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Landslides in clayey soils of western Estonia

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LIST OF ORIGINAL PUBLICATIONS

This thesis is based on the following papers, referred to in the text by their Roman numerals. The papers are reprinted by permission of the publishers.

I   **Kohv, M.,** Hang, T., 2011. Pore pressure fluctuations within quasi-stable slopes in south-western Estonia and their influence on slope stability. Manuscript

II   **Kohv, M.,** Hang, T., Talviste, P., Kalm, V., 2010. Analysis of a retrogressive landslide in glaciolacustrine varved clay. Engineering Geology, 116 (1–2), 109–116.

III Kohv, M., Talviste, P., Hang, T., Kalm, V., 2010. Retrogressive slope failure in glaciolacustrine clay: Sauga landslide, western Estonia. Geomorphology, 124 (3–4), 229–237.

IV Kohv, M., Talviste, P., Hang, T., Kalm, V., Rosentau, A., 2009. Slope stability and landslides in proglacial varved clays of western Estonia. Geomorphology, 106, 315–323.

Author's contribution

Paper I: The author is primarily responsible for research planning, field works, data gathering, statistical analysis, interpretation of the acquired data and slope stability modelling, writing of the manuscript.

Paper II: The author is responsible for field works, sampling, varve correlation and interpretation, slope stability analysis and writing the manuscript.

Paper III: The author is responsible for research planning, field works and sampling, slope stability modelling, installation of the monitoring network and monitoring measurements twice a year, writing the manuscript.

Paper IV: The author is responsible for data collecting, field works, development of the digital elevation model, writing the manuscript.

1. INTRODUCTION

Landslides are one of the most costly natural hazards that mankind has to face, endangering infrastructure, properties and lives worldwide (Brabb, 1991). Statistics from The Centre for Research on the Epidemiology of Disasters (CRED, http://www.cred.be/) show that landslides contribute to about 17% of the fatalities due to natural hazards (Kjekstad and Highland, 2009). Asia had the largest number of catastrophic landslides during the 20th century, while the landslides in Europe are the most expensive ones, each costing up to 23 million $ on average. It has been estimated that in many developing countries economic losses caused by various types of landslides may reach 1%–2% of the gross national product (van Asch et al., 2007). Slope processes may also have serious impact to forest, streams, habitat diversity etc. (Hewitt, 1998; Geertsema et al., 2009). Landslide numbers and cost seem to increase due to increasing urbanization and development, deforestation and increasing regional precipitation related to climate change (Schuster, 1996). Analysing, evaluating and mitigating the hazard and risk associated with mass movements are therefore a challenge to many scientists, engineers, developers and decision-makers. Estonia has a relatively flat topography and is tectonically stable, so landslides are generally rare here (Miidel and Raukas, 2005). However, in recent years, in parallel with increasing storminess (Jaagus et al., 2008) and frequent storm floods (Suursaar et al., 2006) at the coasts and increasing inhabitation of riverside areas, a number of sliding events have drawn attention on the slope stability and triggering mechanism of landslides in proglacial clayey soils of western Estonia.

1.1. Landslides in Estonia

The earliest landslide record from the Estonian territory dates to 1772 when the newly built mansion owned by the Von Uexküll family slid to the Vigala River in western Estonia (Aitsam, 2006). The first geotechnical investigations of several landslides that occurred in Pärnu town were carried out in the 1960s–1970s (Olli and Martin, 1961; Kaljund, 1967; Liblik and Smirnova, 1984; Vilo 1986). These landslides were rotational slumps related to varved glaciolacustrine clays widely distributed in western and north-western Estonia (Pirrus, 1968). Rockfalls, topples and transitional slides are known at the Ordovician and Silurian limestone cliffs (klints) in northern Estonia and on the Island of Saaremaa, West Estonian Archipelago (Vilo, 1986; Miidel and Raukas, 2005; Suuroja, 2009). Relatively weak Cambrian blue clay that outcrops at the base of the Ordovician klint in northern Estonia is also prone to landslides and caused major slope stability issues during the remediation of the Sillamäe radioactive tailings bond (Miidel and Raukas, 2005; Torn, 2008). The most detailed overview of the pre-1980 landslides in Estonia can be found in an unpublished report by Liblik and Smirnova (1984), which lists several river valleys (Pärnu,

Sauga, Vääna, Kasari and Keila Rivers) with potentially unstable slopes and shows their location on the map. This report also mentions landslides related to Cambrian clay in northern Estonia but gives no detailed descriptions. Vilo (1986) defines the Pärnu geotechnical region as most susceptible to landslides in Estonia. Figure 1 presents the areas in Estonia with landslide records known from published sources (Olli and Martin, 1961; Kaljund, 1967; Liblik and Smirnova, 1984; Vilo, 1986; Miidel and Raukas, 2005; Suuroja, 2009) and those mapped within the current study.
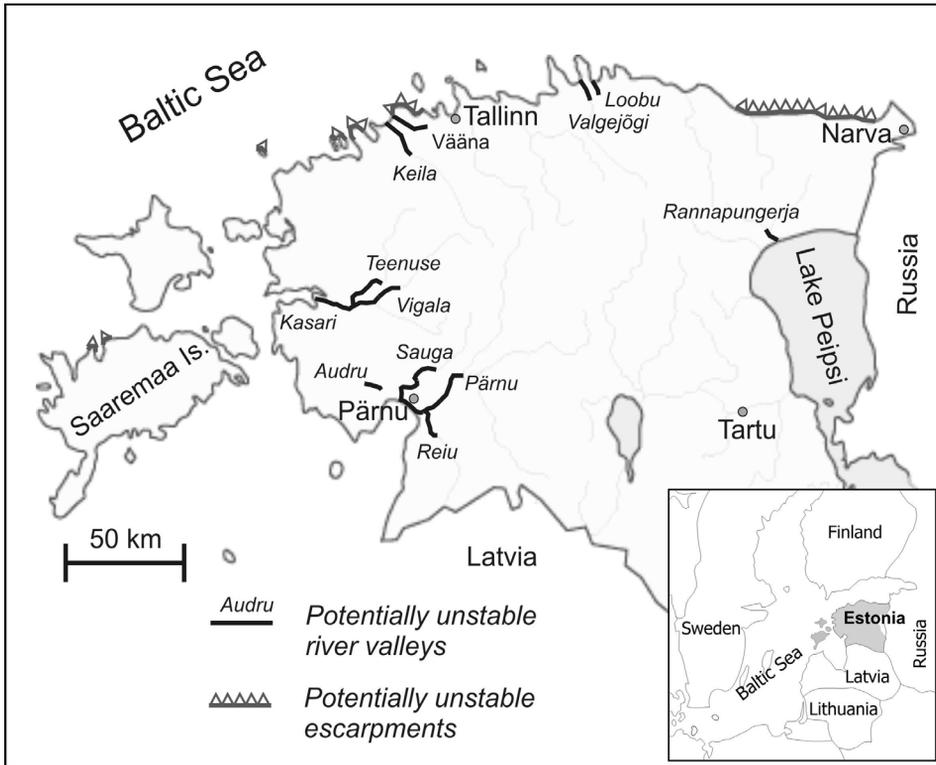
Figure 1. Areas with landslide records in Estonia.

