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Hydrological and sedimentary controls over fluvial thermal erosion, the Lena River, central Yakutia

Nikita I. Tananaev *

P.I. Melnikov Permafrost Institute, Siberian Branch, Russian Academy of Sciences, 36, Merzlotnaya Str., Yakutsk, Sakha (Yakutia) Republic 663200, Russia
Institut National Politechnique de Toulouse (INPT), 6 allee Emile Monso, BP 34038, 31029 Toulouse cedex 4, France

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ABSTRACT

Water regime and sedimentary features of the middle Lena River reach near Yakutsk, central Yakutia, were studied to assess their control over fluvial thermal erosion. The Lena River floodplain in the studied reach has complex structure and embodies multiple levels varying in height and origin. Two key sites, corresponding to high and medium floodplain levels, were surveyed in 2008 to describe major sedimentary units and properties of bank material. Three units are present in both profiles, corresponding to topsoil, overbank (cohesive), and channel fill (noncohesive) deposits. Thermoerosional activity is mostly confined to a basal layer of frozen channel fill deposits and in general occurs within a certain water level interval. Magnitude-frequency analysis of water level data from Tabaga gauging station shows that a single interval can be deemed responsible for the initiation of thermal action and development of thermoerosional notches. This interval corresponds to the discharges between 21,000 and 31,000 $m^3 s^{-1}$, observed normally during spring meltwater peak and summer floods. Competence of fluvial thermal erosion depends on the height of floodplain level being eroded, as it acts preferentially in high floodplain banks. In medium floodplain banks, thermal erosion during spring flood is constrained by insufficient bank height, and erosion is essentially mechanical during summer flood season. Bank retreat rate is argued to be positively linked with bank height under periglacial conditions.

1. Introduction

Riverbanks are dynamic interfaces between fluvial, atmospheric, and soil domains where each medium contributes to their transformation. Stream power is a major force, exerting action on the banks, while resistive properties of bank material restrict fluvial action. Hydraulic erosion rates are controlled by stream power and shear stress values along the eroding bank (Nanson and Hickin, 1986; Darby and Thorne, 1997), position of the eroding segment within the channel section, defining flow 'angle of attack', and effectiveness of shear stress application (Shur et al., 1978; Are, 1983; Nanson and Hickin, 1986), lithology and cohesive properties of overbank sediment (Shur et al., 1978; Julian and Torres, 2006; Parker et al., 2008), bank height (Berkovich and Vlasov, 1982; Nanson and Hickin, 1986), and vegetation (Thorne, 1990; Millar, 2000). In permafrost areas, direct ice impact, solifluction, thaw slumps, detachment slides, and needle ice formation also contribute to the frozen banks' instability and collapse (Prowse and Culp, 2003; Lawler, 2006; Lipowski and Huscroft, 2007).

Fluvial thermal erosion is virtually omnipresent in periglacial environment but is best perceivable along the banks of large alluvial rivers. It is active mostly during a short spring flood period when the streams undercut their frozen banks, forming spectacular thermoerosional niches (Walker and Hudson, 2003). Inception of these niches is a juxtaposition of thermal and hydraulic action and, as such, represents the essence of fluvial thermal erosion. Observations in Arctic Alaska show that mechanical washout generally proceeds slower than thaw, except in the apexes of thermoerosional niches where these processes are assumed to be in equilibrium (Scott, 1978). Heat transfer rate controls particle detachment, thus overriding purely hydraulic impact (Shur et al., 1978; Randriamazaoro et al., 2007). Noncohesive bank material with massive cryogenic texture ('ice cement') and low ice content is more susceptible to thermal erosion than cohesive deposits or organic material having higher ice content (Scott, 1978; Gautier and Costard, 2000; Dupeyrat et al., 2011). Block slumping occurs after its flexural resistance had been exceeded either because of excessive undercutting or active layer thickening.

Preceding quantitative studies of fluvial thermal erosion concentrated on observing and prediction of the process rate given the sediment ice content and water temperature (Randriamazaoro et al., 2007; Dupeyrat et al., 2011; Debolskaya, 2014) and on field and remote observations of bank erosion (Are, 1983; Costard et al., 2003, 2007, 2014). Little notion was given to explain the factors promoting the formation of thermoerosional niches. Notably, however, floodplain sediment heterogeneity and water stage variations discourage their development. Position of the notch formed by thermal erosion is

* Bldg. 8A, 1st District, Igarka, Krasnoyarsk Krai 663200, Russia.
E-mail address: nikita.tananaev@gmail.com.

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important, as it defines implicitly the notch depth as well as volume of slumped material and bank retreat rate (Scott, 1978).

Scott (1978) and, later, Costard et al. (2003) assumed that the thermoerosional niche development is concentrated at the base of the cohesive layers where the bank material is more susceptible to erosion. However, this observation is eventually phenomenalistic and cannot be generalized, i.e., over cases where the banks are uniform and no such boundary is present. The research rationale behind the present paper is an assumption that hydrological controls are somehow responsible for notch inception and thermoerosional niche development.

Effective (dominant, channel-forming) discharge concept is used extensively to evaluate the long-term competence of the streamflow in shaping channels and controlling their hydraulic geometry (Wolman and Miller, 1960; Alabyan and Chalov, 1998; Doyle et al., 2005; Caissie, 2006). Effective discharge is frequently associated with that at the bankfull stage, though it may be significantly lower if considering sediment transport efficiency (Benson and Thomas, 1966) or rivers in degradational mode (Hassan et al., 2014).

The application of effective discharge concept to fluvial thermal erosion in periglacial rivers has several limitations. Bankfull discharge has limited competence in affecting bank erosion insofar as this process occurs normally at discharges well below bankfull level (Wolman, 1959). Complexity of floodplain structure, consisting of multiple levels, complicates the identification of a single discharge causing overbank spill (Nanson and Croke, 1992; Gautier and Costard, 2000). Moreover, recurrent ice jams are known to disrupt a steady stage–discharge relationship during the break-up period (Prowse and Culp, 2003). Stage fluctuations up to several metres are related to variations in flow resistance and hydraulic roughness and can be caused solely by changes in ice conditions (Zaitsev et al., 2006).

This study is based on data collected during multiple field campaigns in the middle Lena River section adjacent to the city of Yakutsk, central Yakutia, between years 2002 and 2008. This river section was subject to previous regional studies in regard to channel pattern development and sedimentary features (Zaitsev and Chalov, 1989; Gautier and Costard, 2000; Costard et al., 2003; Degtyarev et al., 2007), as well as recent climate shift (Costard et al., 2007) and ice breakup and spring flood (Costard et al., 2014) as affecting the thermal erosion process. However, hydrological controls over the latter remain largely understudied; hence the present paper is aimed at partially closing this gap. The Lena River floodplain in the studied reach has complex structure and embodies multiple levels varying in height and origin: high, medium, and low inundation plains (Gautier and Costard, 2000). Performance of fluvial thermal erosion is expected to vary between these levels, and this variability in relation to water regime is the major subject of this study.

