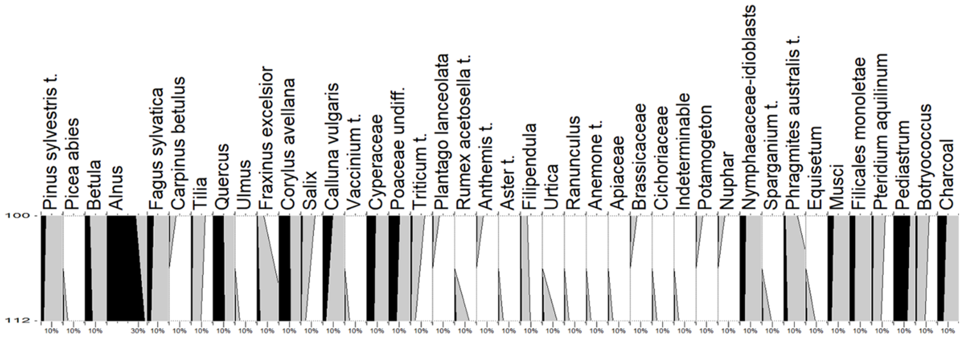

Figure 5. Pollen diagram of the core TS 2 (uppermost lake deposits on the northern shoreline). For the location, see Figure 1.

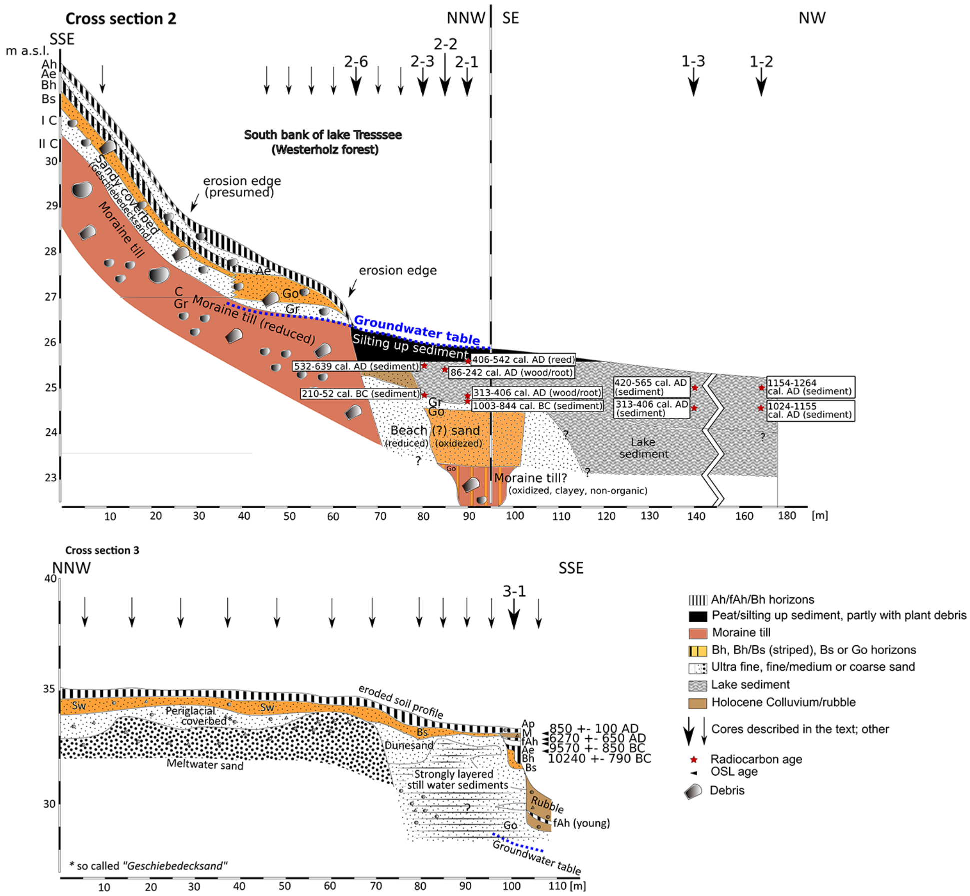

Figure 6. Cross-sections 2 and 3 on the southern lakeshore and northwest of the dune field, with an interpretation of landscape evolution. For the locations, see Figure 1.

2 core were analyzed (depths: 100 and 112 cm; for the results see Figure 5).

After about 55 m to the south, there is the beginning of the morphologic subglacial channel, the actual lake basin. It is filled with dark colored gyttja (in the center of the lake it should be more than 20 m thick; Packschies, 1982). From this location onward, the area is densely overgrown by reed and some trees. It is extremely swampy, and the recent lakeside cannot be accessed on foot. 70 m from the lakeside, we took the 4 m core TS 3. Its pollen and Cladocera contents and chemical parameters were already precisely described and dated with 19 radiocarbon ages (Stolz et al., 2020). The grain size composition of this site is shown in the lower part of Figure 3.

The development of the organic content shows that it initially decreased at about 4000 BC (from 60% to 45%), although the sand input shows an increase prior to that date. Thereafter, the curve is quite constant at an intermediate level (about 45%). Its lowest point was reached at approximately 200 BC (about 30%). Another increase took place between the third and ninth centuries AD, and the last one in the uppermost part of the profile (in the 1960s, when the site dried up). During the last 2000 years, especially, the organic curve has been nearly opposite to that of the sand input.

The southern bank (adjacent to the ground moraines of Westerholz forest; cross-section 2). The slope at the border of the lake basin in the south consists of glacial till containing big boulders (Figure 6). It is covered by a 130 cm thick periglacial cover bed with 9%–73% debris content and a visible enrichment of stones in the upper part (*Geschiebedecksand;* Grimmel, 1973; about 85% is sand (of which 50% is medium sand) and about 15% of it is silt (mainly coarse silt). The soil is a podzol with indications of a former cambisol. The land-use is beech forest. At barely 29 m a.s.l. on the slope, a former erosion edge of the lake can be presumed, but without typical sediments.

Outside of the slope, at a clear visible (second) erosion edge of the former lake (Figure 6: site 2–6) there is a typical strong organic silting up sediment of the lake with many plant remains. In the lower part, there is a gyttja sediment with many plant remains like thatch and woody roots. It is sandy (63%), with about 25% silt and up to 10% clay. The organic content ranges between 50 and 70%. In the first 45 m from the slope (sites 2–3, 2–-2, 2–1), there is a sand layer that is about 170 cm thick. It is slightly organic (0.5–3.5%, up to 14% in one thin layer), dirty-grey colored (gley 5/5B, 2.5/10BG), and strongly sandy (85–93% sand content). In one location of the former beach (site 2–1), we found moraine material with debris content below the sand layer. Dreibrodt and Bork (2005) described similar deposits on the shoreline of Lake Belau. Outside of the former beach zone, the actual lake basin with lake sediments several meters thick begins.

