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Onset of intense permafrost conditions in Northern Eurasia at ~2.55 Ma seen in a cryogenic weathering record from Lake El'gygytgyn

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Abstract

Physical weathering in permafrost landscapes contrasts to mid- and low latitude physical weathering in the way that quartz is less stable in freeze–thaw (F/T) cycles and breaks down to silt sized grains. F/T weathering also produces a distinct single quartz grain micromorphology such as microcracks and brittle textures of the surface. Both of these sediment-mineralogical features have been used for identifying intense permafrost conditions (i.e. intense F/T dynamics) around Lake El'gygytgyn in NE Eurasia using a Pliocene–Pleistocene sediment record from the centre of the lake. This lake provides the longest terrestrial palaeoenvironmental archive in the Arctic and has been a sediment trap for 3.6 Ma. Detritic material marked by F/T weathering becomes distinctive ∼2.55 Ma ago and has been accumulating since then in the lake basin. This time marker coincides with the establishment of a perennial lake ice cover and corresponds with pollen assemblages indicating a significant cooling during that time. It matches fairly well the timing of the Plio-/Pleistocene cooling known from other marine and terrestrial evidence.

The onset of intense quartz weathering is regarded as a first order age assessment for the beginning of persistent Quaternary permafrost conditions and which deepened into the ground since then in settings of high continentality in the non-glaciated NE Eurasian Arctic.

1 Introduction

In the Northern Hemisphere, the permafrost regions occupy approximately 24 % of the exposed land area, principally at high latitudes (Heginbottom et al., 2012). Large portions of the Arctic are undergoing permafrost warming and partial thawing (Osterkamp et al., 2005; Smith et al., 2005, 2010; Romanovsky et al., 2010a, b) and concerns about the effects of permafrost degradation in the North have attracted much attention recently. E.g. it is suspected that because of the great size and volatile nature of the carbon pool, which is contained in permafrost soil (Zimov et al., 2006; Waldrop et al., 2010), decomposition of previously frozen, old organic carbon is one of the most likely positive feedbacks to climate change in a warmer world, if it continues to thaw (Davidson and Janssens, 2006; Heimann and Reichstein, 2008). During IPY 2007/8 various initiatives in the world have banded together to conduct permafrost research, at least partially in the hope of minimising the remaining uncertainties about the effects of temperature forecasts and the impact on fragile polar environments (Romanovsky et al., 2010a).

Little attention is given though to the beginning of permafrost formation and the associated environmental and climate conditions. It is possible that some, perhaps most, permafrost had its origin at the beginning of the Pleistocene era, but this should not imply that the intervening thermal conditions were without significance (Lunardini, 1995). Insights into the permafrost history mostly rely on studying local sections that provide restricted time windows, mostly of the Quaternary (Schirrmeister et al., 2002; Froese et al., 2008; Wetterich et al., 2008; Lacelle et al., 2009; Meyer et al., 2010; Reyes et al., 2010; Kanevskiy et al., 2011; Vaks et al., 2013). However, because by definition permafrost is a temperature and not a stratigraphical phenomenon, drawing an inference about permafrost age is not straightforward from outcrop conditions. What is needed is an environmental archive that stretches beyond the beginning of permafrost formation and capturing the inception of frozen ground conditions. One more IPY emphasis was on past warm Earth analogues and discovering, how the environment behaved before the Quaternary glaciation cycles. Part of this effort to delve into the deep records of a previously warm Earth was coring and studying the El'gygytgyn Crater impact site (Melles et al., 2011, Fig. 1). The dry and cold conditions over at least most of the Plio–Pleistocene in Northern Siberia (Glushkova, 2001; Brigham-Grette, 2004; Hubberten et al., 2004) ensured that this terrestrial sediment trap has remained ice free (Fig. 1a), thus preserving a near continuous record of sediment and climate proxies in this area (Melles et al., 2012; Brigham-Grette et al., 2013). The basin has been created by a meteor impact 3.6 Million years (Ma) ago (Layer, 2000), this is just at the time when temperatures in the Arctic started to grow colder and drier (e.g. Shackleton and Opdyke, 1977; Thiede et al., 1998).

Hypothetically, lake sediments approximately resemble surface soil for the imported minerogenic detritus. The use of lake sediment sequences provides a better temporal resolution of significant events in palaeoenvironmental and palaeoclimate changes than could be achieved using a chronosequence approach. The El'gygytgyn site thus potentially has captured the onset of permafrost conditions in northern Eurasia given that suitable proxy data are available and which are closely connected to permafrost dynamics. Moreover, for the first time a time-continuous terrestrial archive can be used for tracing the permafrost history.