First, sedimentary features of the overbank deposits at two representative key sites, corresponding to distinct floodplain levels, are presented. Second, the 'magnitude-frequency' approach is used to evaluate the effective water stage (or, stages, if multiple) responsible for preferential inception of thermoerosional notches and niches. Finally, these results are overlapped to infer the differences in effectiveness of fluvial thermal erosion in riverbanks of various heights.

2. Study area

Field studies were carried out within the 20-km section of the middle Lena River in the vicinity of Yakutsk, central Yakutia (Fig. 1). At an upstream limit of the studied section, a 120-m-high Tabaginsky Mys terrace (an outcrop of the Jurassic (J_2) sandstones) narrows the Lena River valley from the west. Farther downstream, the main channel approaches the 30-m-high alluvial Bestyakh terrace (experiencing intense fluvial erosion during spring and summer floods) and heads toward Yakutsk after a gentle left turn. The width of the Lena River valley is between 5 and 6 km in the upper section and increases to 15–20 km farther downstream. Active channel deformations occur within 8 to 10 km of the valley width and are also limited by the Tabaginsky Mys terrace outcrop in the upstream section.

The Lena River channel pattern in the studied section is anabranching, after Lewin and Ashworth (2014). The main channel is relatively straight during floods, but its sinuosity increases with decline in water stage. During flow recession, submerged side bars are exposed and start controlling the low-flow channel pattern. A braided pattern emerges within relatively straight sections, where central bars are present. Stability of sand bars is augmented by presence of perennially frozen deposits at their base (Tananaev, 2013). Floodplain is present on both sides of the channel except the Bestyakh terrace cross section and has a well-developed network of the highly sinuous secondary branches, through which the water excess is flushed to the main channel after the flood peak. Floodplain to bankfull channel width ratio is quite low (≤3), reflecting the limited ability of the floodplain to convey flood water and to store the overbank deposits. The vertical structure of the floodplain is complex, with at least three distinct levels, as referenced by Gautier and Costard (2000). Contemporary cryogenic processes are active within the highest floodplain levels and include frost heave and polygonal ice-wedge growth.

Hydrological features of the Lena River were recently described in several papers (Yang et al., 2002; Ye et al., 2003; Berezovskaya et al., 2005; Dzhamalov et al., 2012), and specific attention was paid to the ice break period and associated ice jams (Zaitsev et al., 2006; Kilmjaninov, 2007; Costard et al., 2014). Hence, only a brief hydrological overview is given here based on data from 1938 to 2013, published by Russian Hydrometeorological Agency.

Lena River at Tabaga gauging station (GS; see Fig. 1 for gauge location reference) drains a catchment of 897,000 km^2, which contributes about 42% of total basin flow at the outlet (Ye et al., 2003); mean annual discharge equals 7270 m^3 s^{-1} for the 1938–2013 period. Recent changes in streamflow include significant increase in winter discharges over the last 25 years, accompanied by a visible increase in annual-average streamflow (Fig. 2). Cumulative duration of flood events, i.e., number of days with daily discharge exceeding 25,000 m^3 s^{-1}, reached its maximum in 2012 (63 days), but no clear trend emerges from these data. Spring floods of the 1930s and 1940s were retarded, the fact that can give an impression of a substantial increase in flood severity to what is essentially a 'low base' effect. Peak discharges remain at about the same level as in mid-1950s and throughout the 1960s (Fig. 2).

Water regime of the Lena River is dominated by snowmelt, although heavy rains in the mountainous headwaters can produce storm events comparable to spring flood in terms of peak discharges (Fig. 3). Distinct winter low-flow period, with daily discharges below 3000 m^3 s^{-1}, lasts for 198 days, early November until early May, and has an average discharge of 1520 m^3 s^{-1}. With air temperatures frequently hitting the −50 °C mark during winter seasons, ice thickness in main channels reaches 1.32 m on average and normally exceeds 2.0 m in secondary branches. Rapid (in 10 to 15 days) discharge increase, originating from snowmelt runoff, is fed by the Lena River and its major tributaries, the Vitim and the Olekma rivers. It occurs between late April and late May, frequently accompanied by ice jams (Zaitsev et al., 2006).

Spring flood duration varies between 54 and 96 days depending on snow abundance and insolation of the mountainous headwaters in the Lena River basin. Local meltwater sources are considered to be negligible, as central Yakutia receives on average 120 mm of solid precipitation, which evaporates partially during seasonal transition to positive air temperatures. Spring peak discharge averages 36,500 m^3 s^{-1} and can exceed 50,000 m^3 s^{-1} during extreme flood events. Post-peak flow recession is gradual, and continues from mid-July until early November, when the river again enters the wintry dormant state. Summer low-flow periods are frequently interrupted by rain floods originating from the Vitim and Olekma basins. Rain-induced peak discharges can exceed those of the preceding spring floods.