The areas susceptible to landslides have been determined on the experts knowledge from written sources or personal experience, without targeted mapping or cartographic analyses. After a relatively quiet period in the1980s–1990s new reports of landslides appeared in 2000 when a number of slope failures took place near Pärnu town. Institutional interest also grew, both in the municipality and state level. The municipalities surrounding Pärnu town faced rapid real estate development with planned building activity close to fresh and old landslides sites. This problem aroused interest in landslides and finally led to the work presented in this thesis. The state level interest is bounded by EU

legislation as Estonia joined the European Union in 2004. According to the EU's Thematic Strategy for Soil Protection (COM232/2006), landslides are one of the main eight threats to European soils. Therefore the European Union has also supported several landslide associated research projects like GALAHAD, MUSCL, RUNOUT, ENVASSO, RAMSOIL and SAFELAND. Only the most recent and still ongoing project, SAFELAND covers Estonian territory other initiatives were already terminated when Estonia joined the EU in 2004. Besides other objectives, SAFELAND is aiming to harmonize and improve landslide inventories, susceptibility, hazard and risk mapping within the EU. Further studies and more advanced mapping efforts in Estonia will probably grow out of this activity. The current thesis is a contribution to this initiative and has the following objectives:

- to conduct an inventory of landslides in the river valleys cut into glaciolacustrine clay soils of the low-laying coastal plain in the Pärnu area, on the eastern Baltic Sea coast;
- to map in detail the morphological and geological setting of some recent landslides;
- to model the slope failures and determine the preparatory processes and triggering mechanisms of slide events;
- to determine the critical slope gradient in the given geological–geomorphological conditions and provide a landslide susceptibility map for the region studied;
- to investigate pore pressure fluctuations within quasi-stable slopes and determine their influence on slope stability;
- to monitor the development of the post-sliding slope.

The thesis comprises data from four scientific papers. Two papers (Papers II and III) present the results of case studies describing in detail two different landslides. Paper I focuses on groundwater conditions within the slopes and its role as a slope failure trigger. Paper IV gives an overview of the slides near Pärnu town and the results of GIS-based landslide susceptibility mapping.

2. MATERIAL AND METHODS

2.1. Geological setting of the study area

The studied river valleys lay within an area of ca 200 km^2 around Pärnu Bay, on the eastern Baltic Sea coast, Estonia, between N 58º20' to 58º27' and E 24º15' to 24º40' (Figure 2).
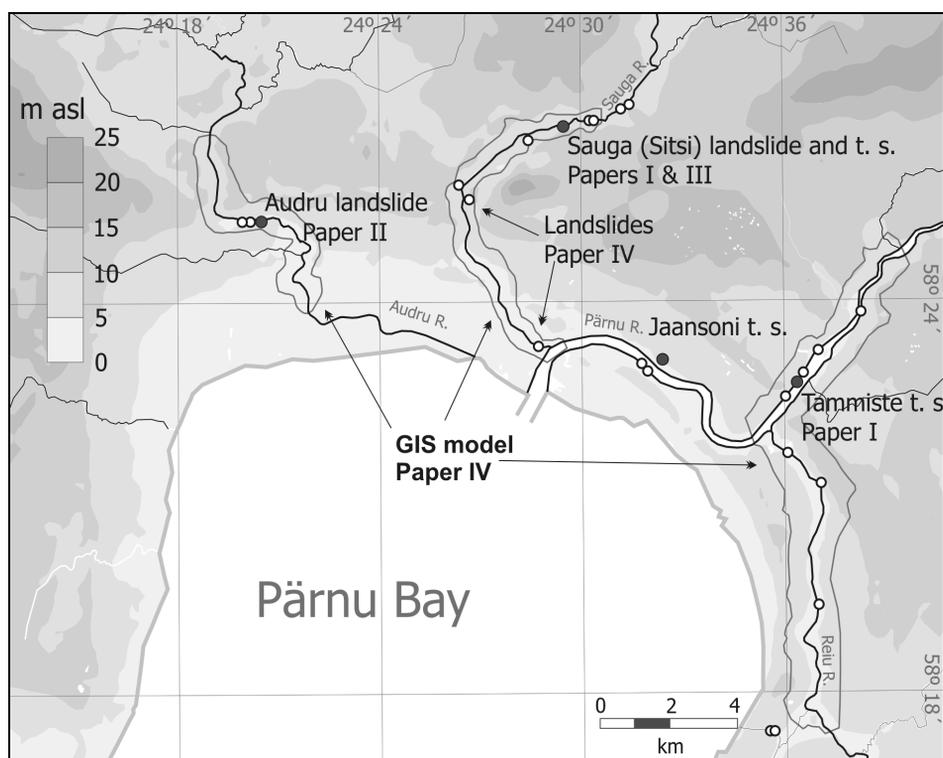

Figure 2. Location of the study area and investigated sites (black dots). White circles mark monitoring stations operated by the Estonian Meteorological and Hydrological Institute. The abbreviation t. s. means test site.

The study area is a coastal plain with altitudes varying between 2 and 15 m, diversified by dunes, raised bogs and river valleys. River valleys, eroded into the lacustrine-marine plain, are up to 15 m deep. The steepest slopes (25–30º) with ongoing erosion occur at the convex bank of the river meanders. The valleys are relatively young in geological sense as their development due to fluvial erosion started only after the Litorina Sea transgression (ca 7400–7200 yrs BP) and progressed in accordance with the lowering of base-level and regression of the Baltic coastline (Veski, 1998). A generalized geological section of the study area is shown in Figure 3.
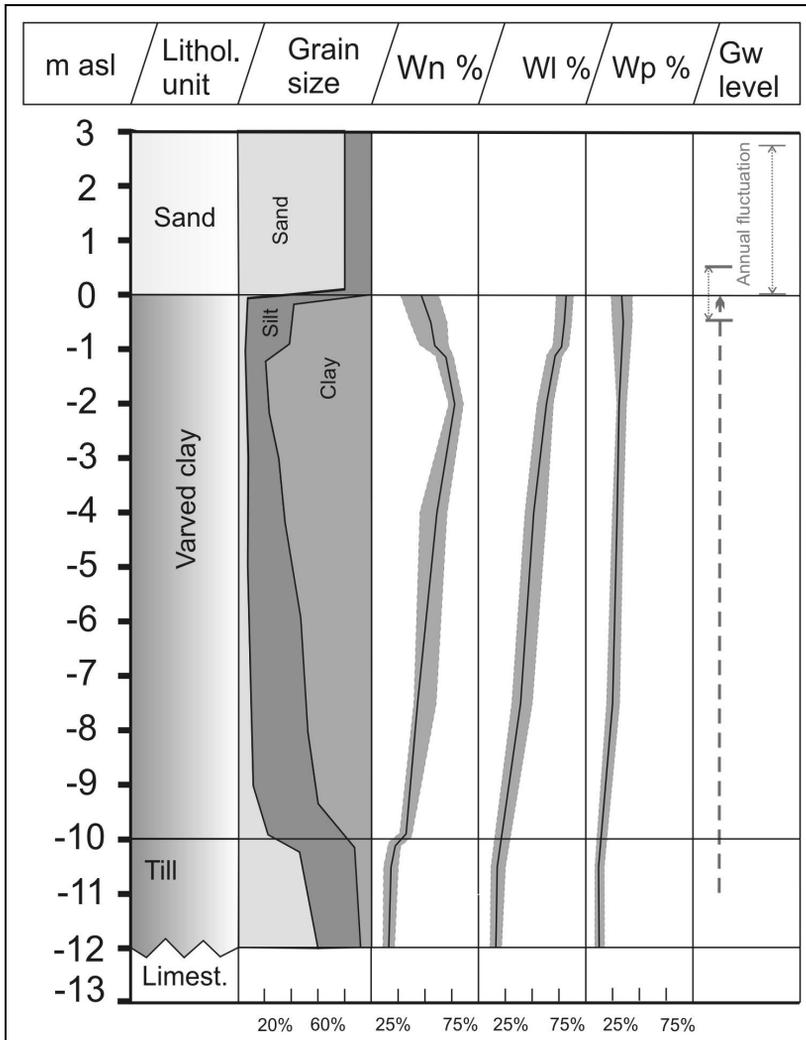
Figure 3. Generalized geological section of the study area. Gw level – groundwater level (two aquifers), Wn – water content, Wl – liquid limit, Wp – plastic limit (Kattel, 1989; Hang et al., 2007).