Identifying mineral detritus that resulted from frost weathering and accumulated in the Lake El'gygytgyn basin is purpose of this paper. It is a follow-up of a study where upper layers in the lake sediment column indicate intense permafrost weathering in the catchment for at least the last 220 000 a, the studied time interval (Schwamborn et al., 2012a). This interpretation is based on the inspection of quartz weathering features including (i) the qualitative assessment of single quartz grain shape and surface features stemming from freeze–thaw (F/T) stress that acts on the grains and is typical for permafrost soil and (ii) the semi-quantitative assessment of intense quartz grain break-up that leads to a quartz enrichment in the silt fraction following F/T cycling. It can be detected in the lake sediment column where the imported particles from the permafrost soil are time-continuously accumulating. The output of the present paper is a first order age assessment for the onset of intense permafrost conditions in the non-glaciated NE Eurasian Arctic and which deepened into the ground since then.

2 Cryogenic weathering

Permafrost archives can be studied to unravel weathering and transport history of grains. Studies of frost weathering consider the larger scale (m-range) frost wedging as well as the smaller scale (µm-to-cm-range) granular disintegration. Granular disintegration including small flaking is a process that leads to the production of fine debris (Konishchev, 1987; Matsuoka, 2001). The collection of field data on such factors as temperature or moisture that influence weathering is commonly at the macroscale. Decades ago there was the assumption weathering processes were related to atmospheric indicators (air temperature and annual precipitation) (Peltier, 1950). However, site conditions and nanoenvironments can be much more significant, especially in "extreme" environments like cold deserts where rock temperature and water supply vary greatly (André et al., 2004). It has been suggested that thermal fluctuations at the grain scale alone can potentially cause rock failure through such processes as thermal fatigue (Hall and André, 2003).

There is a strong association of seasonally driven F/T cycling with granular disintegration in cold regions. The repetition of F/T cycles is responsible for shallow but intensive fragmentation that produces fine-grained debris (Potts, 1970; Lautridou and Ozouf, 1982). Rise or fall in temperature across the 0 °C threshold is more meaningful, if it causes freezing of water or melting of ice (Konishchev and Rogov, 1993). Edge rounding of blocks can be due to granular disintegration under present cold-climate conditions and debris mantles on high ground have matrix-supported clasts where the underlying lithologies are particularly prone to granular disintegration (Ballantyne, 1998). These debris mantles are related to weathering, sorting and mass movement under periglacial condition.

Schwamborn et al. (2012a) have shown that for the sedimentological composition the destruction of quartz grains is a basic process during the formation of cryogenic (i.e. frost shattering including ice as a weathering agent) debris at the El'gygytgyn site. As established in experiments the cryogenic disintegration promotes relative quartz accumulation in the coarse silt fraction. Previously, Konishchev and Rogov (1993) reported that in all the size fractions investigated quartz grains proved to be less resistant as compared with the corresponding grain sizes of fresh feldspars and heavy minerals. This relationship is opposite to the behaviour of these minerals under temperate or warm climates (Nesbitt et al., 1997).

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The dominant role of quartz within detrital deposits thus determines the coarse to middle-sized silt fraction as the most typical product of cryogenic weathering. Single quartz grain shapes and grain surface microtextures illustrate how quartz grains are fragmented; they show sharp edges and microcracks, flaking and brittle surfaces occur frequently on the grains (Schwamborn et al., 2006).

Expressed in a so-called Cryogenic Weathering Index (CWI) the role of cryogenic weathering in frozen soil formation can be estimated according to Konishchev (1998):

$$CWI = (Q1/F1)/(Q2/F2)$$

where Q1 is quartz content (%) in the coarse silt fraction;
F1 is feldspar content (%) in the coarse silt fraction;
Q2 is quartz content (%) in the fine sand fraction;
F2 is feldspar content (%) in the fine sand fraction.

Values > 1 mean relative quartz grain enrichment in the fine fraction when compared with the coarse fraction and argue for cryogenic weathering influencing the grain size distribution. Indication of cryogenic weathering according to the CWI has already been implemented into permafrost modelling spanning the last 400 000 a (Romanovskii and Hubberten, 2001). In other studies it was essential to identify permafrost episodes in terrestrial Quaternary records from the European Russian Plain, Yakutia, and New Jersey (Konishchev, 1999; Demitroff et al., 2007; French et al., 2009).

3 The El'gygytgyn study site

The 18 km wide impact crater is placed in Cretaceous continental volcanics belonging to the Okhotsk–Chukotka volcanic belt (Ispolatov et al., 2004; Stone et al., 2009). They consist mainly of rhyolithic tuffs and ignimbrites framing the 12 km wide lake to more than 75 % of the perimeter (Fig. 1b). The hills on the crater rim rise to between 600 and 930 m above sea level (a.s.l.), and the lake level is 492 m a.s.l. with a bowl-shaped lake basin that has a water depth of 170 m at maximum. Today's catchment-to-lake

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ratio is only 1.6 : 1 and fifty seasonally active inlet streams drain the crater slopes and loose Quaternary deposits cover the crater plain surrounding the lake especially in the north and west (Fedorov et al., 2013). There is no inlet to the lake from the outside of the crater; thus, remote sources of detritic load can be excluded. The active layer (i.e. seasonally frozen soil layer) is about 0.4 m deep in peaty silts and can reach 0.5–0.8 m in sand and gravels on the slopes (Schwamborn et al., 2006). The site is in the continuous permafrost zone with a MAAT (mean annual air temperature) of −10 °C at 3 m above the ground (Nolan and Brigham-Grette, 2007). The MAGT (mean annual ground temperature) below the area of seasonal temperature cycling is −6.1 °C at 20 m depth with a total permafrost thickness around the lake between 330 and 360 m (Mottaghy et al., 2013).