We took six samples for radiocarbon dating from the organic silting up sediments in this beach zone (Figure 6, upper part):

- Site 2–3, 48–50 cm, middle of the organic sediment: 532–639 cal. AD (bulk sample; Migration Period)
- Site 2–3, 98–100 cm, base of the organic sediment: 210–52 cal. BC (bulk sample; Iron Age)
- Site 2–2, 62 cm: 86–242 cal. AD (tree root as a proof of trees in this location; Roman Period)
- Site 2–1 (2019), 47 cm: 406–542 cal. AD (reed; Migration Period)
- Site 2–1, 80 cm: 313–406 cal. AD (root; Migration Period)
- Site 2–1, 85–90 cm: 1003–844 cal. BC (organic sediment; Late Bronze Age).

Further out in the lake basin (75 and 160 m from the slope), the following two sites are located, both entirely consisting of lake sediment:

- Site 1–3, 51–52 cm: 420–566 cal. AD (lake sediment; Migration Period)
- Site 1–3, 86–88 cm: 313–406 cal. AD (lake sediment; Migration Period)
- Site 1–2, 50–52 cm: 1154–1264 cal. AD (lake sediment; High Middle Ages) Site 1–2, 87–89 cm: 1024–1155 cal. AD (lake sediment; High Middle Ages).

The outermost western part of the lake basin (next to the terminal moraines "Fröruper Berge"). To determine the former extension of the lake in the outermost western part of the basin, we took two cores TS 6 (93 cm) and TS 8 (400 cm) (Figure 8). Core TS 6 (located in the small, northwestern bay of the basin (Figure 1) comprises entirely highly organic, terrestrial sediment with occasional reed remains. The upper 23 cm is slightly sandy; the rest is strong clayey-silty. We took three samples at the following depths:

- TS 6, 20 cm: ca. 644–714 AD (plant material; Migration Period/Early Middle Ages)
- TS 6, 50 cm: 807–764 cal. BC (organic sediment; Iron Age, Hallstatt)
- TS 6, 100 cm: 1133–978 BC (organic sediment; Late Bronze Age).

The absolute absence of aquatic indicators (including mostly Cladocera and algae) in smear slides from this sediment sequence proves its terrestrial provenance.

Site TS 8 is located 400 to the east in the center of the western lake basin and comprises highly organic sediments. The upper part, until ca. 100 cm depth, looks more peat-like, with several plant remains (organic content: 51–95%). The lower part consists of more homogeneous gyttja sediment (organic content: 85–95%). We took the following five samples:

- TS 8, 78 cm: 989–1052 cal. AD (plant material; High Middle Ages)
- TS 8, 100–105 cm: 765–895 cal. AD (plant material; Early Middle Ages/Viking Age)
- TS 8, 140–45 cm: 765–895 cal. AD (plant material; same age like before)
- TS 8, 240–245 cm: 55 cal. BC-70 cal. AD (gyttja; Roman Period)
- TS 8, 340–345 cm: 61 cal. BC-65 cal. AD (gyttja; Roman Period).

In the first 105 cm, no remains of Cladocerans or diatoms were present. Also, in the first 245 cm of the core, there were only few Cladocerans, representing solely littoral species (*Alonella excisa, Alona guttata* var. *tuberculata Alona affinis, Chydorus sphaericus*), which indicates shallow waters near the shore. The part below in the depth of 245 cm had more lacustrine character; nevertheless, only the sample from 340–345 cm can be regarded as deposited in deeper water conditions with diatoms and Cladocerans, including planktonic *Bosmina longirostris* and Daphnia sp., well represented with the absence of fen-vegetation indicators (e.g. testate amoebae, brown mosses, and Sphagnum fossils).

Investigations of a small alluvial fan

Small alluvial fans are well suited to studying local soil erosion processes in a specific catchment (mostly investigated in semiarid environments, but also in Central Europe; Bork et al., 1998; Dotterweich, 2008; Dreibrodt et al. (2010); Stolz and Grunert, 2010). A small alluvial fan (about 230 × 280 m long/wide) is located at the north bank of the lake basin. It belongs to a small stream, around 2 km long, originating in the east of the village of Juhlschau (today partly tubed; Figure 1, cross-section 4). The alluvial fan must have been a small delta in the past, when the lake level was higher. Today, there is no longer any direct contact between the fan and the existing lake. To investigate the alluvial fan, we cored a cross-section consisting of 11 corings in the upper part of the fan (Figure 7, cross-section 4). This shows that the sediments of the fan can be divided into a dark-black colored upper part (10 YR 2/1), a dark-brown middle part (10 YR 3/1), and a reddish-brown older part (10 YR 3/4). In some places, a thin pale-colored pure sand layer (probably of aeolian origin) divides the lowest and the middle part. The upper part is about 30 cm thick and quite humic (soil organic matter – SOM 4%–9%) and consists of strong silty fine sands. The middle part is about 20–100 cm thick and consists of weak silty fine and medium sand (SOM: 1–5%). The lowest part is strongly wet and reduced and consists of strong silty fine sand (SOM: 0.5%). From the center of the fan (site 3–2, 80 cm), we took an OSL sample of this layer and obtained a result of 9810 ± 700 BC (end of the Younger Dryas). The upper part of site 3–2 shows a more humic, younger sediment with the OSL age 830 ± 140 AD (Early Middle Ages/Viking Age). In all parts, a clear layering is visible. Thirty-five meters to the WSW, a former channel of the stream is detectable, filled with peaty sediments (Figure 7, site 2–8). This is an indication that the sedimentation process must have had different spatial focuses on the alluvial fan in the past.