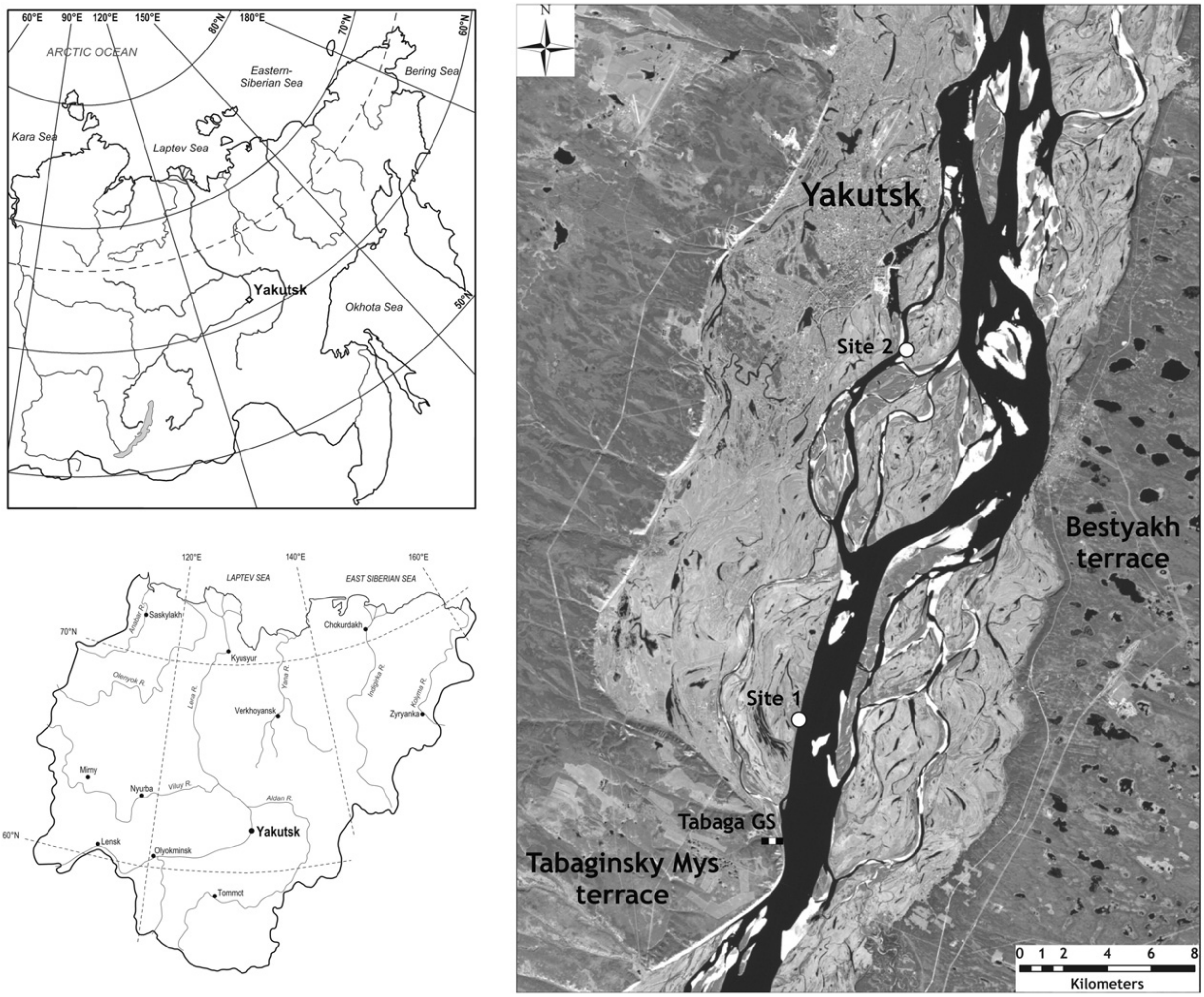

Fig. 1. Geographical position of the studied section of the Lena River near Yakutsk (left) and its detailed overview (right); Landsat 8 image, acquisition date: 25 September 2015.

3. Materials and methods

Fluvial thermal erosion is an ablation-controlled process, and undercutting intensity depends more on the ablation rate than on the stream power (Scott, 1978; Dupeyrat et al., 2011). Under periglacial conditions, the initial ablation rate is low and it can take significant time, i.e., several days, until the ablation accelerates (Randriamazaoro et al., 2007). This requires the water stage, and not discharge, to be stable during at least several days in order to promote thermal bank erosion. We can assume, therefore, that stream competence in shaping riverbanks can be expressed through a certain magnitude-frequency product of water level.

Effective water level, in regard to the joint action of thermal (ablation) and mechanical (particle release and removal) erosion, can be obtained as a certain product of water level and frequency in a manner consistent with suggestions of Makkaveev (1955), reviewed in (Alabyan and Chalov, 1998), and those of Wolman and Miller (1960). Observed range of water levels, derived from the long-term data set, is split into i bins (classes), then effective water level H_{eff} is a class interval, corresponding to a local maximum (maxima) of an efficiency index:

$$H_{eff} \sim \max\left[(Hvd)_i\right] \tag{1}$$

H_i, mean water level of the ith bin (m asl); v, average cross-sectional flow velocity at H_i, m s^{-1} and d, empirical probability density of water level occurrence for the ith bin. The flow velocity is regarded in the presented analysis solely as a scaling metric, attenuating the level magnitude and frequency by its relative physical impact, i.e., stream ability to remove and displace detached sediment particles. In practice, H_i is plotted against the corresponding $(Hvd)_i$ value for each bin. The resulting curve has its local minima and maxima, and the latter correspond to a certain water level, which is called 'effective' (Alabyan and Chalov, 1998).

The present study employs daily water level data for selected years between 1942 and 1989 (23 years in total), published by Russian Hydrometeorological Agency, for the long-term statistical analysis. This data set could not be expanded beyond 1989. Some fragmentary data are available for the most recent period of 2002–2012. These data cover only the summer months (normally June to early October) and are as such incompliant with long-term data. However, we count the employed data set as sufficiently representative for the average hydrological conditions, under which fluvial thermal erosion usually occurs. Mean annual discharge (7160 m^3 s^{-1}), mean peak discharge (35,200 m^3 s^{-1}), and number of days with discharges above 25,000 m^3 s^{-1} (24), obtained for this data set, do not fall far from their long-term (1938–2013) averages.

Water stage data for the Lena River at Tabaga for 2008 were used to compare the hydrological conditions of the field campaign period with the long-term averages. These data were provided by Lena Basin Waterways & Navigation Authority (Yakutsk, Russia) and cover the period from 1 May to 21 August 2008.

Water level data in Russia are published as relative values, in centimetres above the reference height (m asl), assigned independently for each gauging station. Hence, water level data were converted from

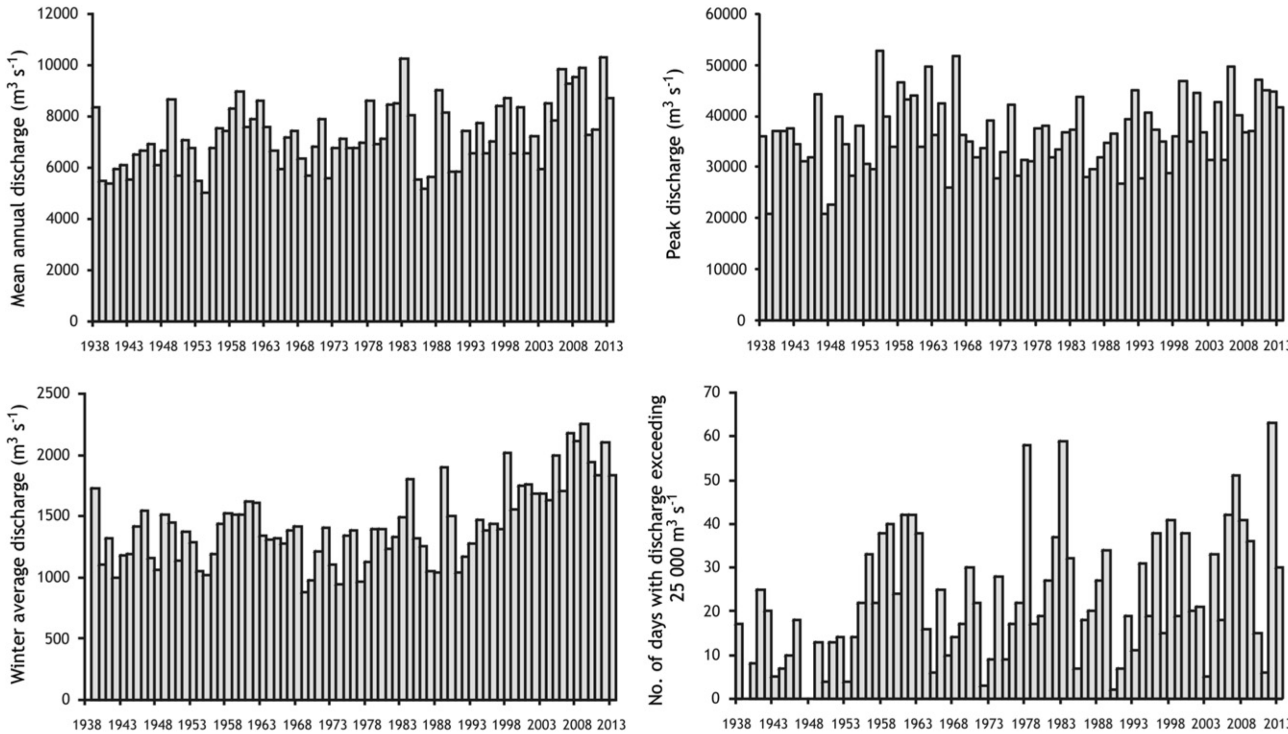

Fig. 2. Water regime of the Lena River at Tabaga GS, 1938–2013.

relative to absolute prior to the analysis, using a reference height of the Tabaga GS, 85.08 m asl.