A tandem drilling at Lake El'gygytgyn has provided two cores from in- and outside the lake (Fig. 1c); a 315 m long sediment record and nearly 200 m from the underlying bedrock were recovered from the centre of the lake (core 5011-1, Melles et al., 2012; Brigham-Grette et al., 2013; Raschke et al., 2013). A 141 m long fully frozen core was extracted from permafrost soil next to the lake (core 5011-3, Schwamborn et al., 2012b, 2013). Lz1024 is a 16 m long percussion piston core taken in 2003 (Juschus et al., 2009) that fills the stratigraphic gap between the lake sediment surface and the top of drill core 5011-1.

Methods

For CWI analysis 51 samples have been taken from Lake El'gygytgyn sediments recovered in lake core 5011-1. These samples cover core depths from 16.82 to 231.27 m. They build on 37 samples of core Lz1024 from the upper 12 m of the lake fill, which have been analysed previously (Schwamborn et al., 2012a). Whereas Lz1024 samples represent 4 cm thickness of carefully selected sediment layers, 5011-1 samples are mixed samples from one-meter core intervals; here, after liner opening and the compression release the samples have been extracted from the expanding sediment in the half-open liner tubes. One individual mixed sample of core 5011-1 represents a random mixture of sediment layers spanning 20 to 30 ka in the upper core part (0–160 m core depth) and about 10 ka in the lower core part (160–318 m core depth) (Nowaczyk et al., 2013). The 5011-1 mixed samples have been separated into the grain size fractions 32–63 µm and 63–125 µm in a vibrating sieve tower.

Downcore changes of the relative amounts of quartz and feldspar in the two grain size fractions have been assessed in a semi-quantitative way using X-ray diffractometry (XRD). XRD instrumental settings were chosen according to Petschick et al. (1996) and non-textured pulverised samples have been measured in a Philips PW1820 goniometer applying CoKα radiation (40 kV, 40 mA). Feldspar was measured for reference as another ubiquitous mineral, which was found to be less susceptible to F/T weathering than quartz in Siberian samples (Konishchev and Rogov, 1993). CWI values were calculated based on the quartz occurrence that is defined using the area below the 4.26 Å peak, whereas feldspar is measured by summing the 3.24 + 3.18 Å peak areas (Vogt et al., 2001).

The quartz grain micromorphology (single grain shape and surface features) were examined and imaged in detail using an SEM (scanning electron microscope). As demonstrated by work summarised in Krinsley and Doornkamp (1973) and Mahaney (2002) it is mostly the quartz grain morphology that is preferred to obtain a qualitative assessment of weathering and transport traces and to aid the identification of sedimentary environments. Samples of two presumably contrasting weathering regimes have been selected; the first SEM sample set is from core depth 318 to 350 m representing early basin and lake deposition 3.6 Ma ago (Raschke et al., 2013; Brigham-Grette et al., 2013). The selection of this sample set was guided by the assumption that they represent a Pliocene weathering regime before the establishment of permafrost conditions. A second set of samples has been taken from core depth 12 to 0 m with material that is presumably affected by cold-climate weathering in the catchment during the Late Quaternary glaciation cycles.

Grains were washed for 10 min in $SnCl_2$ (5 %) to remove iron before they were cleaned with distilled water and doused in an ultrasonic bath (2 min) to remove adher-

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ing particles. Finally they were boiled in ethanol for 5 min before being washed again (Schirrmeister, 1995). Quartz grains were selected randomly under a binocular microscope; a group of 20 grains per sample was mounted on aluminium stubs and coated with gold-palladium. The grains were then examined and photographed in detail using an SEM (Ultra 55 Plus Carl Zeiss SMT).

4 Results

4.1 Sedimentology and stratigraphy of lake sediment core 5011-1

The details about the sedimentology and stratigraphy in core 5011-1 have been covered elsewhere (Melles et al., 2012; Brigham-Grette et al., 2013) and are summarised below:

The lake core 5011-1 consists of clayey to silty layers with a minor sand portion (Francke et al., 2013). This is also true for the core Lz1024 (Juschus et al., 2009). A succession of four main facies types have been identified: facies A consists of dark gray to black silt and clay showing sub-millimetre laminations and is linked to glacial/stadial conditions, facies B has olive-gray to brown massive to faintly banded silt and is linked to moderate interglacial/interstadial conditions, facies C consists of reddish brown silt with pale white laminae and is linked to an exceptionally warm interglacial climate, facies D is characterised by turbidites and other forms of mass movement deposits. Whereas episodes of facies B (warm) and facies D (turbiditic) sediments can be found along the whole lake sediment core spanning the last 3.6 Ma, facies C (exceptionally warm) occurs first at 140 m core depth (~ 2.88 Ma) and facies A (cold) first appears at 124 m core depth (~ 2.60 Ma).