In the longitudinal transect 1 (Figure 7, lower part), the fan is gently sloped. In the downward direction, it increasingly consists

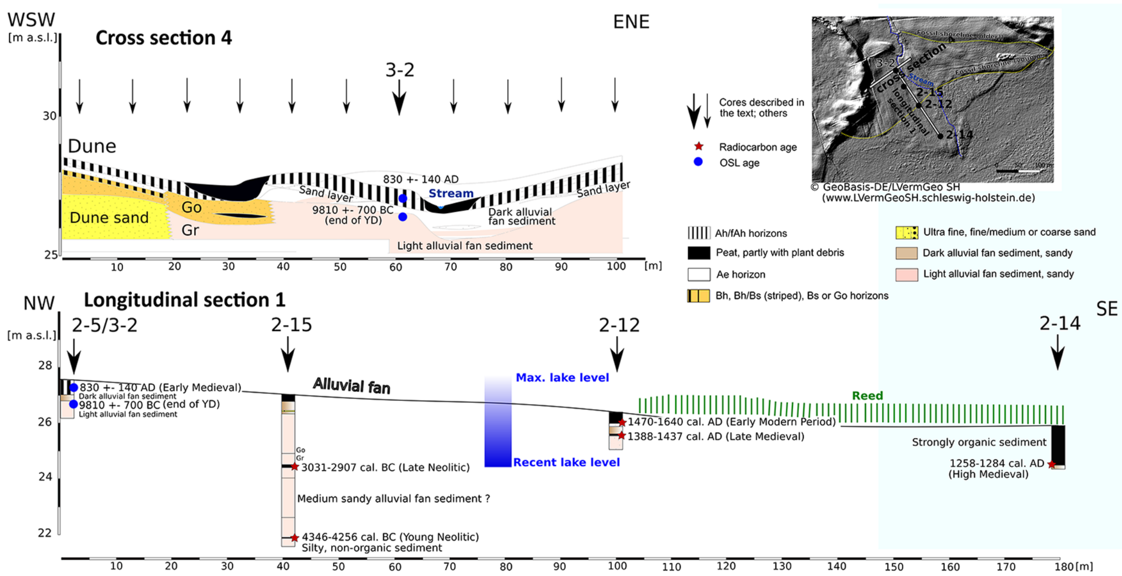

Figure 7. The structure of a small alluvial fan at the northern bank of the lake as an indicator of soil erosion. The dating results enable us to reconstruct sedimentation phases (with cross-section 4). The included sand layers are indicators of aeolian activities in the surrounding areas.

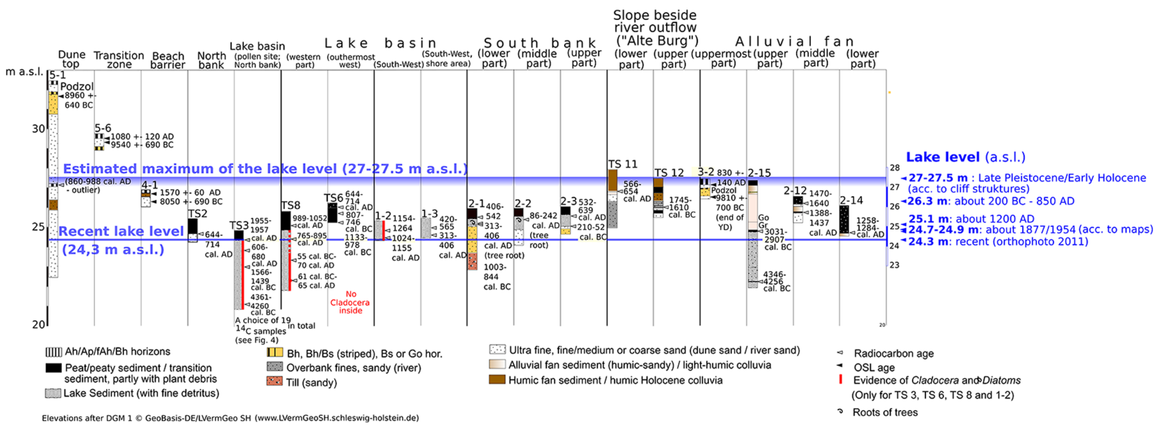

Figure 8. Comparison of the altitude of several sampling sites that were used in this study to indicate former lake levels. Samples of site TS 3 from Stolz et al. (2020).

of sandy layers alternating with strongly organic deposits. At site 2–15, there is a slight organic fine sandy-silty partly layered sediment in the upper part down to 110 cm. Afterwards the sediment becomes more medium and fine sandy in alternating layering and the organic content decreases. At depths of 264 and 535 cm, there are thin organic layers suitable for radiocarbon dating. The results were 3031–2907 cal. BC (Late Neolithic) and ca. 4346–4256 BC (Young Neolithic). Below, there is a coarse silty sediment, poor in organic content. It differs strongly from the typical lake sediment of Lake Tresssee in core TS 3, which is strongly organic and contains significantly less clastic content (Stolz et al., 2020). Site 2–12 is an alternating profile of dirty clayey sand layers and strong organic peat-like layers. We carried out two radiocarbon ages and obtained results of 1470–1640 cal. AD (early modern period; 40–42 cm depth) and 1388–1437 cal. AD (Late Middle Ages; 82–91 cm depth). The pH values vary between 5.5 and 6. This is quite high in comparison to other locations in the surroundings and indicates for a former influence of the lake in this location on the lower part of the alluvial fan. The area between this site and site 2–14, located 80 m below, is very wet and densely overgrown by reed. Site 2–14 shows a well-layered, dark brown, sandy-silty, and strong organic sediment down to 141 cm. It is extremely wet in the upper part. Below, there is a dirty silty medium sand. A sample from the lowest part of the mud at 139–141 cm gave the result 1258–1284 cal. AD (High Middle Ages).

A slope situation at the outflow of Treene River (“Alte Burg” hill)

The southwestern slope beside the outflow of Treene River from the lake basin belongs to an oval hill (its top is 30.0 m a.s.l. and 5.8 m above the riverbank; Figures 1 and 8). It is located close to the small settlement of Augaard. Recently, it has been used as grassland and it is commonly known as “*Alte Burg*” (meaning “old castle”; Röschmann, 1963). However, as yet no structural remains have been found at this location. We cored a transect from the top of the hill to the riverbank and found no evidence of a building on it. On the top and along the slope, we only found a partly eroded pseudogleyic soil with a plowing horizon as evidence for former agriculture. The substrate is of natural origin (*Geschiebedecksand* over till; this hill probably formed as a kame). On the north-facing slope, 15 m (site TS 11) and 21 m (site TS 12) downwards from the top, we found the profiles covered by Holocene colluvial. At site TS 11, this sediment is 117 cm thick, quite homogenous and light-brown colored. Below, there is a gleyic reduced bright-colored fine sand, free of debris. We took a