Average cross-sectional flow velocity (v, m s^{-1}) data have the same source as water level data. The relationship between observed H and v values is steady and linear (Fig. 4). It was used to evaluate v_i for each H_i using the linear equation:

$$v_i = 0.136\ \ H_i\text{–}11.3, \qquad (2)$$

making the assessed product $(Hvd)_i$ eventually bivariate, but still 'weighted' by a fluvial action parameter.

Frequency statistics were assessed using RStudio (2014), an open-source integrated development environment (IDE) for R language, with the standard *hist*() method. Bin (class) width h was estimated from sample size n using *nclass.Scott*() function following an approximation known as Scott's (1979) rule:

$$h = 3.5sn^{-1/3} \qquad (3)$$

where s is the standard deviation of the sample and n, sample size. Derivation of the bin width using a strict statistical relation was important, as previous works (Makkaveev, 1955; Alabyan and Chalov, 1998) allow substantial uncertainty in its assignment. Table 1 summarizes the statistical features of long-term and 2008 data sets.

In 2008, bank exposures at two locations were surveyed, referred here as sites 1 and 2 (see Fig. 1 for spatial reference). Selection of the key locations followed the floodplain structure, as described by Zaitsev and Chalov (1989) and Gautier and Costard (2000), i.e., comprising three major levels: high, medium and low. Lowest floodplain level was either inundated or showed no visible bank erosion during our field campaigns. Hence, two key sites were selected representative for high and medium floodplain levels. Site selection was based on field experience, satellite imagery, and available topographic maps of the Lena River valley.

Bluff zones and slopes of the banks at both key sites were sampled for grain-size analysis. Grain-size distribution was determined by automated dry sieving using a Fritsch Analysette 3 shaker, with subsequent weighing of the sieved subsamples. At the time of the analysis, sieves of

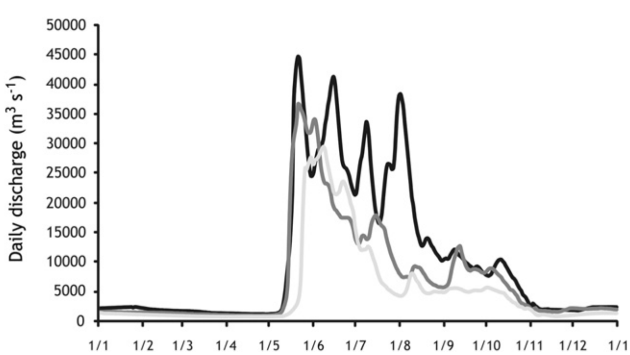

Fig. 3. Observed hydrographs of the Lena River at Tabaga GS; bold, 2012 (high flow); dark grey, 2002 (average flow); light grey, 1986 (low flow).

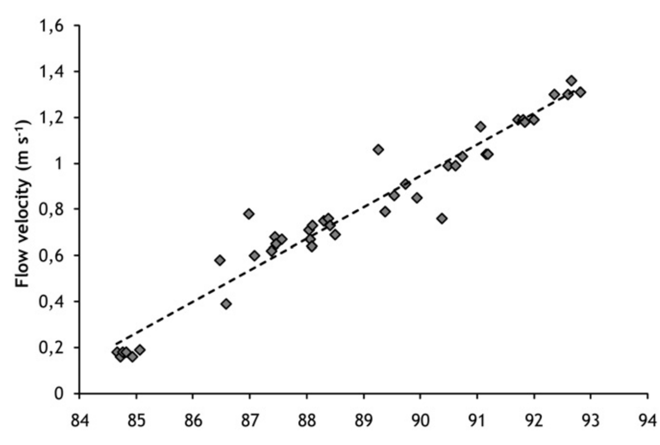

Fig. 4. Average cross-sectional flow velocity related to the water level of the Lena River at Tabaga.

Table 1
Statistical features of the employed hydrological data sets.

	H_{min}	H_{max}	H_{mean}	n	s	Bins
Long-term	84.05	95.63	87.48	8403	2.32	30
2008	85.76	93.99	89.94	100	2.08	6

five mesh sizes were available: 1000, 500, 250, 100, and 50 μm. In total, 14 samples were processed, of which 10 were from site 1 and 4 from site 2. Moreover, extensive field descriptions of the Lena River banks were conveyed by the author during the field campaigns from 2002 through 2011, mostly within two weeks after the spring flood peak.

The Circumpolar Active Layer Monitoring (CALM) project database, hosted by George Washington University web site, was used to acquire active layer depth data from CALM R42 (Tuymada) plot for 2008–2014. This plot is nonstandard and reports the active layer depths captured by thaw tubes on a random grid rather than standard (rectangular plot of 121 points). However, because of its location on the alluvial surface of the second terrace within the Lena River valley as well as the resemblance in sedimentary (sands, loamy sands) and landcover (meadow) features, it can be considered representative at least for the high floodplain surfaces. Active layer measurements are in general preformed at the time of maximum ground thaw, which is around late September to early October. Our field campaigns, however, were over by late August, while the active layer was not fully developed; and this is why CALM data are used instead.

4. Results

4.1. Sedimentary units

Bank exposures at two key sites (see Fig. 1 for spatial reference) were surveyed and sampled for grain-size analysis in 2008.

Site 1 represents an eroded section of the high floodplain level with surface altitude between 94.3 and 94.5 m asl (Fig. 5A). It is exposed to main channel flow of the Lena River and is separated from the first alluvial terrace by a minor intermittent branch. Bank height is ca. 8 m above the low-flow water stage and was ca. 3.8 m on the survey date (8 August 2008). The eroded bank has a bluff zone of about 2 m height, followed by a gentle slope toward the water edge. The inflection point between the bluff and the slope roughly corresponds to the visible boundary between dark-coloured cohesive material and light-coloured sands. At the time of survey, the adjacent bank segment was locally covered with vegetation and root patches of previously slumped block (Fig. 5B). Bank material was thaw at the surface, but frozen sands were found to be underlying the sandy slope. Active layer depth was about 1.1 m beneath the inflection point and decreased to 0.25 m right beneath the water's edge (Fig. 6).