The uppermost part of the sedimentary cover in the Pärnu area, overlying the Proterozoic crystalline basement, includes Silurian limestone and Devonian sandstone. The surface of those sedimentary rocks lies at an altitude of −10 to −15 m (Tavast and Raukas, 1982). Paleozoic rocks are covered by Quaternary sediments; the lowermost section is usually loamy till of Late Weichselian age (Raukas, 1978). The latter is overlain in average by 10 m (max 30 m) thick laminated glaciolacustrine silty clay. Varved clay is usually covered with 2–3 m (max 19 m) of Holocene marine sand and silt; the sand cover may be locally absent (Figure 3).

The weakest soil type in the Pärnu area is glaciolacustrine clay which has been deposited in a vast proglacial sedimentary basin (Pirrus, 1968; Hang et al., 2007) during the last deglaciation. Laterally geotechnical qualities of clay do not change notably within this former sedimentary basin (Vilo, 1986; Saarse, 1992). The clay fraction (<0.002 mm) is dominant in varved clays, particularly in the upper 2 m thick portion, where it constitutes 69–79% (Hang et al., 2007). The silt fraction increases and clay decreases (to 46–51%) towards the depth. In the clay fraction illite (70–80%), smectite-illite (9–19%), kaolinite (6–8%) and chlorite (2–5%) dominate, while silty layers are composed primarily of quartz (46–59%), feldspars (14–18%) and carbonate (11–21%) (Kattel, 1989). According to the Casagrande plasticity chart (Coduto, 1998) the upper portion of clays are classified as CL (fat clay) and the lower half as CH (lean clay). The water content of clay decreases with depth, being 70–90% in the upper portion and lowering downwards to 30–60% (Kattel, 1989). However, the uppermost part of varved clays may be dehydrated and mixed by freeze-thaw cycles, forming an up to 2 m thick desiccated layer with an approximately twice lower water content compared to underlying varved clay. This layer is generally absent below 5 m altitude (Hang et al., 2007)

Two groundwater aquifers occur in the Quaternary sediments in the Pärnu area. The upper, unconfined aquifer is bound to marine sands, and the lower, confined one, to till underlying glaciolacustrine clay. Glaciolacustrine clay acts as an aquiclude between the groundwater aquifers. The upper aquifer is controlled by precipitation, transpiration, evaporation and by the water level of adjoining rivers. The aquifer in till below clay is pressurized and the piezometric level fluctuates between 0 and 1.5 m asl in Pärnu town (Kohv, 2005), which is 11 m above the base of the clay layer. In period 1990–2000 the pressurized ground-water level rose up to 12 m in Pärnu town (Talviste, 2004).The recent landslides and traces of historical ones were mapped along the major river valleys in the study area during the field surveys in 2001. Complementary analyses were done in 2008 after the Estonian Land Board opened access to high-resolution (0.25 m pixel size) orthophotos in their website (http://www.maaamet.ee/). The morphology of nine recent landslides was instrumentally mapped with the Dhalta 010B tachymeter (angle measurement error ± 10") and in the case of Sauga landslide (Paper II), with the Sokkia 5A total station (accuracy 15").

The geological setting of the studied landslides was described through drill cores, obtained with either machine or manually operated Russian-type peat corer. The Russian-type peat corer is a side filling chamber-type sampler operated by metal rods. It has been designed by Belarus geologists and was first presented to the international audience by Jowsey (1966). This corer gives relatively undisturbed and uncompressed samples that are needed for varve correlation and indication of slip surfaces. The 1 m long clay sections were wrapped in plastic film to avoid drying and placed in half-cut PVC tubes for transport. Subsamples for water content measurements were taken at the coring site and transported within hermetical containers. Undisturbed soil samples for the triaxial shear test were collected with a thin wall sampler (110 mm inner

diameter) that was pushed into varved clays. All analyses and geotechnical tests except some water content measurements and varve correlation were conducted by the accredited Geotechnical Laboratory of the Estonian Environmental Research Centre according to the following standards: water content (CEN ISO/TS 17892–1), Atterberg limits (CEN ISO/TS 17892–12), grain-size distribution (CEN ISO/TS 17892–4), triaxial consolidated undrained shear tests (CEN ISO/TS 17892–9) with 15 strain points. Vane shear tests were made by the geotechnical company REI with a 10 x 20 cm vane, following the test procedure described in the Estonian standard EVS 1997–3:2003. Locations of failure zones were identified via varve correlation. An overview of the analyses performed during the current study is presented in Table 1.

Table 1. Analytical material used in the current study.

	PAPER I	PAPER II	PAPER III	PAPER IV	In total
Water content	–	42	56	–	**98**
Grain size	–	2	7	–	**9**
Triaxial CU tests	–	–	5	–	**5**
Atterberg limits	–	2	12	–	**14**
Vane shear tests	41	–	9	–	**50**
Boreholes	3	8	8	–	**19**
Piezometers	9	–	–	–	**9**
Investigated landslides	–	1	1	9	**11**

Pore pressures were measured at two test sites (Figure 2) with automatic piezometers (pressure hysteresis <0.1% and linearity error <0.15%) manu-factured by BAT Geosystems AB, Sweden. The piezometers were installed in vertical cluster configuration at two test sites: six piezometers at the Sauga and four at the Tammiste test site (Figure 2). Standard metal water pipes and couples with the 25 mm inner diameter were used to install porous polyethylene plastic filters (filter height 35 or 20 mm) to the desired depth. The piezometers were pushed to the ground and the pore pressure and temperature were measured with an 1 hour interval. The pore pressure data analysed in Paper I were collected between 01 May 2009 and 01 May 2011. The meteorological and hydrological data used in Paper I were obtained from the Sauga meteorological station (3 km SW from the Sauga test site) and the hydrological posts operated by the Estonian Meteorological and Hydrological Institute.