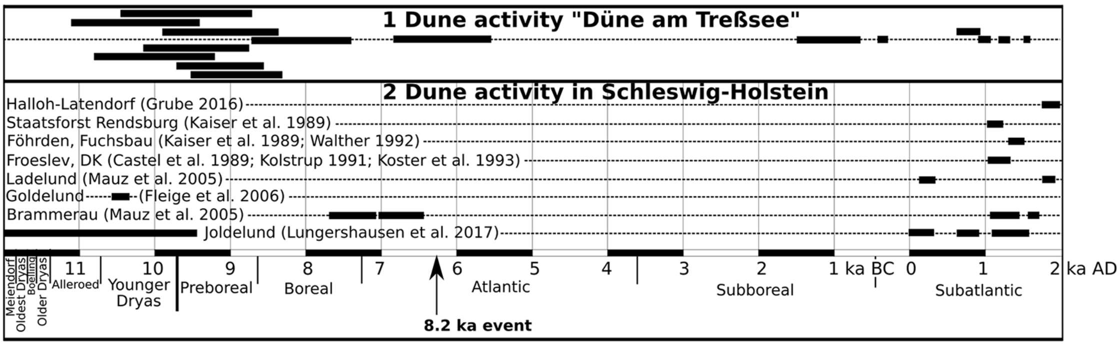

Figure 9. Sand drift activities for the last 13,000 years within the dunefield "*Düne am Treßsee*" with a comparison to the results from seven other dune fields in Schleswig-Holstein and South Denmark (for the locations, see Figure 2).

sample for radiocarbon dating from the basis of the colluvium at 100–117 cm depth (weak humic sediment): the result was ca. 566–654 AD (Migration Period).

Site TS 12, on the foot slope, consists of a strongly organic Holocene colluvium that is 116 cm deep. It is loamy, finely sandy, and dark colored, with a loamier and more organic section in the middle (45–81 cm). Flood plain sediments may have influenced its formation. Below the colluvium is a well-layered sediment sequence of river origin with thin alternating strata of fine sand and organic sediment (116–157 cm). Below that is a bright medium sand layer (until 162 cm) and a dark-brown peat layer. At the base of this profile, we detected organic free river sand that had highly visible fine layering in some areas. For dating, we took one sample from the base of the alternating sand/organic sequence (147–157 cm). This yielded a date of 1745–1611 cal. BC (early/middle Bronze Age).

Discussion

Dune dynamics

Land-use changes on dunes can trigger sand-drifting processes. There are studies dealing with young dune activity – mostly sand-drift events and partial dune reactivation – in Central Europe (Böse et al., 2002; Grube, 2016; Hilgers, 2007; Köster, 2010; Küster et al., 2014; Lungershausen et al., 2018; Schirmer, 1999; Völkel et al., 2011). Tolksdorf and Kaiser (2012) gave an overview about aeolian dynamics in the European sand belt. Sand covers above ancient surfaces lead to phases of higher pressure in land use, locally with a strengthening of sand drifts (Schirmer, 1999). Periods of abandonment and reforestation can be detected from younger soil formations and initial soils. These covered horizons can be dated using the ^{14}C method. By using luminescence dating, subdivisions of single sand covers are possible (Hilgers, 2007).

The formation of inland dunes in Schleswig-Holstein (Germany) and Southern Jutland (Denmark) on the Weichselian outwash plain and the young glaciated area was frequently presumed to have occurred between the Younger Dryas – or earlier – and the Boreal (10.73–7.27 ka BC; Fleige et al., 2006; Müller, 1999; Zölitz, 1989; cf. Litt 2007) and to have partially lasted until the Early Atlantic (OSL: 6380 ± 260 BC; Mauz et al., 2005). Tolksdorf and Kaiser (2012) presume a human impact on sand drifting processes already for Mesolithic cultures in the north of the Central European sand belt. Younger sand covers, formed by sand drifts triggered by human impact, were mostly dated to the early modern period (after 1500 until about 1800 AD; Grube, 2016; Mauz et al., 2005), to the High Middle Ages (about 1000–1300 AD; Castel et al., 1989; Kaiser et al., 1989; Kolstrup, 1991; Köster et al., 1993; Mauz et al., 2005; Walther, 1992) and, in part, to the Early Middle Ages (about 750–900 cal. AD) and the late Roman Period (about 150–300 cal. AD; Lungershausen et al., 2018). The dependence of regional settlement intensity and archaeological cultures must also be considered (Tolksdorf and Kaiser, 2012). Müller (1999) refers to the presumably different ages of dune fields, depending on their distance from possible sand sources.

In comparison to the other inland dune fields in Schleswig-Holstein, the dune field "Düne am Treßsee" is the only one in the north of Schleswig-Holstein that is located in the area of the Weichselian ice sheet (Müller, 1999). All the other ones are mostly situated on the Weichselian outwash plain (*Niedere Geest*) or on older Saalian moraines (*Hohe Geest*) in the west (Figure 2). The source of the sand is not clear. Wiethold (1998) and Zölitz (1989) presumed an origin from the Weichselian outwash plain in the west, which was drifted in by the frequent westerly winds. One indication of this theory is the presence of at least one crescent-shaped dune and its open side in the southeastern direction (see the marking in Figure 1). Another possibility would be an origin from the directly neighboring Tresssee basin and/or from a local outwash plain in the east of the Weichselian ice margin, drifted in by easterly winds. These winds could have been triggered during the Late Glacial Period by high-pressure areas over the withdrawing ice sheet in the east of the Tresssee area (Figure 2).

From 15 OSL of sands and radiocarbon ages of buried surfaces, we observed seven different phases of aeolian activity within the younger Pleistocene and the Holocene (Figure 9).