Site 2 represents a medium floodplain level, with surface altitude between 90.0 and 90.5 m asl (Fig. 7). It is eroded by one of the secondary Lena River branches. Bank height is ca. 7 m above the low-flow water stage, and was ca. 3.5 m on the survey date (21 August 2008). Bank morphology at this location resembles that of site 1, although its offset from the main channel leads to lesser bluff zone height (ca. 1.5 m) and steeper (>50°) successive slope toward the channel. In the lower part of the slope, two minor steps were observed, originating from wind- and vessel-induced waves (Fig. 8). Evidences of thermoerosional niching and significant bank collapse were not observed in course of the survey neither in the surveyed location, nor within the adjacent 1-km section.

Grain-size characteristics appear to be similar at both studied locations, adjusted for the distance from the main channel (Table 3). Hence finer-grained material is deposited at site 2, situated in the secondary channel system. Grain-size analysis results resemble in general the previous quantitative descriptions (Gautier and Costard, 2000). Three distinct sedimentary units can be distinguished visually, varying in colour and texture.

Unit A at both sites represents the topsoil, upper 0.7 to 0.8 m of the profile beneath the rootmat, and consists of dark-yellow fine sands with minor silty layers at site 1 (median diameter d_m = 240 μm), shifting to sandy loams at site 2 (d_m = 160 μm). This unit is associated with the most recent overbank deposition, partially influenced by ice-jam damming and subsequent spilling (Gautier and Costard, 2000). *Unit B* is a thick (0.8 to 1.0 m) layer of dark-coloured, organic-rich loams and sandy loams (d_m = 110…130 μm) clearly of overbank origin. Frost weathering could also contribute to the material fining and increase the content of fines (<50 μm) to 30–50% (Fig. 9). The boundary between the two upper layers is sharp and wavy, wave height not exceeding 5 cm. Light-yellow fine and medium (d_m = 230…310 μm) cross-bedded sands are exposed in the basal part of the banks at both sites (*unit C*). At site 1 this layer is interbedded by several thin (<4 cm) organic layers, which were not subsampled for analysis. The boundary between this basal layer and the overlying overbank deposits is sharp and smooth.

Cumulative curves show all samples falling into two groups, roughly corresponding to overbank (unit B) and channel fill (unit C) deposits (Fig. 9). Typical overbank deposits of unit B are poorly sorted (*So* = 4.1…4.2), with d_{50} between 25 and 50 μm. Channel fill deposits of unit C are well sorted (*So* = 1.8…1.9), with d_{50} between 100 and 210 μm and visible cross-bedded stratification. Topsoil (unit A) stratum position within this classification is variable and reflects the complexity of contemporary deposition patterns. At site 1, adjacent to the main channel, unit A resembles well sorted (*So* = 2.2) channel fill deposits. At site

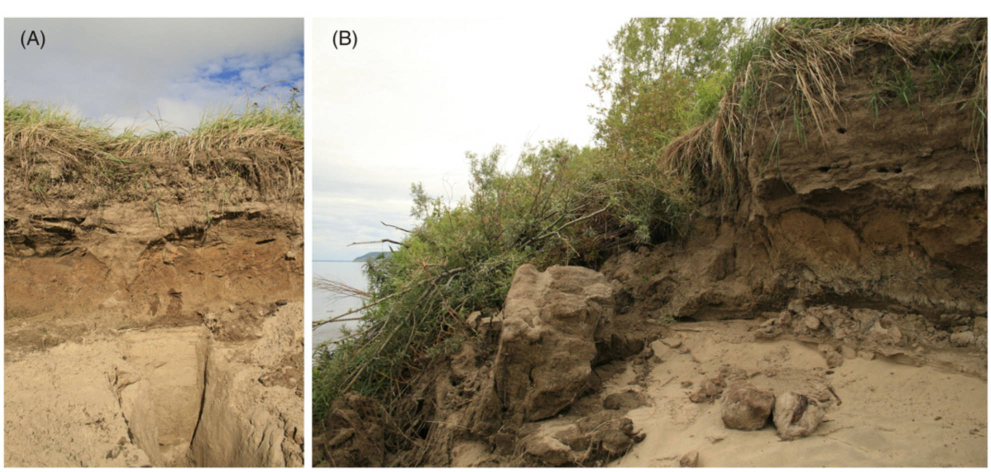

Fig. 5. Bank profile photo, site 1 (A); adjacent bank segment looking upstream (B). Numerous remnants of slumped blocks are visible on the background.

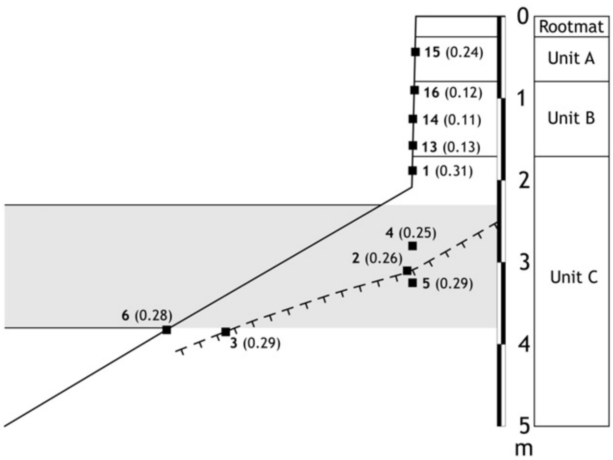

Fig. 6. Bank profile, site 1. Sample numbers, bold; median grain diameter d_m, in parentheses; permafrost table, dashed line; effective water level range, shaded.

2, located in a secondary branch system, this unit occupies the intermediate position between distinctly different overbank and channel deposits (Fig. 9).

4.2. Effective water level calculations

The Lena River, from a hydrological point of view, rests in a dormant state under the ice cover for the largest part of the year. Winter low-flow period lasts for seven to eight months, largely contributing to the most frequently observed water stages ranging from 85 to 87 m asl (Fig. 10A). Occurrence frequency decreases gradually toward higher stages without significant peaks or elevated plateaus. High floodplain levels above 94 m asl, i.e. at site 1, are inundated only on rare occasions.

Effective stage calculations reflect the same hydrological pattern (Fig. 10B). Peak 1, a major maximum of the $(Hvd)_i$ product, corresponds to an interval between 86 and 87 m asl, observed during summer and early winter low-flow periods. Other minor peaks follow: peak 2, occupying an interval around 88.1 m asl, which can be attributed to the falling limb of spring and summer floods. A major peak in the upper part of the graph, peak 3, covers the water level range from 91.05 to 92.62 m asl. This interval roughly corresponds to discharges from 21,000 to 31,000 $m^3\ s^{-1}$ (Table 2). Relevant discharge values were obtained from the stage-discharge curve for Tabaga gauging station. This upper peak reflects the ultimate importance of the spring period, as well as extremely high rain floods, in shaping the Lena River banks.