4.2 SEM features of quartz grains in early basin and lake deposits

Between 350 and 318 m core depth the 5011-1 material is interpreted to represent mostly suevite with a minor portion of welded ignimbrite at 337 to 333 m depth

(Raschke et al., 2013). Between 328 and 316 m depth the sequence consists of reworked suevite with a minor portion of intercalated sand layers representing a transitional zone to the overlying lake deposits. Single quartz grains taken from the sand layers have mostly angular to subangular shapes with high relief and sharp edges (Fig. 2a–h). Rounding effects due to transportation are minor. Clean grain surfaces illustrate a lack of chemical weathering or mechanical abrasion (e.g. Fig. 2b and e) and conchoidal fracture planes (e.g. Fig. 2a and d) indicate that the particles likely have resulted from fragmentation from greater host rock particles (Fig. 2b and d). Occasionally grains show linear to slightly curved cracks that are a few micrometers in size (Fig. 2i–k). Because of their linear shape on smooth surfaces they may be the result of the rock shattering during the impact event; shock metamorphism has been observed in the intercalated suevitic deposits (Raschke et al., 2013).

4.3 SEM features of quartz grains in upper lake sediments

In the core part 12 to 0 m quartz grains are slightly more rounded (Fig. 3a, b, f). Partly they have subangular outlines and sharp edges (Fig. 3c). Frequently surfaces are scarred with various microcracks (Fig. 3c) that have irregular courses in a zigzag (Fig. 3a) or dendritic way (Fig. 3b). The microcracking produces brittle textures and fragmented surfaces (Fig. 3a, d, f). This internal grain fracturing may initiate grain fragmentation and produce silt particles (Fig. 3e).

4.4 CWI as a semi-quantitive measure of quartz grain break-up

CWI values obtained from core 5011-1 range between 1.6 and 0.4 with a mean of 1.0 (Fig. 4). A scattering of values around 1.0 – this is the reference that separates distinct cryogenic quartz grain weathering with values > 1 from non-cryogenic weathering with values < 1 is apparent. A trend from lower values to higher values is obvious though when tracing the CWI values upcore. To illustrate this increase a linear trend has been calculated for the sample set (5011-1 and Lz1024 samples) having a determination

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coefficient R^2 of 0.468 (Fig. 4). This linear trend intersects the reference value of 1.0 at 121.6 m core depth. Based on the 5011-1 age model (Nowaczyk et al., 2013) this core depth is equal to an age of 2.55 Ma. CWI values from core Lz1024 covering the upper 12 m are mostly > 1; they range between 1.0 and 2.2 with a mean of 1.5 (Schwamborn et al., 2012a).

5 Discussion

5.1 Evidence of permafrost formation

When combining insights from the quartz grain micromorphology, the quartz grain break-up and changes of the relative quartz enrichment in the silt fraction a distinct change of environmental conditions between the early and the late lake period becomes obvious. Quartz grain micromorphology changes from angular to subangular shapes with fresh surfaces in the early basin deposits to slightly more rounded grains with internally fragmented grains and brittle surfaces in the upper lake sediments. Permafrost soils in the catchment of Lake El'gygytgyn are known to produce grains with microcracks and brittle surfaces (Schwamborn et al., 2006). The quartz grain features are inherited from catchment related weathering processes and form especially in the active layer of the subaerial slope deposits. These grain features survive the transport way (max. 3–4 km) to the deep lake basin of Lake El'gygytgyn, where grains accumulate over time (Schwamborn et al., 2012a). The grain surface textures appear to be particularly diagnostic for a permafrost environment, since their production can be directly linked to the destructive effect of F/T alternation in the active layer. Comparable textures associated with mechanical encroachment are only sparsely documented but appear distinctive for permafrost soils (Kowalkowski and Mycielska-Dowgiallo, 1985; Konishchev and Rogov, 1993). They are not known from any other environment, i.e. a glacial, fluvial, or aeolian environment (Krinsley and Doornkamp, 1973; Mahaney, 2002).

The mechanism of cryogenic disintegration has been assigned to a wedging effect when ice forms in microcracks and produces volume widening following repeated F/T as discussed in Konishchev and Rogov (1993). Another specific cause for widening is seen in the freezing of gas-liquid inclusions, commonly containing salts (Rogov, 1987). Fluid inclusions are a common feature of quartz minerals and are known to modify quartz grain shapes and microstructures during weak plastic deformation (Tarantola et al., 2009). Based on TEM (transmission electron microscopy) imagery of El'gygytgyn quartz grains it has been suspected that these inhomogeneities and other in-grain impurities may be weak zones where quartz grains can break apart when physical stress such as F/T and seasonal temperature cycling is applied to them (Schwamborn et al., 2012a).