Phase 1. Triggered by aeolian activity, the dune field's development lasted from the onset of the Younger Dryas (10,080 ± 870 BC) until the end of the Preboreal (OSL: 9110 ± 790 BC). This result fits well with Schirmer's (1999) so-called "dune period" for the European sand belt. As we did not take samples from the base or from the heart of the dunes, however, some areas of the dune field may have already formed during the Older Dryas (11.59–11.4 ka BC; tundra vegetation; Litt et al., 2001), the Oldest Dryas (11.85–11.72 ka BC; tundra vegetation) during the Meiendorf interstadial (12.5–11.85 ka BC; shrubby vegetation). Initial dunes could have been even formed already toward the end of the Mecklenburg phase (about 15–13 ka cal. BC, marked by the last Weichselian Ice advance in Schleswig-Holstein and steppe vegetation). In the intervening Boelling and Alleroed interstadial periods (partly marked by wood vegetation), surfaces can be assumed to have been largely stable (Litt et al., 2001). The dunes cannot be older than that; however, as prior to that the area was still glaciated (further north in Scandinavia, comparable dune fields are younger and originate from the early Holocene; Matthews and Seppälä, 2014). There is still some controversy over the exact period when the ice in eastern Schleswig-Holstein and southern Denmark – and especially in the area around the lowland fjords (German: *Förden*) – disappeared (Litt et al., 2007; Smed, 1997). Lungershausen et al. (2018) assume that the aeolian sand covers in the Joldelund area were primarily formed during the Younger

Dryas (Figure 9). Müller (1999) presumed that inland dunes in Schleswig Holstein took shape during the Preboreal. At 3–1 in the western part of the dune field, there is an apparent age inversion. However, the error ranges of both samples overlap. Therefore, it can be assumed that the aging is nearly identical. We measured 10,240 ± 790 BC at a depth of 65 cm, and 9570 ± 850 BC at 100 cm. At site 3–1 and others (sites 4–1 and 5–6), it is noticeable that samples with small depth spacing in the same profile sometimes show quite large differences in OSL ages (cf. Table 2). This can be explained by undetectable intermediate erosion events or hiatuses.

Phase 2. A second phase of aeolian activity took place in the middle of the Boreal period (8050 ± 690 BC). What is probably a reworked sand layer from the shore zone of the lake at site 4–1 was OSL dated to the mid-Boreal (8050 ± 690 BC). These ages fit to the age of sand deposits from Brammerau in Schleswig-Holstein from the late Boreal and the beginning Atlantic (Figure 9; Mauz et al., 2005). This is remarkable, given that we must already assume forest vegetation for the Preboreal and the Boreal (9610–7270 BC) periods in Northern Germany (Behre and Lade, 1986; Dörfler et al., 2012). However, these ages are close to the “dune and dune transformation period” of Schirmer (1999).

Phase 3. Holocene sediment dynamics on the dunes could be evidenced for the mid-Atlantic and Mesolithic periods (OSL age: 6270 ± 650 BC; site 3–1). However, this sample shows a positively skewed dose distribution, which is more typical for colluvial sediments. We presume a natural influence of the 8.2 ka event, in the form of rapid cooling and changes in vegetation (Kobashi et al., 2007). Mauz et al. (2005) processed a similar OSL age in Brammerau (6380 ± 260 BC; Figure 2).

Phases 4–7. Younger phases could be demonstrated for the Late Bronze Age (phase 4; 1130 ± 340 BC; OSL; site TS 3), the Iron Age (phase 5; 400–351 cal. BC; ^{14}C; site 1–20), and the Viking Age/High Middle Ages (phase 6; 968–1046 cal. AD and 1160–1264 cal. AD; ^{14}C; sites TS 5 and TS 4; and 1100 ± 340; OSL; site TS 4). A last documented phase (phase 7) happened during the early modern period (1550 ± 30 AD; OSL, site TS 4). Drifting processes for entire dunes can be excluded for all of these phases; we only expect sand to have traveled over short distances. By contrast, for the Joldelund area (28 km WSW) Lungershausen et al. (2018) detected newly formed dunes during the Holocene, as a consequence of human activity.

In cross sections 1 and 3, located at the edges of the dunes, we found references to colluvial processes by water, plowing, or short-distance wind drift from the High Middle Ages to the early modern period (12th–16th century). The evidence for this comes from positively skewed dose distributions of OSL analyses, which are more typical for colluvial sediments and sedimentary features like organic contents and soil colors.

The results regarding grain size distribution of TS 4 site are remarkable (Figure 4), showing more fine sand and less coarse sand in the upper, younger layers of the dune. This could be an indication of stronger winds during the Late Glacial Period.

From the lake’s sand records (results from core TS 3; Figure 3), the phases of sand drift can be clearly confirmed for the Bronze Age (OSL: 1100 ± 340, the Iron Age (400–351 cal. BC), and the High Middle Ages (12th–13th century), in particular. It is not possible to make a statement for later time periods, due to a hiatus in site TS 3; Stolz et al., 2020). However, the fact that there were long periods of sand input (consisting of dune sands, especially medium and fine sands) to the lake is remarkable. Higher values are already visible starting with the Young Neolithic (ca. 3800 BC), and later, especially from the Late and End Neolithic (ca. 3500–1800 BC) and the Viking Age (ca. 750–1050 AD). These values suggest a long history of land-use; predominantly pasture farming, within the dune field. However, the values for the sand input must be used carefully, as sands of all subfractions can also enter the lake via its tributaries. Wind-related sand drifts require the absence of vegetation, at least on very local sites. In the younger Holocene, this is a clear proof of anthropogenic land use in this area, especially grazing. In particular, the presence of *Calluna* pollen of core TS 3 (Stolz et al., 2020) fits well with our results for sand records on the lake (Figure 10, line 4). In the past, a sand drift toward the lake basin must have been possible whenever the heath areas in the dunes were grazed extensively – especially during the winter and spring, when there were strong westerly/northwesterly winds and less vegetation cover. Lungershausen et al. (2018: Table 1) give a good overview of the existing research on sand-drift events, and even the abandonment of complete villages due to sand drifts. The assumed forest share in the Schleswig area yields values similar to ours (Härdtle, 1996; Wiethold, 1998). On the other hand, climatic effects are hard to find within these data. It is unclear whether the cattle driven over the historic *Ochsenweg* (ox way; Danish *Hærvejen*) road from Wedel (Hamburg) to the north of Jutland (Müller, 2002) influenced the destruction of vegetation on the dunes.

Most of the other studies of dunes in Schleswig-Holstein confirm our results for sand drifts during the High Middle Ages (ca. 1000–1300 AD; Mauz et al., 2005; Brammerau; Kaiser et al., 1989; Rendsburg; Köster et al., 1993; Froeslev, South Denmark; for locations see Figure 2). Lungershausen et al. (2018) indicated that the strongest young sand drift was in late medieval times (ca. 1300–-1500 AD), and they even demonstrated the formation of new dunes during the early modern period. Similar results are known for the dunes next to Föhrden/Rendsburg (Walther, 1992); Ladelund (Mauz et al., 2005; next to the German-Danish border) and Halloh-Latendorf (Grube, 2016; south of Neumünster). In Ladelund, they also evidenced sand drifts for the Roman Period, and for Joldelund starting with the Roman Period and the Early Middle Ages/Viking Age (Lungershausen et al., 2018). In Halloh-Latendorf, Grube (2016) described eleven quite young soil formations, one above the other, in a profile that is 1 m thick.