Slope stability calculations were done with the software package Slide v 5.0, which couples the finite element groundwater model with slope stability analyses. The latter is realized with limited equilibrium methods that are most widely used for slope stability calculations (Coduto, 1998; Aryal, 2006). Corrected calculation schemes of Janbu and Spencer (Janbu, 1954; Spencer,

1967) were used for this study. These belong to a group of rigorous methods that satisfy both moment and force equilibrium. Variance of the calculated factor of safety is relatively small when whatever rigorous method is used (Sjöberg, 1996; Zhu et al., 2003). Detailed calculation schemes and differences between various limited equilibrium methods are presented by Fredlund and Krahn (1977). Initial morphology of the analysed slopes was obtained from topographic maps (Paper III), by measurements from the adjoining slopes (Paper II) or by accounts provided by local people. Groundwater levels used in the models were interpolated from the nearby measurements (Paper II) or measured after the investigated slope failures (Paper III). Measured pore pressures were used in the slope stability models in Paper I. Hydraulic conductivities of different soil types were taken from the literature (Kaljund and Mets, 1976; Kalm et al., 2006). Soil shear strength, necessary for the slope stability models, were obtained from vane shear tests, triaxial tests or calculated with the empirical model outlined by Talviste (2002). Strength parameters for marine sand were taken from Kaljund and Mets (1976). Till was considered as a bedrock for the model, i.e. it has infinite strength.

A potential failure surface with the minimum factor of safety was located with the grid search option. The slip centres are placed above the analysed slope model, forming a grid of nodes. In this study we used a grid of 20 x 20 m. Each slip centre has 10 evenly spaced circular potential failure surfaces. The software calculates a total of 4851 safety factors for each analysed slope configuration, finding the weakest circular surface within the search limits set by the user.

In order to analyse long-term changes in the pressurized groundwater level, monitoring data from 51 groundwater monitoring wells (opening Quaternary, Devonian and/or Silurian aquifers) operated by the Estonian Geological Survey or the company of IPT Projektijuhtimine OÜ were used. The datasets covering the period from 1960 to 2009 were cleaned of measurement errors, computed to yearly averages and organized as a GIS-database. Interpolated groundwater surfaces with the grid size of 300 x 300 m were generated for selected years (1966, 1976, 1986, 1991, 1996, 2001) with the kriging interpolator (Paper I).

The methodology suggested by Brabb (1991) was followed to generate the landslide susceptibility map for the Pärnu area. The digital elevation model (DEM) was created from elevation data, digitized from topographic maps at a scale of 1:10 000 (isobases with 0.5 m intervals and irregularly distributed elevation marks). The DEM was generated by using the linear solution of Natural Neighbour interpolation with the grid size of 4 x 4 m. Slope angles were derived from the DEM using the terrain slope operator. The classes of slope angles were compared with critical values for the underlying geological setting and thus potentially unstable cells were identified. The GIS package MapInfo 8.5 (with add-on Vertical Mapper) was used for terrain analysis and landslide susceptibility mapping.

The Sauga landslide (Paper III) was monitored twice a year with measurements of the installed fixed points (1.5 m long steel rods installed into the ground) and the bank of the river channel. The corners of the nearby living

house were used as stable reference points. Comparative photos were taken from same vantage points in order to document the slope development. Results from the period 2005–2010 are presented in the current study but the monitoring programme is still going on.

Statistical analysis presented in Paper I was done in the following way: datasets were checked against normality with the Shapiro-Wilk test; Spearman's rank order correlation and cluster analysis with Ward's linkage method were used to find and exemplify relationships between the analysed meteorological, hydrological and pore pressure time series. A good overview of the statistical methods used is presented by Davis (2002). Significance level p=0.05 was used in all analyzes.

3. RESULTS AND DISCUSSION

3.1. Landslide inventory

The GIS database of slope failures near Pärnu town includes currently 26 records. The earliest record goes back to 1905 and the largest known landslide occurred in 1976 (Table 2, Figure 4).
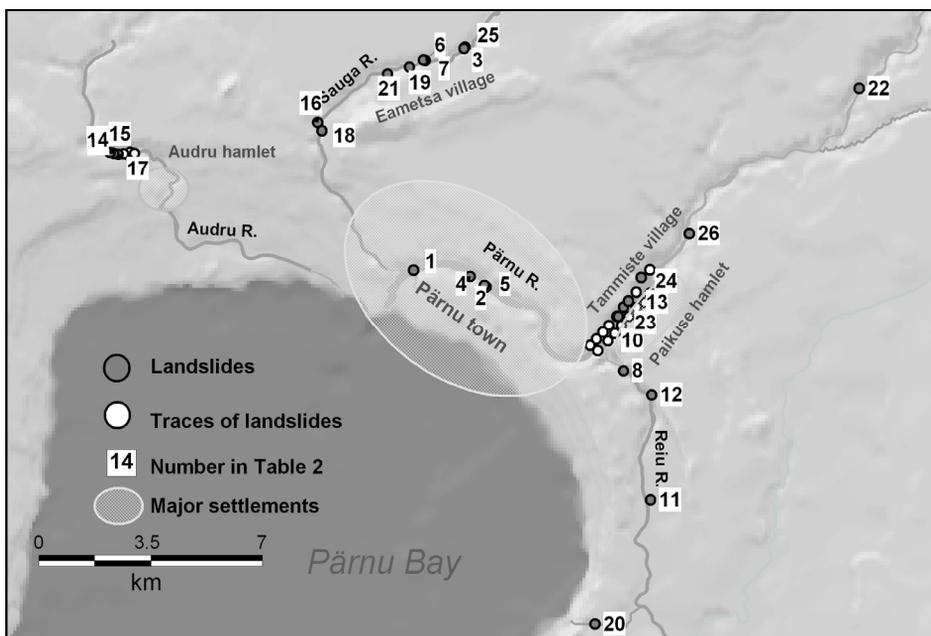

Figure 4. Location of the landslides known in the Pärnu area. The numbers indicate the record number in Table 2.

Landslides clearly converge in three areas: Tammiste village at the Pärnu River, Eametsa village at the Sauga River and a smaller group about 1 km upstream from Audru hamlet at the Audru River. Other slope failures are more sparsely distributed along the river valleys.

Table 2. Location, morphology and classification of slope failures in the Pärnu area.

Nr	Time	Name	Location	Width (m)	Main scarp (m)	Length (m)	Group
1	1905	Nahksilla	58°23 .3' 24°28.7'	–	–	–	–
2	1927	Viisnurk 1	58°23.6' 24°31.8'	–	–	–	B
3	>1960	Nurme mill	58°26.9' 24°31.1'	90	–	45	A
4	1960	Fishindustry	58°23.1' 24°31.3'	80	–	25	B
5	20.11.1966	Viisnurk 2	58°23.6' 24°31.8'	180	4	45	B
6	9.12.1974	Nurme bridge	58°26.7' 24°29.8'	160	–	–	A
7	1976	Nurme bridge 2	58°26.x' 24°29.x'	200	–	100	A
8	**2000**	**Reiu 1**	**58°21.6' 24°36.2'**	**32**	**1.2**	**22**	**B**
9	Spring 2001	Uulu channel 1	58°17.x' 24°35.x'	–	–	–	B
10	2002	Pärnu 2	58° 22.6' 24° 36.2'	150	2.5	30	B
11	**2002**	**Reiu 3**	**58°19.4' 24°36,9'**	**7**	**2.5**	**15**	**B**
12	**18.02.2002**	**Reiu 2**	**58°21.2' 24°37.1'**	**23**	**1.5**	**16**	**B**
13	**14.03.2002**	**Pärnu 1**	**58°22.7' 24°36.3'**	**80**	**5.4**	**42**	**B**
14	**Spring 2002**	**Audru 2**	**58°25.3' 24°19. 8'**	**8.4**	**0.3**	**3.5**	**A**
15	**Spring 2002**	**Audru 3**	**58°25.3' 24°19.9'**	**16.2**	**0.4**	**4.3**	**A**
16	**Spring 2002**	**Sauga 2**	**58°25.7' 24°26.4'**	**13**	**1.4**	**13**	**A**
17	**12.05.2002**	**Audru 1**	**58°25.2' 24°20.1'**	**75**	**1.2**	**36**	**A**
18	11.01.2005	Eametsa	58°25.6' 24°26.6'	80	1	60	A
19	**19.12.2005**	**Sauga**	**58°26.3' 24°29.4'**	**137**	**4**	**80**	**A**
20	May 2006	Uulu channel 2	58°17.4' 24°35.1'	43	11	18	B