Surface chipping and flake generation illustrate that silt sized grains are produced. It links the SEM features to the CWI signal and could explain the relative quartz enrichment in the silt fraction given that F/T dynamics have fragmented the quartz grains. The CWI signal is an in-situ signal of cold-climate physical weathering (Konishchev, 1982) and since the grains at Lake El'gygytgyn are transported off-site on the slopes, through the inlets, and into the lake, the lake sediment column looses the in-situ information, but preserves an integral catchment signal rather than an immediate microenvironment. It is noteworthy that CWI values in core Lz1024 are more consistent with almost all of them placed in the indicative zone of intense cryogenic weathering. These samples are from distinct sediment layers, whereas core 5011-1 samples combine a random mixture of sediments reflecting glacial, moderate, and exceptionally warm environments and, in addition, mass movements, which can occur along one individual core meter in 5011-1. Sauerbrey et al. (2013) calculated that 33 % of the Quaternary sediment record consists of mass movement deposits (facies D in Fig. 4). Especially these deposits, which include turbidites, are suspected to alter the mineralogical composition due to additional hydrodynamic sorting during deposition. This may have led to blur the signal to some extent. Still, the upcore increasing CWI values in 5011-1 have captured the transition from non-cryogenic to cryogenic weathering conditions around the

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lake. If accepting the timing of 2.55 Ma based on the entry of the linear trend into the window that is indicative for intense cryogenic weathering, this would match the timing of the northern cooling at the Plio–Pleistocene boundary (2.58 Ma). It may thus serve as a first-order assessment for the onset of permafrost conditions as reflected in the cryogenic weathering record and which is further discussed below. However, presumably the establishment of permafrost in the area occurred gradually. From a biological viewpoint permafrost is expected already at 3.3 Ma in the area based on findings of the permafrost tolerant *Larix* in the lake sediments of that time (Brigham-Grette et al., 2013). Between ca. 3.305–3.275 Ma treeless tundra- and steppe-like habitats became dominant in the regional vegetation. Climate conditions were similar to those of Late Pleistocene cold intervals (Andreev et al., 2013). The CWI record has its first value > 1 at about 3.1 Ma and matches this vegetation interpretation to some extent. On the other hand, CWI values are < 1 between 2.1 and 1.7 Ma and indicate non-cryogenic conditions in the catchment during the period. This contrasts the palaeoenvironmental reconstruction for that time, which points to pervasive glacial episodes at Lake El'gygytgyn that gradually increase in frequency from ~ 2.3 to ~ 1.8 Ma (Melles et al., 2012, facies A in Fig. 4). This mismatch in interpretation suggests that single CWI values should not be overstated; rather the overall temporal trend in the CWI data set should be used for interpretation.

5.2 Inception of icehouse conditions on the Northern Hemisphere

When considering the onset of cold-climate conditions elsewhere in the Northern Hemisphere there is direct observation of sedimentary sections with ice in the ground in non-glacial environments or indirect observation from lake archives that recorded distinct climate cooling. More indirect evidence is available from marine cores, which reflect glacial history as IRD (ice rafted detritus) or ice volume change proxy data. In high northern latitudes, permafrost, or possibly most of it, probably started to develop at the beginning of the Pleistocene era in the periglacial realm of glaciers and ice caps (e.g. Harris, 1994; Murton, 2005). During the Pliocene temperatures were ~ 3° to 6 °C

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warmer than in modern times (Lunardini, 1995; Ravelo et al., 2004; Brigham-Grette et al., 2013), before temperatures were greatly reduced and massive glaciations occurred in the Quaternary period. Permafrost likely started to form at many northern places during that time. At some places permafrost formation might be older though, as is outlined below and summarised in Fig. 5.

5.2.1 Terrestrial evidence

At Lake El'gygytgyn the first cold "glacial" sediment facies A occurs at 2.602 Ma (Melles et al., 2012, Fig. 4); this facies, indicative of a perennial summer lake ice cover and a mean annual air temperatures at least 3.3 (±0.9) °C colder than today (Nolan, 2013), becomes more common after 2.3 Ma. The CWI linear trend enters the window that specifies intense cryogenic weathering at ~ 2.55 Ma before present and, thus matches well the onset of cold-climate conditions around the lake inferred from other climate proxies. For example the vegetation reconstruction based on pollen in core 5011-1 shows that at the beginning of the Pleistocene ~ 2.6 Ma ago, a noticeable climatic deterioration occurred; forested habitats changed to predominantly treeless and shrubby environments and cold steppe habitats, which reflect a relatively cold and dry climate (Andreev et al., 2013).