Lake level changes

Former changes in the level of the lake can be detected using ecological markers in records of sediments from the lake (Dörfler et al., 2012; Dreibrodt and Wiethold, 2015; Zolitschka et al., 2000) and its former shorelines, and by dating limnic sediments in the outer parts of the lake basin (Kaiser, 1998). The resizing of Lake Tresssee and the lowering of lake levels was a common consequence of human impact (Stolz et al., 2020; Wiethold, 1998). Climatic influences can have comparable effects on lakes (Punning et al., 2005).

Lake Tresssee acts as a sediment trap for its tributaries, reducing sedimentation processes on the flood plains directly downstream of it (Stolz et al., 2016). The lake basin, which is probably 15 to 25 m deep according to Holocene lake deposits (Packschies, 1982), was already full by about 2800 BC (Stolz et al., 2020); thereafter, the flow rate through the lake rose and the sedimentation rate fell (Figure 10, line 3). For the same period, Dreibrodt and Bork (2005) indicated the first phase of an enhanced sedimentation into Lake Belau (demonstrated by the K input to the lake). In consequence, today the lake is quite shallow and sensitive to variations in its outflow. We think that when the flood plain of the upper Treene valley was covered by dense vegetation, such as alluvial forest, higher lake levels were possible. Strong flush events could have made the outflow more passable again. The same effect was obtained by human impact activities on the floodplain such as forest clearing and grazing, and at the river with respect to its regulation, such as straightening and deepening of

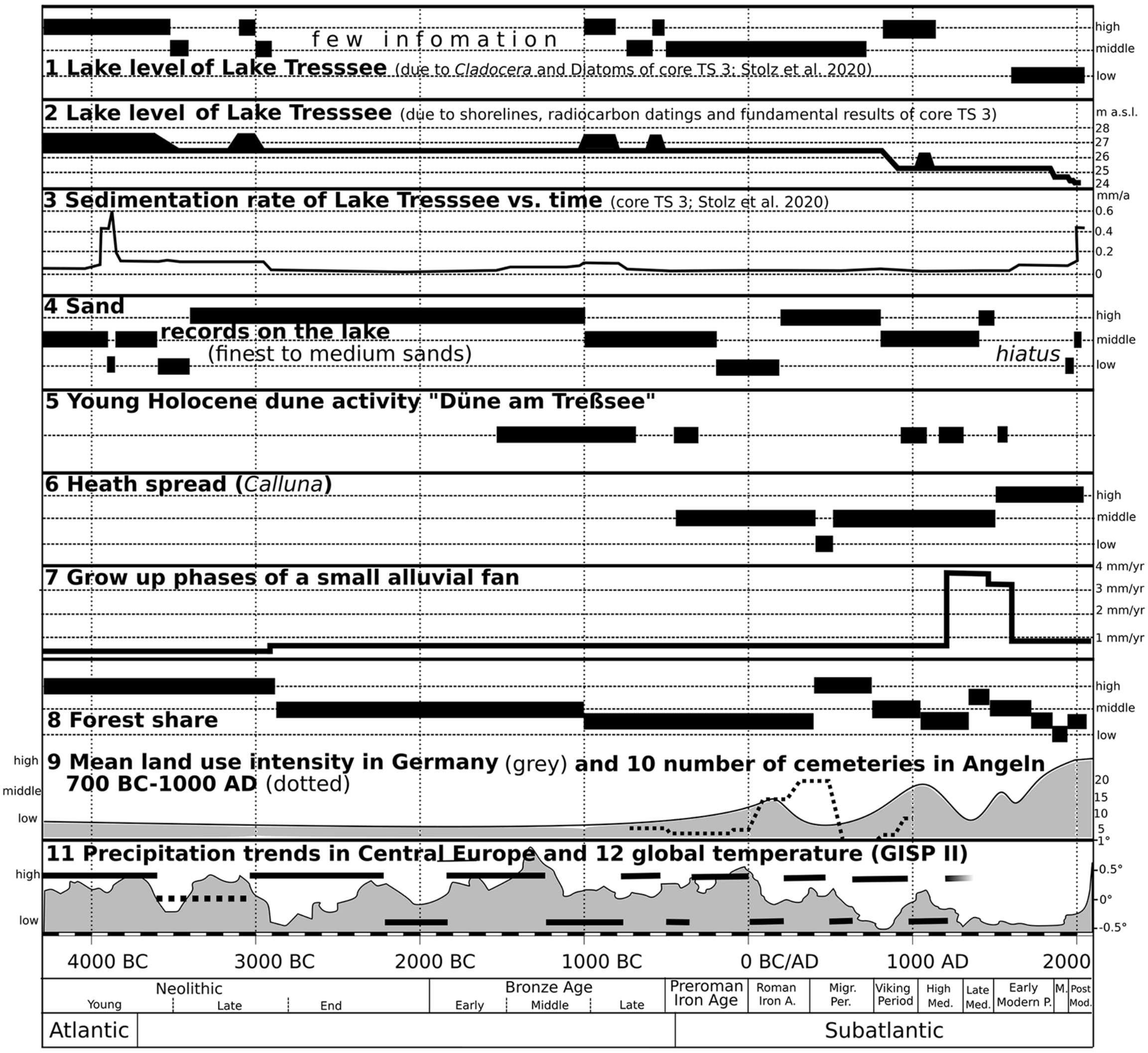

Figure 10. Timelines for Lake Tressee and their comparison to related results: (1) Lake Tresssee water level, as derived from Cladocera and diatoms of core TS 3 (Stolz et al., 2020). (2) Lake Tresssee water level, as derived from shorelines, radiocarbon dating, and the basic results of core TS 3 (this study). (3) Lake Tresssee sedimentation rate (Stolz et al., 2020). (4) Sand records into the Lake (this study). (5) Young Holocene dune activity in "Düne am Treßsee" over the last 14,000 years (this study; see Figure 9). (6) Heath spread (*Calluna* pollen; core TS 3 (Stolz et al., 2020). (7) Developmental phases of a small alluvial fan. (8) Forest share (after Härdtle, 1996; partially complemented by data from Dörfler et al., 2012 and Nelle and Dörfler, 2008). (9) Mean land-use intensity in Germany (Dotterweich, 2008). (10) Number of cemeteries in the Angeln landscape in 700 BC–1000 AD (after Willroth, 1992). (11) Precipitation trends in Central Europe (Jäger, 2002). (12) Global temperature of the GISP II ice core (Alley, 2004).

the riverbed) (Stolz et al., 2020). In this regard, a comparison with deeper lakes (cf. Magny et al., 2003) makes less sense. From High Middle Ages at latest (ca. 1000 cal. AD, but probably earlier), water level fluctuations were predominantly influenced by human impact activities rather than climatic effects (cf. Packschies, 1982; Wiethold, 1998).