Nr	Time	Name	Location	Width (m)	Main scarp (m)	Length (m)	Group
21	22.01.2007	Sauga 3	58°26.0' 24°27.2'	10	1	5	A
22	Spring 2007	Urumarja	58°26.2' 24°43.8'	30	2	20	–
23	12.05.2007	Tammiste	58°22.5' 24°36.0'	38	5	38	B
24	–	Tammiste Oja str.	58°23.1' 24°36.8'	50	3	40	B
25	–	Sauga 4	58°27.0' 24°31.2'	60	2	40	A
26	17.04.2009	Sindi cemetery	58°23.8'' 24°38.3'	20	4	15	A

Bold landslide records in Table 2 were analysed in Paper IV. Those were relatively fresh slides with distinctive borders and open scarps during the landslide mapping by the authors of Paper IV in 2002. The length and width of landslides in Table 2 are presented according to the nomenclature proposed by Cruden and Varnes (1996). As earlier records come from literature of variable reliability, some of the parameters are unknown and therefore marked with a dash in Table 2.

Landslide frequency has risen in recent years (Figure 5). A large portion of the known slope failures, 8 out of 26 (31%), took place in 2002.
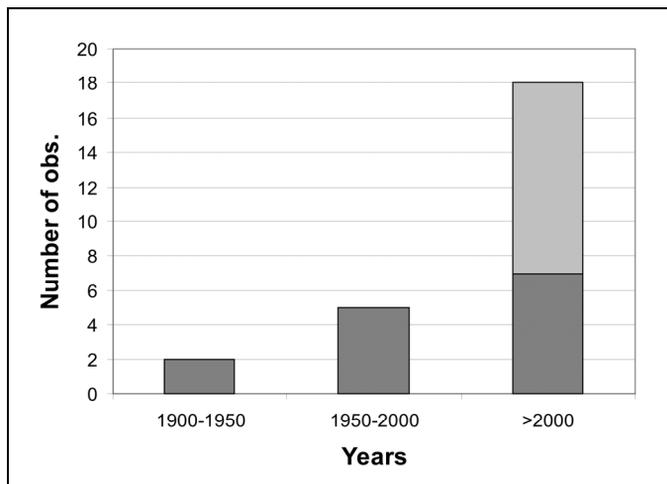

Figure 5. Landslide frequency in the Pärnu region. Light grey marks the landslides detected during the 2002 survey, dark grey records from geotechnical reports, newspapers, etc.

Increasing slope instability has been reported from many regions and explained by changing land use and deforestation (Guthrie, 2002; Glade, 2003) and by more frequent extreme weather with increased precipitation (Buma and Dehn, 1998; Dehn and Buma, 1999; van Beek, 2002, Zogning et al., 2007). To the authors knowledge land use changes in the Pärnu area have not been investigated but preliminary analysis of the historical maps and orthophotos available on the Estonian Land Board website (http://www.maaamet.ee), covering the period 1900–2010 shows mild enlargement of Pärnu town until the 1980s, followed by faster development. The surrounding farmlands were relatively unchanged until 2000, then rapid expansion of the Pärnu suburbs begun. This recent trend caused partial deforestation and construction activities within and in the neighbourhood of the river valleys.

Morphological, geological and hydrogeological properties were used to classify recent landslides. Group A consists of landslides with multiple, retrogressive failures. They have elliptical failure surfaces or failure zones that penetrate the whole varved clay section and usually occur after the spring thaw. Group B includes landslides that are single slope failures occurring on the slopes capped with substantial sand cover (>3 m). They are usually triggered by intense rain infiltration or thaw. In addition to those two groups group C was distinguished in Paper IV. Those small-scale slides occur on the river banks and are caused by river undercutting. It was concluded in Paper IV that those small slides may be initial stages for the group A slides. Therefore they are not brought forth as a separate group but within the group A landslides.

Following the classification nomenclature by Cruden and Varnes (1996), group A slides can be described as retrogressive rotational earth slides and those of group B as rotational earth slides. The proposed classification differs from the earlier classification by Kaljund and Vilo (1967). The last authors described landslides in the Pärnu area as landslides with an elliptical surface of rupture, caused by human-induced stress at the upper part of the slope. However, the second type of landslides in their paper, characteristic of the Vääna and Keila River valleys (Figure 1), is almost identical to group B landslides in the current study. Vääna and Keila slides were described as rotational landslides with a nearly perfect half-circular surface of rupture and caused by undercutting and/or lowering of the river water level (Kaljund and Vilo, 1967). We will analyse the group B landslide mechanism later, but a preliminary conclusion is that those slides are very similar to those described by Kaljund and Vilo (1967). Landslide inventory has right now 26 records. Recently published LIDAR based digital terrain model from Estonian Land Board, covering almost whole Estonian territory, may reveal currently unknown sites. This approach has already proven itself very promising in many cases around the world (Schulz, 2006; van den Eeckhaut et al., 2007, Razak et al., 2011)

3.2. Morphological and geological properties of the studied landslides

Group A slope failures occur primarily in clayey slopes with slope angles between 10° and 20°. Group B landslides occur in sandy slopes with slope angles around $20-30^0$ (Paper IV). Travel angles for group A landslides are around 5–6° and for group B around 15–17°. The travel angle is a measure of the vertical difference of the lowest and highest points on a landslide as a ratio of the length to the travel path (Geertsema et al., 2009). Geertsema and Cruden (2008) show that landslides in glaciolacustrine sediments in Canada have usually travel angles between 6.5 and 14°. However, those sediments are strongly overconsolidated by overriding glaciers, whereas Estonian glaciolacustrine sediments are normally consolidated.