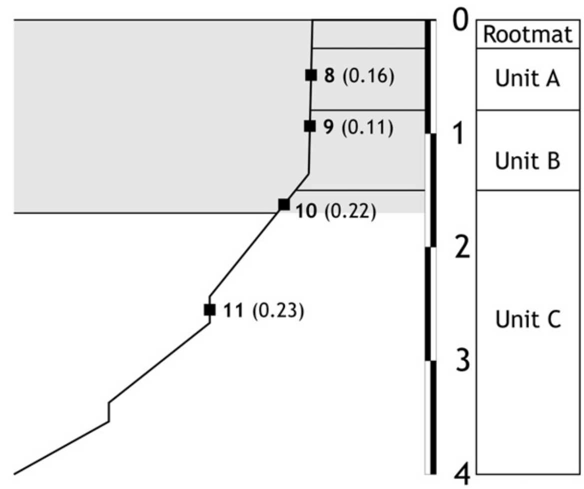

Fig. 8. Bank profile, site 2. Sample numbers, bold; median grain diameter d_m, in parentheses; effective water level range, shaded.

5. Discussion

5.1. Long-term vs. 2008 water stages

Field surveys and sampling were performed in 2008. Effective water level was evaluated using data for this particular year in order to verify whether the observed bank erosion patterns are in line with the average long-term conditions. Dataset was separated into six classes using Eq. (3). Number of classes is significantly less than for the long-term period because of smaller sample size.

Two distinct peaks present on the frequency histogram (Fig. 11A). Since the data only covers a period from 1 May to 21 August, winter low-flow peak is absent. The upper peak is by far more important in exerting control over bank erosion (Fig. 11B), and its range is similar to that of a long-term period (Table 2). We can thus conclude that the bank erosion patterns observed in the field are (in general) converging to those of a long-term period.

5.2. Site 1: high floodplain, thermal erosion

Thermoerosional niching, as observed during field surveys, is omnipresent within this floodplain level. Notches are developed mostly within the sedimentary unit C (channel fill deposits). The observed position of the niches ranged about the effective water level interval, that at site 1 falls between 2.3 and 3.8 m from the floodplain surface (Fig. 6).

Fig. 7. Bank profile at site 2 (A); adjacent bank segment looking downstream (B).

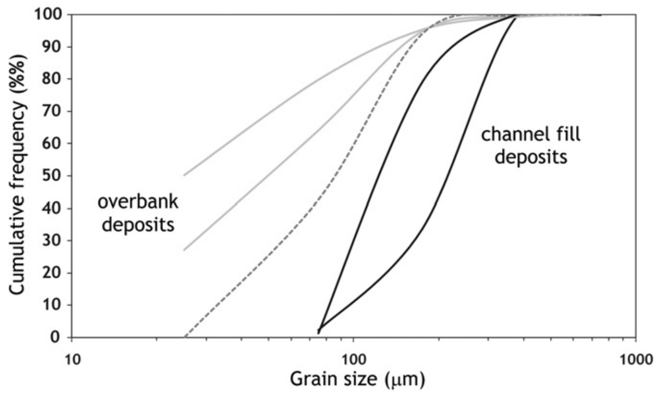

Fig. 9. Representative cumulative grain size distributions of major sedimentary units observed at the key sites: channel fill (dark grey); overbank (light grey); intermediate (dashed).

Channel fill deposits are highly susceptible to thermal erosion. Lower soil humidity and ice content in sands promote rapid ablation and particle detachment. Low cohesion, in its turn, allows particles to be immediately removed by the stream. No field evidence was found that the niche formation corresponds to the boundary between channel and overbank deposits, as assumed by Costard et al. (2003). Only on the later stages of this process, when the sand layer above the developing niche is thawed and desiccated, does it crumble down leaving cohesive loams exposed at the niche top. Observing the niches at this stage can lead to conclusions similar to those drawn by Costard et al. (2003).

During the spring flood, when the high floodplain banks are essentially frozen, their erosion is driven solely by thermal impact. The stream begins undercutting the banks and the notches are incepted when the water level is fluctuating within its effective interval. Notch depth rarely exceeds 1.5 to 3.0 m. Bank block slumps owing to gravity after its flexural resistance threshold had been exceeded because of active layer development or excessive undercutting. Collapsed material is deposited at the bank base and is further eroded by the stream. New niches are being formed in the background if hydrological conditions are favourable.

High summer temperatures, frequently above +30 °C, are usual for central Yakutia. Hence, bank sidewalls are thawed during summer, and bank erosion loses its thermal component. Mechanical washout during summer flood events effectively removes slumped blocks and freshly thawed material, and the stream is progressively undercutting the bank. Whence the rate of mechanical erosion exceeds the rate of thawing of the bank sidewall, the notch apex reaches the thawing front and contacts the frozen deposits. From that point and further on, hydraulic action is switching to ablation mode, and thermoerosional niches are formed during summer floods. This process can be repetitive, retaining its thermal component, throughout the summer flood season.

Average active layer depth in central Yakutia, as illustrated by CALM R42 plot data (Fig. 12), is around 2.0 m. Channel fill (unit C) deposits are only affected by heat influx from the exposure sidewall and not from the floodplain surface. This assures that the thawing front in the sidewall zone is always inclined toward the channel, as we have observed at site 1 (Fig. 6).

5.3. Site 2: medium floodplain, mechanical washout

Medium floodplain levels remain mostly inundated during spring flood peak. Effective water level corresponds to the topsoil stratum (unit A) or the overbank deposits (unit B) of the bank cross section (Fig. 8).

On the onset of spring flood, shallow active layer is already developed in the rootmat and topsoil (unit A) of the medium floodplain, therefore fluvial erosion lacks the thermal component. Shallow thermoerosional niches are formed within the overbank deposits layer (unit B). Cohesive properties and/or higher ice content of overbank deposits constrain accelerated ablation and limit fluvial thermal erosion rates. Notch development is limited by the flexural resistance of the block above the innermost point, or apex, of the thermoerosional niche. Slumped material, which includes turf and vegetated rootmat, shields the collapsed section from the fluvial action, preventing further erosion.

During extreme summer floods, when the water level is back to its effective range, these bank sections are mostly thaw, and purely mechanical washout of the bank material takes place. The upper strata of the medium floodplain level fall within the active layer, thus bank erosion progresses in thawed material above the thawing front.

5.4. Fluvial thermal erosion and bank height

Bank height was previously regarded as being inversely related to bank retreat rates, explicitly (Berkovich and Vlasov, 1982) and implicitly (Nanson and Hickin, 1986). Our results show, though qualitatively, that in permafrost areas with active thermal erosion, bank height should be regarded as positively linked to its retreat rate. Several lines of evidence can support this suggestion.