The recent water level of the lake is 24.3 m a.s.l. (state in the year 2011, with seasonal fluctuations; Figure 8). This must be the lowest level ever recorded after artificial efforts to lower it, especially in the nineteenth and twentieth centuries (Wiethold, 1998). Packschies (1984) presumed the presence of lake sediments on about 125 ha; however, he did not carry out any analyses on macroscopic remains or similar. This value roughly corresponds to the total area of the morphologic lake basin. With the help of a digital elevation model (resolution of 1 m), we modeled the lake level necessary to flood the whole basin in this area. The calculated value is between 27 and 27.5 m a.s.l. Furthermore, we took the lake areas from the 1877 and 1954 topographic maps (Landesvermessungsamt Schleswig-Holstein, 1954; Preußische Landesaufnahme, 1877;) and calculated average lake levels between 24.7 and 24.9 m a.s.l. (both states of the lake are also still visible on orthophotos and in the field from the vegetation patterns).

From the paleoecological analyses of the core TS 3 (Cladocera, diatoms, and others; Stolz et al., 2020), we know roughly when there were higher and lower lake levels in the past. However, these data rather show long-term tendencies, and the temporal resolution, especially during historic periods, is not always very clear. Therefore, we used geomorphologic methods with an exact comparison of profiles, sediment layers, and dated samples of former lake levels. Additionally, we checked the sediments of selected profiles for the ecologic significance of the included Cladocerans and diatoms (Figure 8, red lines).

At the north bank of the lake, within the dunes (Figure 6, lower part; cross section 3), we regularly found silty, partly clayey, and slightly organic sediments. These are supposed to be of Late Glacial origin and may represent very high meltwater levels within the subglacial channel of Lake Tresssee in this period. The first indication of the recent lake at the north bank of the basin is an erosion edge or beach barrier (site 4–1). It consists of probable reworked aeolian sands from the middle of the Boreal period (OSL: 8050 ± 690 BC), presumably from an active shore zone. This would point to a lake level of barely 27 m a.s.l. at this time. Site TS 2, located further down toward the lake basin, represents a littoral silting-up zone. The base of the sediment was dated to the seventh and eighth centuries AD. This indicates a lowering of the lake level during the Viking Age (8th century and later). The high algae content (different *Pediastrum* species) and the presence of reed taxa and *Nuphar* pollen and Nymphaeaceae idioblasts of the samples below (100 and 112 cm depth) confirm the presence of a shallow lake zone. At the same time, these taxa indicate the eutrophic status of the lake water during this period. The composition of pollen spectra is comparable with the spectra from the TS 3 core (Figure 3; Stolz et al., 2020) after the *Fagus* peak (Migration Period) and the start of the increase of the *Calluna* pollen curve. Human indicators of arable fields such as *Triticum* t. pollen are present, as well as indicators of pastoral activity (*Plantago lanceolata, Rumex acetosella*). No *Secale* pollen was found. The age of this deposit as an indication of a shrinking water level at least partly correlates with the results of Magny et al. (2003), even if a connection between these two factors is unlikely. According to Jäger (2002), the climate became drier in the two centuries before 1000 AD. European lakes predominately indicated lower lake levels during the first millennium AD.

The neighboring location of site TS 3 (Figure 1) dried out after 1960. This confirms that the lake level was slightly higher at the time than it is today. In this site, changes in the clay (still-water conditions), finest/fine/medium sand (input from dune sand and river sediments), and coarse sand (input from river sediments) contents are well visible for the last 6300 years, as well as some larger fluctuations (Figure 3).

At the south bank of the basin (in the Westerholz forest), we found confirmation of the lake conditions during the Late Bronze Age (1003–844 cal. BC; site 2–1) as well as from the 3rd century BC (Late Iron Age) until the 5th to 7th century AD (sites 2–1 and 2–3). The lake level must have been high in this time. Especially for the Hallstatt and Early La Tène, Magny et al. (2003) indicated higher water levels for European lakes and Haas et al. (1998) demonstrated a cold-humid phase for Switzerland. The upper part of the profiles there consists of a more terrestrial sediment, confirming a drying out of this location and further decreasing lake water level from the Viking Age onward. However, in the same location, tree roots were dated from the first and second centuries AD and from the 4th century AD, probably from trees growing near the shoreline (sites 2–1 and 2–2). This fact points to a lowering of the lake level in comparison to the Bronze and Iron Age levels (until approx. 200 BC) to at most 26.3 m a.s.l. However, we must also assume that it underwent short-term fluctuations during this period.

Further out, in the southwestern part of the basin, we confirmed a still relatively high lake level of at least 25.1 m a.s.l. in the High Middle Ages (11th to 13th centuries; site 1–2). In the sediment of this core, we found diatoms (mostly benthic forms) and Cladocerans (represented mostly by littoral, plant-associated taxa – *Acroperus harpae, Alonella nana, Camptocercus rectirostris, Eurycercus lamellatus*, and *Chydorus sphaericus*, and in the uppermost diagnosed level, also by planktonic *Bosmina longirostris*, which may indicate the slightly increased water level), with both groups pointing to lacustrine rather than terrestrial conditions.

A high lake level during the Bronze Age (1745–1610 cal. BC) can also be assumed from the age of the base of an alternating peaty stratum, below a thick Holocene slope colluvium at the outflow of Treene River from the lake basin (site TS 12; Figure 8).

Our results for the lake basin's outermost western rim (site TS 6) were not easy to interpret. Between the Bronze Age (1133–978 cal. BC) and the late Migration Period (644–714 cal. AD), the absence of typical lake indicator microorganisms in the sediment shows that this location was never permanently flooded. However, the sediment in the lower part of this core is similar to other cores from areas of the basin that contain well-proven lake sediment. The upper part (younger than the seventh or eighth centuries AD) was peaty and (semi-)terrestrial. Site TS 8, also located in the western part, proves lake conditions in the Roman Period (61 cal. BC-65 cal. AD) through the evidence of lake organisms in the core, but also points to a shrinking process and/or strong fluctuations until Viking Age (765–895 cal. AD). After this time (989–1052 cal. AD), the location of about 26 m a.s.l. felt completely dry.