Other morphological parameters used in this chapter are given in Table 2. Figure 6 is a landslide width versus length scatterplot with fitted linear regression trendlines. Group A landslides (11 cases) have a linear trendline with a very high r^2 value (0.97) while the group B (10 cases) trendline has $r^2 = 0.29$; also the significance (p) of the group B trendline is below the standard 0.05 value.
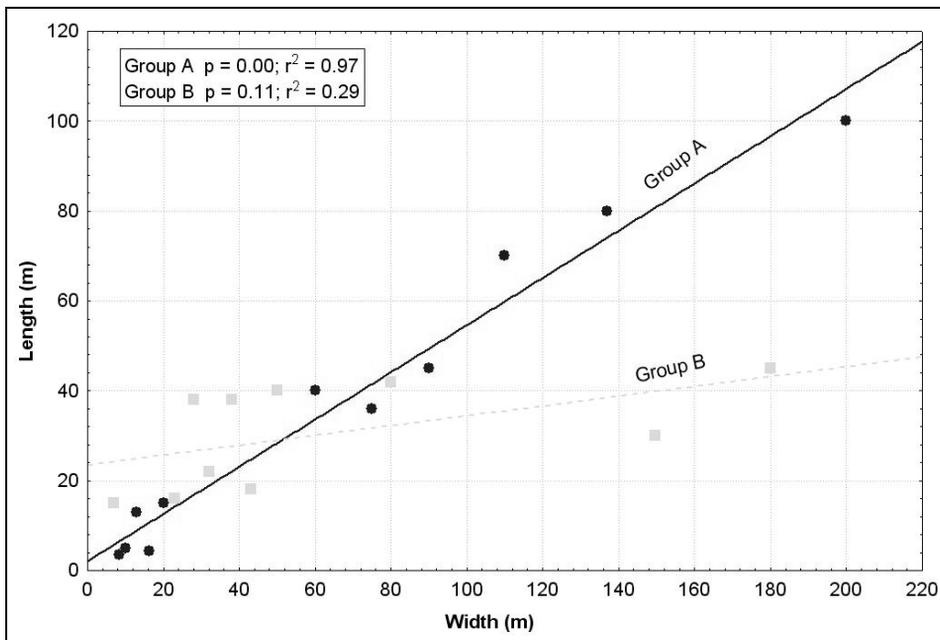

Figure 6. Landslide width vs length. Analysis is applied to 21 landslides with known morphological parameters presented in Table 2.

The difference between two groups is mainly in the slope component of the trendline. Increasing width causes longer landslides in case of group A land-

slides compared to group B. Group B landslides occur in steeper, shorter slopes and usually do not affect significantly the areas outside the valleys, causing a relatively low rise of the trendline in Figure 6. Group A landslides occur in longer clayey slopes and retrogress outside the valleys, thus the Group A trendline in Figure 6 is steeper than in case of group B.

The explanation given above also manifests itself in size comparison: on average group A landslides (ca 4500 m^2) are roughly three times larger than those of group B (ca 1500 m^2). Generally, group B landslides are smaller and have higher travelling angles than group A landslides. The tendency of smaller landslides to have higher travelling angles is pointed out by Corminas (1996) and Hungr et al. (2005), while Geertsema and Cruden (2008) do not show this relationship clearly but point to the importance of the failed soil type. In our case both factors seem to b some significance but the number of observations is too small to allow further analysis.

Retrogressive landslides of group A (Table 2) are multiple slope failures that start at the river bank. The initial failure will make the upper part of the slope unstable, causing a retrogressive failure pattern (Papers II and III). Successive failures also widen parallel to the river channel. This mechanism produces distinctive "stepping" morphology of the sliding body (Figure 7) which is described in Papers II and III.

Figure 7. Surface morphology of the retrogressive Sauga landslide (Group A).

The area affected by sequential failures increases roughly two- to threefold with each failure. The area gain is quicker (threefold) at the initial stages and lowers to twofold at the final ones. This generalization is based on the morphological maps of the Viisnurk-2, Audru-1 and Sauga landslides (Table 2).

Varve correlation between the sediment cores shows that group A slope failures occur along the ruptured zone that starts as a plane in the upper part of the landslide and may evolve to a 1 m thick ruptured zone (Papers II and III).

Group A landslides are characteristic of the areas west and north of the Pärnu River valley. All landslides recorded in the Sauga and Audru River valleys belong to this group (Figure 4, Table 2). The biggest cluster of group A landslides occurs in Eametsa village 5 km north of Pärnu town (Figure 4, Table 2).
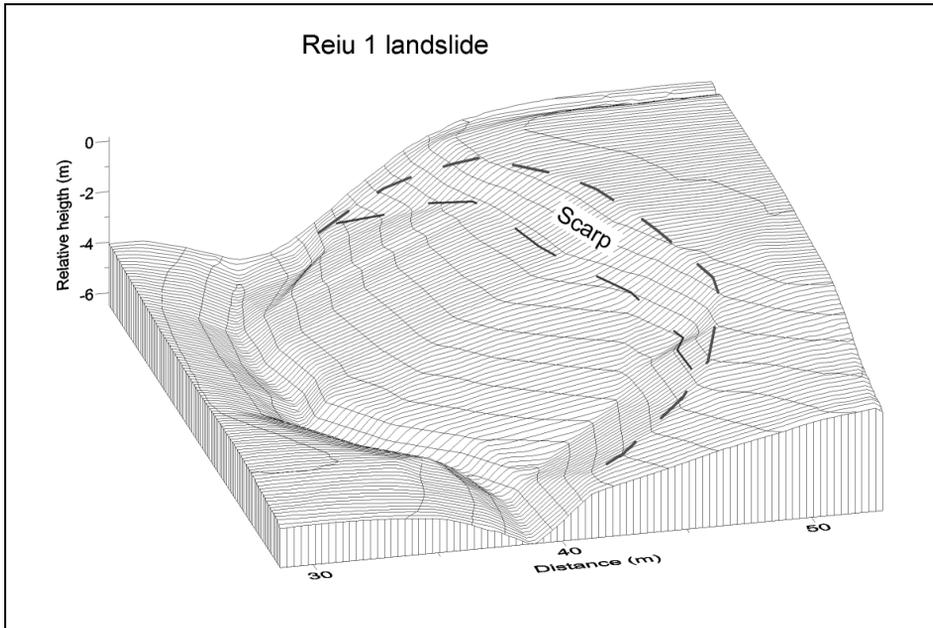

Figure 8. The Reiu 1 landslide is a single rotational failure in sandy slope and belongs to group B. Isolines have 0.5 m spacing. For location see Figure 4.

Group B slope failures are generally smaller than group A landslides and occur primarily in sandy slopes. Although generally smaller, the highest landslide scarp (11 m in the Uulu channel 2 landslide) and also the second and third highest ones belong to group B landslides. Higher scarps are caused by the position of these landslides (upper part of the slope) and by the fact that their rupture surfaces do not extend deeper than the bottom of the valley. This positioning leaves the deeper part of the valley open for the landslide body to move in. Also, at steeper slopes a larger component of the gravitational force is acting parallel to the slope surface and causes acceleration of the mass movement after the slope failure. Those two factors contribute to a longer vertical travelling distance of the sliding body and the resulting scarp is higher compared to group A landslides. Group B failures occur usually after heavy

precipitation periods or thaw, pointing to increasing pore pressures as a potential trigger to the slope failure (Paper IV).

Group B landslides are characteristic of the Pärnu and Reiu River valleys (Figure 4, Table 2). The section of the Pärnu River valley in Tammiste village is most affected by group B slope failures. Here the density of landslides is often so great that they form chains of failures along the valley slope. A nice example is Pärnu 1 landslide which formed of three individual failures occurring side by side along the river bank within a three-day period (Paper IV).