In the northern lowlands of the Chukotka Peninsula sediment strata are exposed (the "Ryveemsker Series") that have ice wedge casts reaching into underlying fluvial diamicton of Upper Pliocene age. The pollen composition in the upper diamicton layers indicates a diminishing portion of tree pollen and a growing portion of hypo-arctic tundra vegetation. Presumably soon after that climate deterioration the ice wedges started to form and a low temperature regime allowed the establishment of permafrost (Arkhangelov et al., 1985). In northern Yakutia syngenetic ice-wedges in the Yana–Indigirka region have existed since the Middle Pleistocene continuously based on the time of sediment deposition, which is dated back to that time (Konishchev, 1999). Pseudomorphic (postsedimentary) ice wedge casts start to appear already during the Plio–Pleistocene transition of this sediment section. Uncovered buried ice wedges in the

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Indigirka River lowland are reported to date to the early Pleistocene and the same age is reported for cryogeneous disturbances, which have been observed with numerous wedge-shaped structures in a group of strata in the Kolyma lowland (Arkhangelov and Sher, 1973).

The first cold interval in Lake Baikal sediments starts at 2.8 Ma expressed as an "early glacial diatom minimum", which is accompanied by an "early glacial clay maximum" of the same age (Prokopenko et al., 2001). Vorobieva et al. (1995) deduced considerable climate change at the Early Pliocene–Late Pliocene boundary (3.2 Ma) from palaeopedological studies in the Baikal area, and cooling during the Late Pliocene with first occurrence of permafrost gley about 2 Ma ago. Also at 2.8 Ma a large climatic shift from long-lasting warm-humid conditions to large-amplitude cold-dry and warm-humid fluctuations are seen in the Chinese loess sequences (Yang and Ding, 2010).

In areas that were covered by the Pleistocene ice sheets, thermal models suggest that preexisting permafrost and ground ice would have disappeared from beneath warm sectors at the base of the ice sheets (Humlum et al., 2003). Thus, in areas that were glaciated, much of the permafrost and ground ice has developed in post-glacial time (during the late Pleistocene and Holocene Epochs). In non-glaciated parts of northern Russia, Alaska, and Yukon, permafrost and ground ice may be much older. In northwestern Canada, for example, some ground ice is ~ 740 000 a old and because it is found in warm permafrost of the discontinuous zone this raises questions regarding the resilience of permafrost through sustained periods of warmer-than-present climate (Froese et al., 2008), and, in Svalbard, permafrost in the high mountains is thought to be perhaps as much as 800 000 a old or even older (Humlum et al., 2003).

In Northern America a cooling in the Western Cordillera that suggests a high-latitude and high altitude glaciation, which presumably was accompanied by permafrost formation, began already in the Miocene and Pliocene. In southern Alaska tillites have been found that are 3.6 and 2.7 Ma old (Denton and Armstrong, 1969). McDougall and Wensink (1966) reported a till overlying a basalt in northeastern Iceland dated at 3.1 Ma and suggested that a glaciation occurred at this time.

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5.2.2 Marine evidence

Indirect evidence for permafrost conditions comes from the glaciation history recorded in marine sediments if, land-based glaciations have created a periglacial belt with permafrost conditions in the continental terrain. The Arctic Ocean has maintained perennial sea ice since 47 Ma (Darby, 2008; John, 2008; Moran et al., 2006; Stickley et al., 2009) and an intensification of the Northern Hemisphere Glaciation in Late Pliocene time is clearly documented in marine sediment cores (e.g. Thiede et al., 1998; Jansen et al., 2000; Haug et al., 2005). Major cooling took place between 3.1 and 2.5 Ma and Maslin et al. (1998) suggest that the onset of glaciation in the Eurasian Arctic occurred at about 2.74 Ma, 40 ka prior to that in Alaska (2.70 Ma) and 200 ka prior to that in northeast America (2.54 Ma). Böse et al. (2012) note that IRD records from the Nordic Seas show a stepwise inception of large-scale glaciation near the North Atlantic region during the Plio–Pleistocene. The first major influx of IRD, derived from the Greenland ice sheet, occurs at 3.3 Ma with further IRD fluxes recognised between 3.1–2.9 Ma (Jansen et al., 2000). These major IRD events have been detected within the Eastern Equatorial Pacific and suggest a strong Northern Hemisphere glacial signal (Prueher and Rea, 1998). Records of IRD from around the North Atlantic/Barents Sea at 2.74 Ma (Knies et al., 2009) and of alkenone and diatom oxygen isotope ratios from the North Pacific (Haug et al., 2005) indicate the presence of glacial margins reaching tidewater. Evidence consists of the first appearance of ice-rafted erratic clasts from northern Britain within cores from the upper Hebridean Slope, plus IRD horizons from the Porcupine Seabight along the Irish margin; records from the Irish margin demonstrate 16 major IRD events between ~ 2.6 and 1.7 Ma (Thierens et al., 2012). Bailey et al. (2012) have identified iceberg calving from multiple circum-North Atlantic terranes (i.e. Greenland, Scandinavian, N-American and possibly British) for IRD in the sub-polar North Atlantic dated around 2.5 Ma. Periglacial belts with permafrost are assumed for those areas.