Scarce direct observations, performed previously within the river reach in question, suggest the enormous difference in retreat rates between the floodplain banks. For the high floodplain section several kilometres downstream of Yakutsk, annual retreat rates up to 50 to 60 m were reported by Chistyakov (1952). This is well above the average values, between 7.5 and 8.5 m, for all floodplain levels, measured in the field by Are (1983) for the adjacent Lena River section farther downstream, and by Shur et al. (1978) for the lower Indigirka River near Chokurdakh.

From the sedimentary perspective, higher banks provide higher probability of effective water level alignment with channel fill deposits. This highly erodible, noncohesive material is susceptible to fluvial thermal erosion through mechanical washout and ablation processes, or thawing of the textural ice (so-called 'ice cement'). This consideration is fully applicable solely to the river sections resembling those described

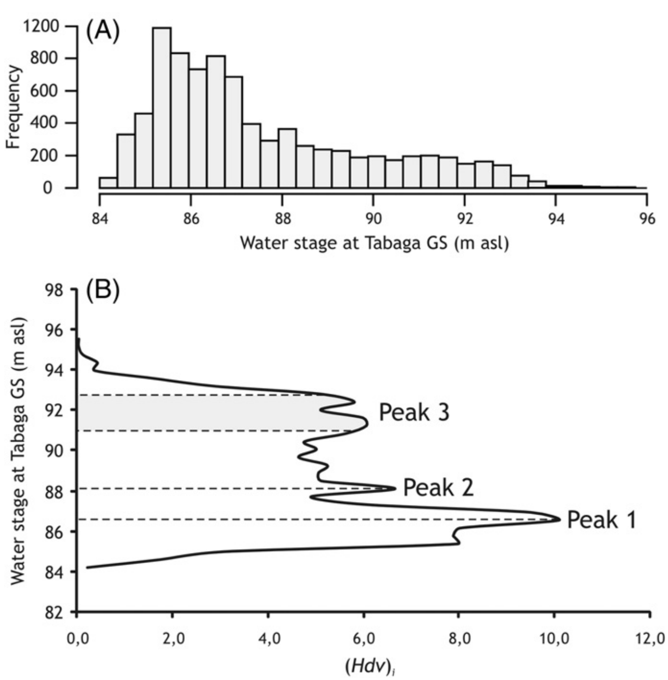

Fig. 10. Long-term data set: water level frequency distribution (A); effective water level calculation from plotted efficiency index $(Hvd)_i$ (B).

Table 2
Effective water stage estimates for the long-term and 2008 data sets.

		Bin medians, m asl	Stage range, m asl	Discharge range, $m^3 s^{-1}$
Long-term	Peak 1	86.55–86.94	86.35–87.13	4800–6200
	Peak 2	88.11	87.92–88.30	8000–9150
	Peak 3	91.25–92.42	91.05–92.62	21,000–31,000
2008	Peak 1	89.19	88.50–89.85	9600–14,600
	Peak 2	91.94	91.25–92.63	22,000–31,000

in this paper; in valleys with a different accumulation history this may not be the case.

Cryolithological evidence assumes that higher banks allow the flow to contact with frozen bank material for a longer time, insofar as active layer propagation down the bank profile is time-consuming. When the bank is sufficiently high, its lowest sedimentary units are only affected by solar radiation through sidewall heating but not by active layer development. Higher banks allow deeper thermoerosional undercutting as the frozen material above the notch has higher flexural resistance. Subsequently, size of collapsed bank sections is larger if the bank is higher, and this increases the retreat rate and the talus volume.

5.5. Bank erosion in periglacial environment: fluvial vs. thermal

In permafrost areas, rivers erode their banks largely in the same manner as their counterparts do elsewhere in the world. Periglacial river banks are not affected exclusively by fluvial thermal erosion, and whence the bank starts being eroded, it is not necessarily owing to fluvial and thermal erosion at once. It requires a thermal component (ablation) to exist and to precede mechanical washout.

Periglacial rivers are in a stagnant state, in respect to geomorphic action, during the largest part of the year (Fig. 13). Woken up by the intensive meltwater release, they start their work in eroding their channels (which are thawed) and banks (which are mostly frozen). Water level rapidly reaches its peak and continues with a gradual decline. Most of the fluvial thermal erosion occurs within the falling limb of a flood, as ablation requires several days to accelerate (Randriamazaoro et al., 2007). In high floodplain banks, fluvial thermal erosion takes place, and thermoerosional niches are formed. Lower floodplain levels are either inundated or thawed, though limited thermal action can still be observed.

After the meltwater pulse declines, rivers start reworking the lower portions of their banks, including collapsed material. Rain events can raise the water level to its effective value. At the initial stage of these events, however, bank material is thawed because of solar heating of the sidewalls. Fluvial thermal erosion is then possible when and if either the thawing front is above the notch apex or the active layer does not propagate below the effective water level interval.

Relative contributions of the mechanical and thermal components of bank erosion vary on a seasonal basis (Fig. 13). This fact clouds the discussion on the rates of fluvial thermal erosion, as not all bank erosion in permafrost is thermal, and on the comparison of bank retreat rates in periglacial and nonperiglacial environments. *De facto*, only bulk erosion rates are observed and thus are legitimate for direct comparison, and no conclusions can be drawn concerning higher effectiveness of thermal erosion compared to its mechanical counterpart. Ablation rate observations in experimental settings under variable conditions (water temperature, ice content, flow velocity, etc) solely allow such conclusions and are an important part of future research.

5.6. Fluvial thermal erosion: research perspectives

Fluvial thermal erosion is a well-described process acting in permafrost areas, yet these descriptions are mostly qualitative. Field and experimental studies stress that the knowledge of physical properties of bank material is critical to our quantitative description of this process. Determining the physical and mechanical properties of frozen material is an uneasy task, but it is an essential part of future research. Major parameters to be assessed include, but are not limited to, grain-size distribution, bulk density, porosity, friction angle, effective cohesion, and ice content. Along with these laboratory tests, thorough field descriptions of eroded sections are required, noting the type of surface being eroded (floodplain, terrace), bank angle in a bluff zone and in

Table 3
Grain-size distribution of the samples collected at the key sites, in %.