3.3. Geotechnical properties of soil

Strength properties of the uppermost marine sand in the study area were obtained from Kaljund and Mets (1976). The current study mainly focuses on the geotechnical properties of varved clay close to analysed landslide sites (Papers II and III). General geotechnical characteristics of clay will be presented as an example from the case study at the Sauga landslide (Figure 9, Paper II).
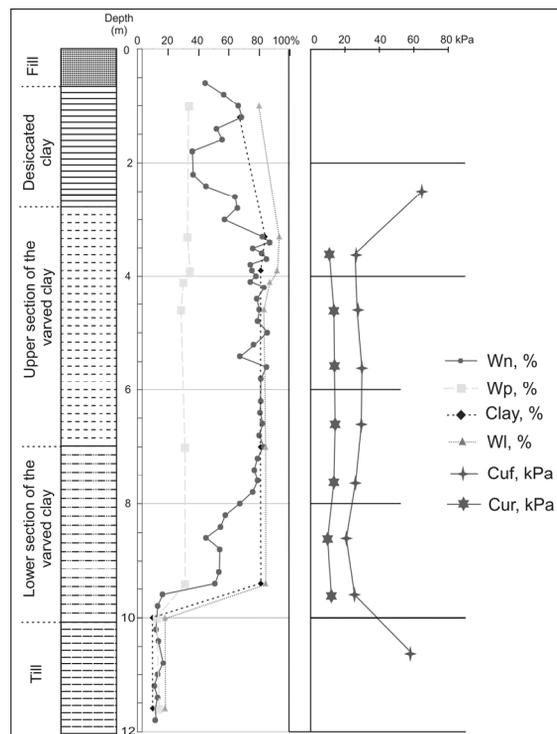

Figure 9. Geotechnical parameters of the varved clay at the Sauga test site and results of the vane shear tests. Wn-water content, Wl-liquid limit, Wp-plastic limit, Cuf-undrained shear strength, Cur-undrained residual strength (Paper II).

The uppermost clay interval at the Sauga test site, with a thickness up to 2 m, is desiccated varved clay. This layer may be locally absent and, if present, caps the upper section of the varved clay (Figure 9). The water content within the upper section of the varved clay is very close or may even exceed the liquid limit (Wl), pointing to the high probability of flowing behaviour of the sediment. In contrast, some samples from the uppermost desiccated clay layer have Wn values close to the plastic limit (Wp), indicating possible brittle consistency, especially during the high evapotranspiration periods. The lower 2–3 m thick section of the varved clay is resting on glacial till. Water content decreases gradually downwards from 80% to 50% within the lower section and may vary up to 20% between alternating silty summer and clayey winter layers, making the results of analyses heavily dependent on the actual sampling interval.

Undrained shear strength (Cuf) in Figure 10 drops sharply from 64 kPa to 26 kPa at the transition between desiccated clay (Wn ca 40%) and the upper complex of varved clay (Wn ca 80%). This follows the well-known correlation between water content and other geotechnical param, pointed out as characteristic of varved clays by many authors (Vilo, 1962; Võrk and Vilo, 1977; Kattel, 1989). However, the correlation between Wn an Cuf changes to positive within the depth interval 7.6–8.6 m as Cuf decreases from 25 kPa to 20 kPa and Wn from 80% to 50%. The decrease in Cuf values in the 7.6–8.6 m interval was double-checked with manually operated vane shear tests done within recovered sediment cores, and the results confirm the Cuf decrease. This decrease of strength downwards is unusual for the varved clay complex (the "common" profile can be seen in Figure 10) and the cause(s) is(are) currently unknown.

Slope stability analyses in case of the Sauga landslide (Paper II) showed that undrained shear strength measured with vane shear tests and effective strength values are too high for a landslide to occur. A river channel more than 4.5 m deep is required for the slope to fail. According to the local landowner and depth of the nearby river sections the actual river channel was around 3 m deep before the landslide occurred. Varved clay may loose its strength (softened) after the incision of the Sauga River into the varved clay plain. Due to the erosion of the upper part of the clay unit, deeper clay layers became relatively overconsolidated. The reduction of normal stress causes slow decompression and weakening of the once buried clay down to the critical state/fully softened strength values. If we use critical state strength parameters for varved clay in slope stability analyses, the river bank indeed becomes unstable when the channel reaches around 3 m in depth (Paper II).

3.3.1. Long-term changes in shear strength

In order to investigate probable long-term changes in the varved clay shear strength a series of vane shear tests was conducted at the Jaansoni test site (Figure 2). A similar series of vane shear tests was carried out at the same site in 1985 by the company IPT Projektijuhtimine OÜ. The results of those tests, together with second-order polynomial fitting line and 95% confidence limits, are presented in Figure 10. The result from two vane shear test series conducted in 2010 almost copy each other and show linear positive correlation with depth. The results from 1985 show a rather different pattern with a considerably (20–30%) weaker interval in the upper part (5–9 m) and a relatively stronger lower part of varved clay compared to 2010 results. This shows that the undrained shear strength of varved clay has changed within the 25-year period, from 1985 to 2010. During the same period the groundwater level rose significantly in the Pärnu area due to decreased consumption and relocated pumping stations but possible link between those two phenomena needs further investigation.
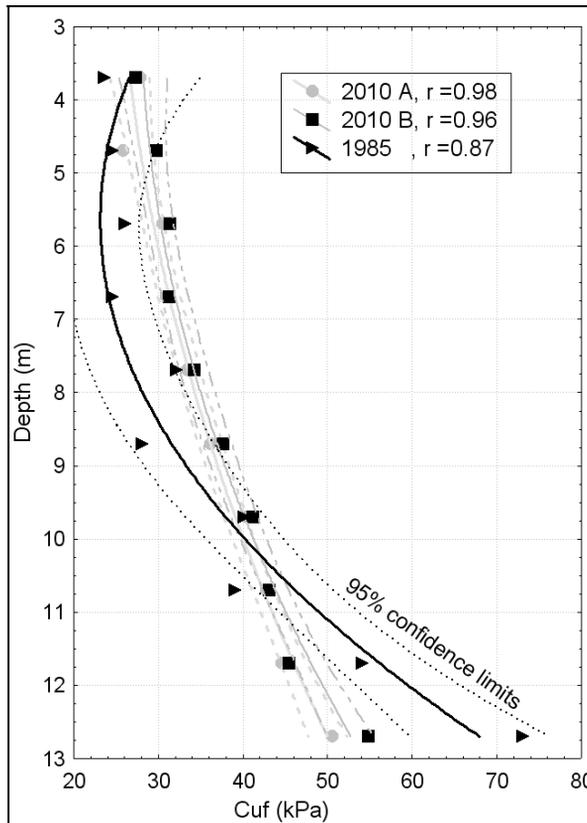

Figure 10. Undrained shear strength profiles from the Jaansoni test site conducted in 1985 and 2010. For location see Figure 2.

3.4. Groundwater conditions

Fluctuating groundwater levels are one of the most common triggers of landslides (Coduto, 1998; Clerici et al., 2002; Persson, 2008). In the current study we analysed long-term groundwater level changes in the Pärnu area. Also seasonal and short-term pore water pressures were measured at two test sites during the two-year period from 1 May 2009 to 1 May 2011. The effect of changing pore pressures on slope stability was also analysed (Paper I).

3.4.1. Long-term changes in the groundwater level

Talviste (2004) reports an up to 12 m rise of the piezometric level of the pressurized aquifer capped by varved clay during the period 1990-2000. This phenomenon could influence slope stability all around Pärnu town (Paper III). In order to understand long-term changes in the pressurized aquifer below the varved clay complex, the groundwater monitoring data available from the Pärnu area were analysed. A general development is described below and examples of interpolated isohypse maps for the years 1986 and 2001 are presented in Figure 11.