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6 Conclusions

The history of cryogenic weathering is linked to the long-term climate evolution and parallels the glaciation history; in the sediment record of Lake El'gygytgyn traces of cryogenic weathering are absent in the Pliocene core part and become distinct in the Pleistocene part of the core.

Cryogenic conditions around the lake are established at ~ 2.55 Ma based on the quartz grain fragmentation and enrichment in the silt fraction of the basin fill after that time.

Even though this permafrost time marker is a first order age estimate it matches well the beginning of cold-climate sedimentation in the lake and is in line with the timing of northern hemispherical cooling known from other marine and continental records of the Arctic.

A continental setting in the NE Eurasian Arctic, which escaped Quaternary glaciations, was favourable to expose the continental terrain to the full magnitude of seasonal temperature cycling. This also includes multiple transitions between the water and ice phase in the soil, which promote the quartz grain shattering. Quartz grain micromorphology such as microcracking and brittle grain surfaces supports the interpretation of intense mechanical stress in this environment.

A fairly uniform geology, a small catchment-to-lake ratio, and the preservation of the weathering features during the transport on the slopes were favourable conditions to yield an integral signal of the cryogenic weathering dynamics. These environmental conditions might be representative for wider areas of the continental North.

Likely permafrost in the area is as old as the onset of the intense cryogenic weathering and has deepened into the ground since the Plio-/Pleistocene transition.

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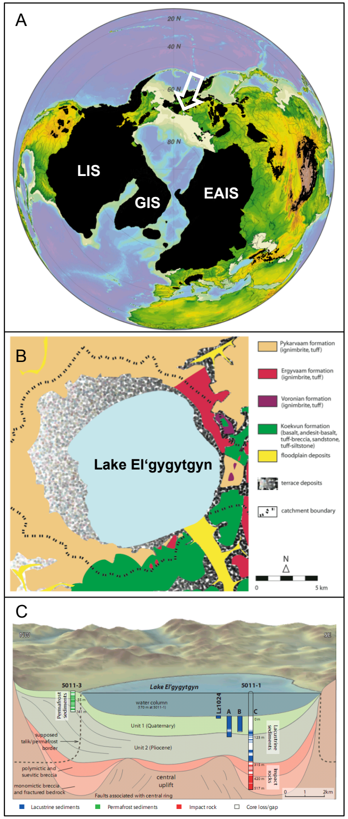

Fig. 1. (A) The maximum extent of Cenozoic northern glaciation (dark areas); LIS = Laurentide Ice Sheet, GIS = Greenland Ice Sheet, EAIS = Eurasian Ice Sheet (based on Ehlers and Gibbard, 2007). Lake El'gygytgyn is located in the non-glaciated uplands of the Chukotka mountain belt (see arrow). **(B)** Geological setting around the lake (based on Nowaczyk et al., 2002). **(C)** Schematic cross section of the El'gygytgyn basin showing the location of sites Lz1024, 5011-1, and 5011-3 (Melles et al., 2011).

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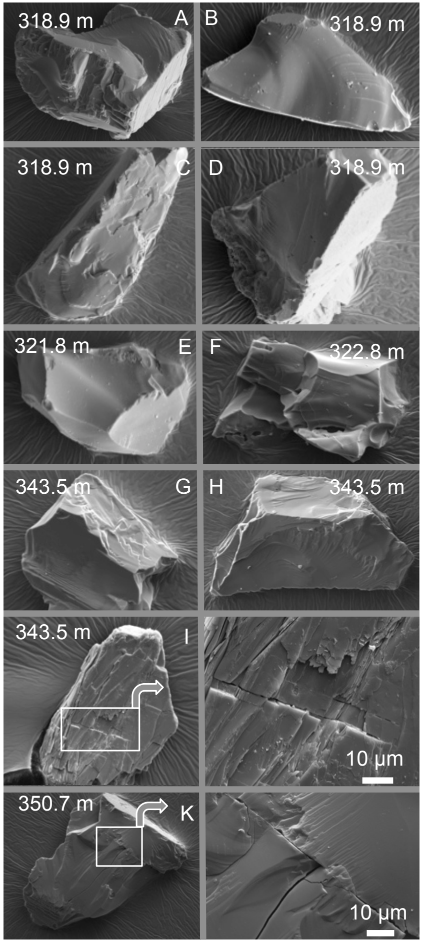

Fig. 2. SEM imagery of quartz grains (63–125 µm fraction) from lower lake sediments including depth information.