Sample #	500–1000 μm	250–500 μm	100–250 μm	50–100 μm	<50 μm	d_m, μm
Site 1						
1	0.85	65.43	31.42	2.31	–	0.31
2	0.02	42.26	57.07	0.65	–	0.26
3	0.03	59.17	40.08	–	–	0.29
4	0.16	35.83	62.32	1.69	–	0.25
5	0.04	56.83	41.87	1.25	–	0.29
6	0.05	55.50	42.42	2.02	–	0.28
13	0.53	4.55	30.72	37.05	27.15	0.13
14	0.69	3.90	15.55	29.76	50.10	0.11
15	0.03	34.68	68.44	1.37	–	0.24
16	0.31	2.25	27.00	30.72	39.72	0.12
Site 2						
8	0.26	5.56	51.38	42.80	–	0.16
9	0.17	1.69	19.22	30.65	48.27	0.11
10	–	20.88	77.80	1.31	–	0.22
11	–	28.19	71.57	0.24	–	0.23

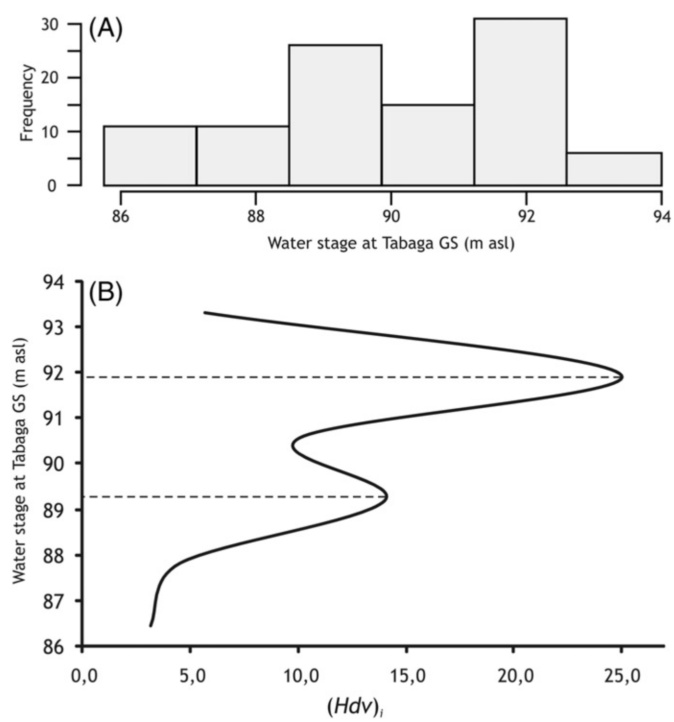

Fig. 11. Year 2008 data set: water level frequency distribution (A); effective water level calculation from plotted efficiency index $(Hvd)_i$ (B).

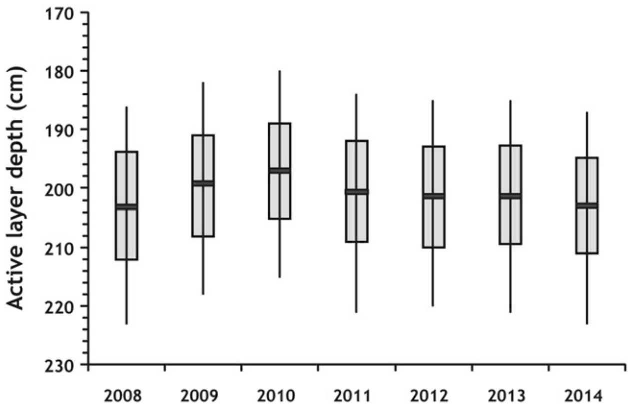

Fig. 12. Active layer depth at CALM R42 plot (Tuymada), 2008–2014; standard deviation range, grey; extreme values shown by whiskers.

the adjacent slope, active layer depth, and if possible, soil temperatures at various depths, thawing front position, vegetation, state of slumped material, if present (frozen or thawed), and water temperature at several points close to the notch.

Satellite imagery sources are abundant, and remote sensing data are becoming more accurate and accessible than ever before. Regardless of this, several issues are to be solved prior to its proper use in fluvial thermal erosion studies. First, manual bank delineation is time-consuming and person-dependent and requires extensive field experience in the region under study. Automated methods, in its turn, still require adequate training samples and are subject to several types of errors, i.e. caused by shadows from banks, trees, and clouds (Merwade, 2007).

The ultimate objective of fluvial thermal erosion studies is a valid numerical model of this process. A conceptual framework is to be developed based on existing bank stability models for a non-permafrost environment, supplementing them with computational modules of ablation, thawing front propagation, and hydraulic data assimilation. The other possibility to catch is to use a 'reverse engineering' approach. First, a hydraulic model is applied to the river section in question in order to assess (imitate) basic flow parameters: water level, flow velocity and direction, water surface gradient. Overlapping these data with field survey results and sediment properties should allow us to relate bank erosion rates to hydraulic and to sedimentary controls.

Forecasts of fluvial thermal erosion are important for the communities occupying the river banks, though they remain a distant goal. Hydrological models are useful tools in assessing future streamflow distribution under a changing climate, and modelled hydrographs can be subsequently employed in hydraulic models to evaluate flow parameters. Given the sedimentary structure and modelled hydraulic parameters, future fluvial thermal erosion rates can be evaluated.

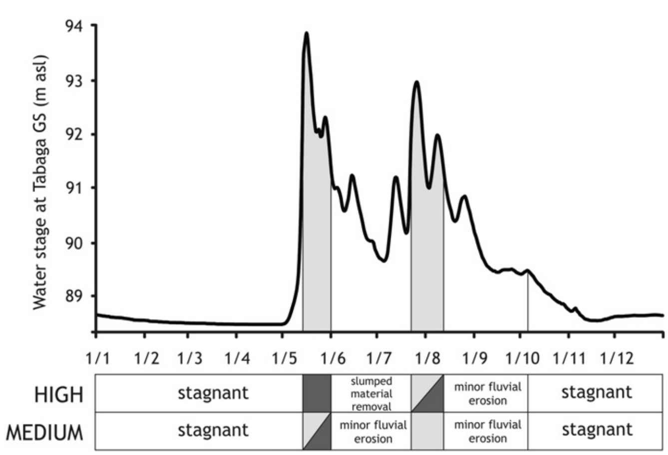

Fig. 13. Geomorphic processes acting in banks of different floodplain levels and their hydrological controls; water stage equal or above the effective water level, shaded portions; bars below the graph: fluvial thermal erosion, dark grey; fluvial erosion, light grey.

6. Conclusions

The Lena River floodplain is a complex sedimentary body, built up in several levels, varying in height. Fluvial thermal bank erosion is assumed to be a dominant force in shaping the Lena River banks. Our results show that this process can be selective when different floodplain levels are affected.

Fluvial activity occurs in all ranges of water levels. In a long-term perspective some water levels are assumed to be more efficient in performing the geomorphic action, i.e., more frequent and implying higher stream energy. Effective water level interval was inferred from magnitude-frequency analysis, marking hydrological conditions under which the fluvial thermal erosion is active in a long-term perspective. For the studied middle Lena River section, it lies between 91.05 and 92.62 m asl, and corresponds, roughly, to water discharges between 21,000 and 31,000 $m^3 s^{-1}$.

Effective water level represents the hydrological driver of fluvial erosion, but the erosional response of the river banks depends nonetheless on the mechanical features of the surface being eroded. Overlapping this effective water level with studied bank profiles allows drawing some conclusions on the differences between the sites and subsequently between various floodplain levels and banks in terms of processes involved in bank retreat.

In high floodplain banks, effective water level corresponds to the noncohesive layer of sandy channel-fill deposits, remaining frozen during the spring peak discharges. Repeated thermoerosional niche failures and subsequent removal of collapsed material represent the dominant bank erosion process within this floodplain level. Medium floodplain level is significantly lower and is subject to mechanical washout, largely unaffected by thermal interactions.

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