Dark and strongly organic sediment at the outermost part of the small alluvial fan between 24.5 and 26 m a.s.l. (site 2–14) proved to be from the high medieval period (1258–1284 cal. AD). This suggests that the lake level was at least 25 m a.s.l. thereafter.

Alluvial fan growth and processes on slopes

The initial part of the alluvial fan had already formed during the end of the Younger Dryas (OSL: 9810 ± 700 BC; Figure 7). The organic content of the well-layered sediment is low, and the profile above shows soil formation. The middle and lower parts of the fan, in particular, must have had frequent contact with the lake during higher lake levels. This is clearly visible based on the comparison of all relevant profiles with different lake levels in Figure 8. From this we can see that the lake influenced the growth process of the alluvial fan, given that the latter is almost entirely located between the highest detected lake level and the recent lake level. Furthermore, from the Neolithic onward (3031–2907 cal. BC), the alluvial fan extended into the lake. Several thin, light-colored sand layers could be evidenced inside the fan. This also proves an aeolian deposition on the fan (see site 2–12; the sand layer is located between a late medieval (1388–1437 cal. AD) radiocarbon sample and an early modern (1470–1640 cal. AD) one. This fits well with the documented sand drifts in the area. To make a rough calculation of the growth process, we took the ages from the sites 2–12, 2–15, and 2–14. The results, given in Figure 10 (line 7), show an enhanced growth process starting in the Late Neolithic (lower than 1 mm/year). Clearly anthropogenic-influenced sedimentation processes started in the 9th century AD, a date that corresponds nicely with a resettlement phase in the region. The highest sedimentation rates happened during the late medieval period (about 3.8 mm/year) and in the 200 years thereafter (3.2 mm/year). These results fit well with those found for other parts of Central Europe concerning anthropogenically triggered soil erosion and its maximum in the high medieval and late medieval periods (Bork et al., 1998; Dotterweich, 2008; Dreibrodt et al., 2010; Lang, 2003; Schmitt et al., 2005; Stolz and Grunert, 2010).

At the outflow of Treene River from the lake basin we detected an enhanced slope activity, probably due to agricultural activities (e.g. Dotterweich, 2008) that date back to the Bronze Age (1745–1610 cal. BC). The sample originated from a Holocene colluvium from the slope foot and was presumably connected to the flood plain sedimentation of the Treene River and, with it, to the sedimentation of the lake. A further sample from the middle part of the slope indicates enhanced soil erosion after the sixth and seventh centuries AD. Both results fit well with the archaeological evidence from the surrounding areas. The influence of a former castle or similar object on the hill remains unclear. The hill itself is of natural origin.

Conclusion

Through the use of 30 new radiocarbon ages (plus 19 ages from Stolz et al., 2020) and 16 OSL ages, it was possible to create a new timeline for the interaction between the Lake Tresssee, an adjacent dune field, a small alluvial fan, and an adjacent slope systems in response to climatic and human influences since the Late Glacial Period. This is the most precisely dated inland dune chronology of Schleswig-Holstein and Southern Denmark in conjunction with the development of a lake basin and other sediment sinks. In all, seven different phases of dune activity were demonstrated for the last 13,000 years. The dunes arose in consequence of sand drifts from the Weichselian outwash plain and probably also from a local subglacial basin and local meltwater deposits from the Younger Dryas to the Early Boreal (OSL ages: 10,080 ± 870 BC until 8050 ± 690 BC; phase 1;). Further sediment drifts, probably due to water or surf effects in the intermeshing zone between the dunes and the lake, took place in the mid-Boreal (OSL age: 8050 ± 690 BC; phase 2). For the 8.2 event, a climatic influence can be presumed. During the Young Holocene, there were several further phases with sand drifts, mostly triggered by grazing and an enhanced land use (demonstrated by young sand covers and pollen results). These phases took place during the Late Bronze Age (OSL age 1130 ± 340 BC; phase 4), the Preroman Iron Age (^{14}C age: 400–351 cal. BC; phase 5), the High Middle Ages (^{14}C ages: 968–1046 cal. AD and 1160–1264 cal. AD; OSL age: 1100 ± 340; phase 6), and the early modern period (OSL age: 1550 ± 30 AD; phase 7). This is the first proof of sand drift events in the Bronze and the Iron Age for Schleswig-Holstein. Furthermore, remarkable sand records for the lake, especially during the Bronze Age (1800–500 BC), the Preroman Iron Age (500 BC-0), and the 12th and 13th century AD can be well confirmed. For a long time, Lake Tresssee was influenced by strong sediment impacts due to fluvial and aeolian processes. Changes in its water level can be attributed predominantly to the condition of the outflowing Treene River and its flood plain. An easily permeable outflow caused a decrease in lake levels, triggered by artificial interventions and – more rarely – climatic impacts like flush events must have been the triggers in the past. At the northern edge of the former lake basin, we presume that the highest water levels occurred during the Late Glacial Period as a consequence of melting ice (more than 27.5 m a.s.l.). The highest Holocene water level can last be presumed for the Boreal (27–27.5 m a.s.l.); thereafter, levels must have fluctuated in response to climatic and the first human effects. However, the lake level and its extension must have been generally high (26.5–27. m a.s.l.). The last time period for which we were able to prove this higher level was the Late Bronze Age (10th/9th century BC); thereafter, from the first to the 4th centuries AD, the lake's water levels clearly declined due to the presence of tree roots next to the southern edge of the lake basin starting at about this time. From the 5th to the 8th century AD, this process increased, and the outermost western part of the basin dried out. In High Middle Ages (11th to 13th centuries AD), the water level reached a minimum of 25.1 m a.s.l. During the 18th and 19th centuries, inhabitants reduced the lake area artificially to obtain new agricultural land. Today, the lake has its smallest extension ever recorded.

We detected that in the process of the growth of a small alluvial fan beside the dune field, the biggest sedimentation rates occurred between the 14th and 17th centuries. These results fit well with the results for other parts of Central Europe concerning anthropogenically triggered soil erosion.

At the outflow of Treene River from the lake basin, we detected an enhanced slope activity with the formation of colluvial sediments after the middle Bronze Age (1610–1745 cal. BC) and again after the sixth and seventh centuries AD.

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