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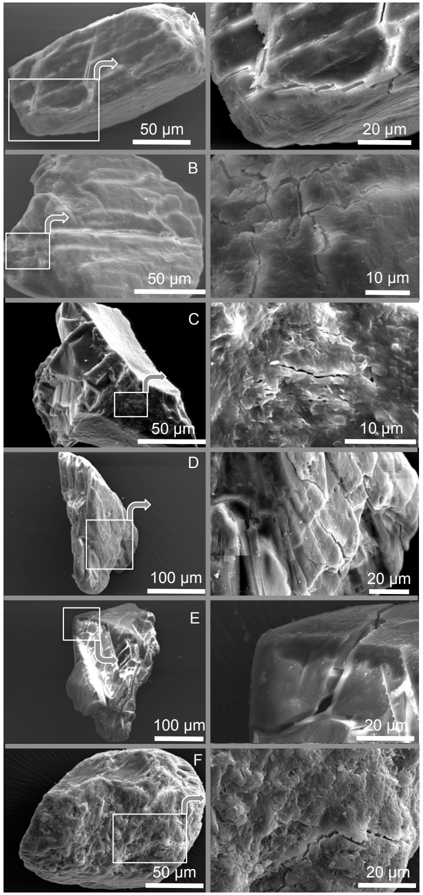

Fig. 3. SEM imagery with zoomed features of quartz grains (63–125 µm fraction) from topmost 12 m of lake sediments.

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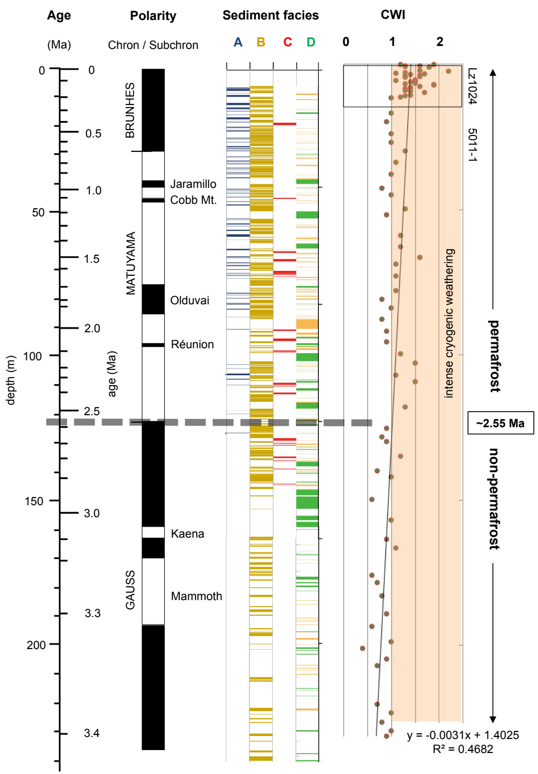

Fig. 4. Interpretation of lake core sedimentology based on Melles et al. (2012) and Brigham-Grette et al. (2013): facies A = glacial/stadial conditions, B = moderate/warm conditions, C = exceptionally warm, D = turbidites (orange) and other mass movement deposits (green) and the cryogenic weathering index (CWI) measured on cores Lz1024 and 5011-1 including a linear trend. The coloured background marks the area, which is indicative for intense cryogenic weathering. Geomagnetic polarity adopted from Haltia and Nowaczyk (2013).

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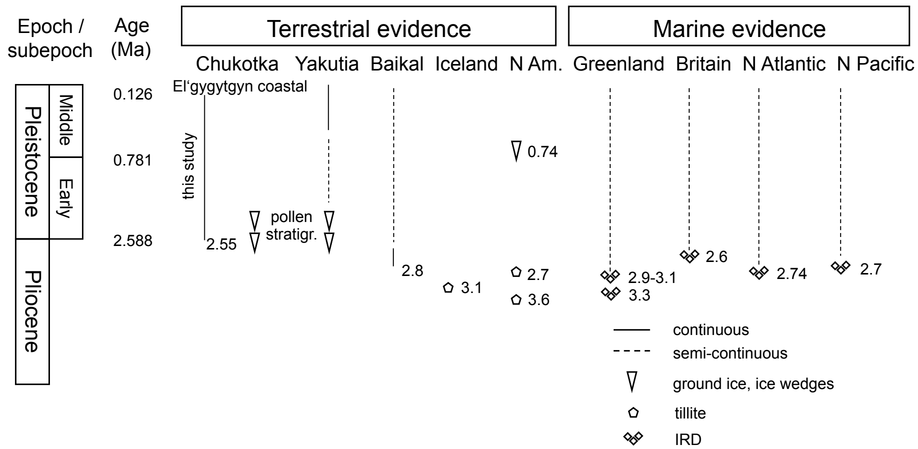

Fig. 5. Maximum ages of Plio–Pleistocene cooling in Northern Hemisphere continental terrains based on terrestrial and marine evidence. References are given in the text. Timescale is according to Gibbard et al. (2010).

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