The groundwater depression during the 1960s was located in the old town and its center was at an altitude of -8 m. Depression gradually deepened and widened due to increased pumping and reached its maximum in the late 1980s. In 1986 the centre was lowered to an altitude of −12 m. The groundwater depression was partly refilled during the period 1990–2000 when the observation data show a mean 5 m (max 12 m) rise in the piezometric level within the town. The refilling was mainly caused by decreasing water consumption due to collapse of the Soviet era industry and rising water prices. The centre of the depression also moved eastwards due to extensive opening of new groundwater wells outside Pärnu, with corresponding lessening of water pumping from the wells located within the town. The piezometric level of pressurized groundwater in Pärnu town has been stable since the year 2001, at an altitude between 0 and 1.5 m with seasonal variation within 1.5 m (Paper I). The main problem of analysis is the inconsistency of the spatiotemporal distribution of data. However, the general development and evolution of the groundwater depression caused by water withdrawal for Pärnu town are still observable (Paper I).
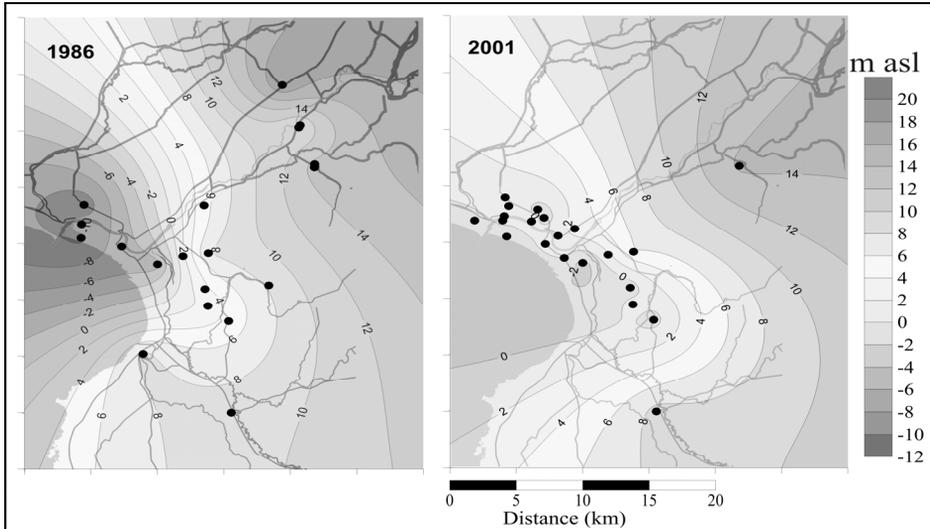
Figure 11. Interpolated isohypse maps marking the piezometric level of the artesian aquifer for the years 1986 and 2001. Black dots indicate monitoring wells (Paper I).

The interpolated isohypse maps in Figure 11 allow us to determine changes in the groundwater level between the monitoring wells and also somewhat outside the area covered by data. We can elicit that the pressurized groundwater level in 1986 was about 4 m lower than in 2001 at the Tammiste test site and 2 m lower at the Sauga test site (Figure 2). The Tammiste test site is located close to monitoring wells; therefore the above estimation is rather credible. The Sauga site is outside the reliable data coverage, thus leaving space for further discussions (Paper I).

3.4.2. Seasonal fluctuations of the groundwater level

Pore pressure is one of the major unknowns in slope stability calculations due to probable large variability both in space and time. In order to record the distribution and seasonal changes in the pore pressures within the studied slopes, the Tammiste and Sauga test sites were equipped with automatic piezometers at various depths. Pore pressure measurements were taken with an 1-hour interval during two-year period from 1 May 2009 to 1 May 2011 (Paper I). The Tammiste test site was chosen because of abundant marks of sliding events in the Pärnu River valley (Paper IV). The test site is situated ca 100 m downstream from the landslide Pärnu 2 (Table 2, Figure 4). The Sauga test site was set up in the middle of one of the most landslide prone areas in the Pärnu region (Paper IV), right at the top of the Sauga landslide (Table 2, Figure 4, Paper III).

3.4.2.1. Tammiste test site

Four automatic piezometers were installed in a vertical cluster set-up at the Tammiste test site in order to monitor pore pressures inside different lithological units (Figure 12).
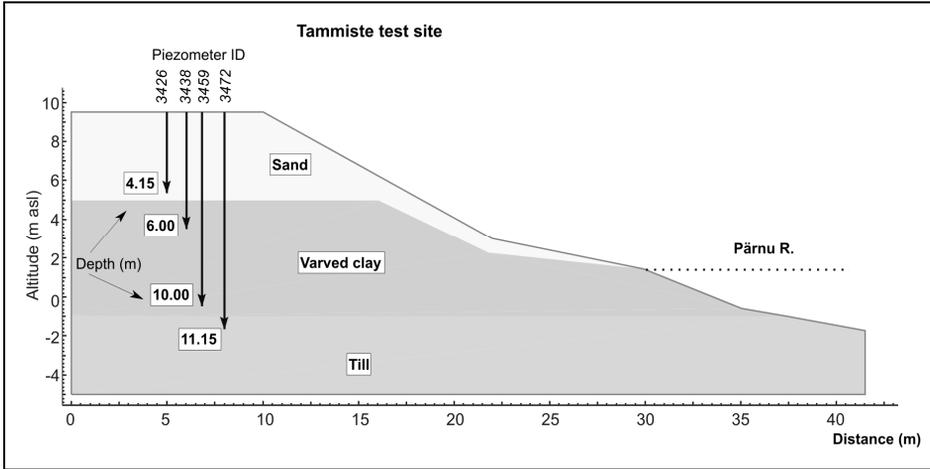

Figure 12. Geological setting and piezometers set-up at the Tammiste test site. For location of the test site see Figure 2.
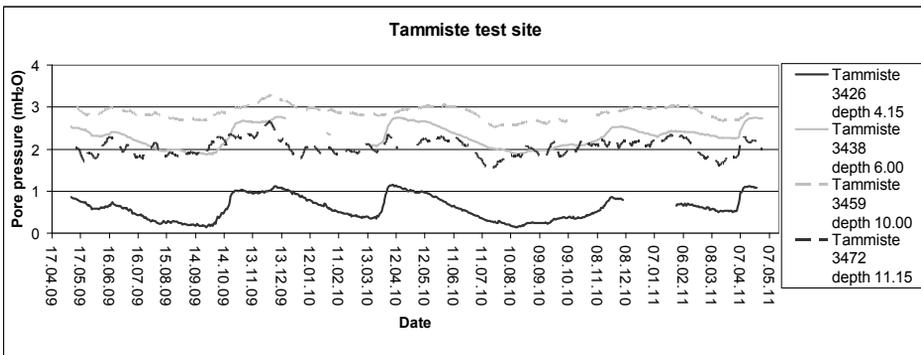

Figure 13. Graphs of the pore pressures, expressed as water column height above the filter, from the Tammiste test site. For location of the test site see Figure 2, for geological setting and piezometer set-up Figure 12. Gaps in the data series were caused by failed batteries.

Pore pressures fluctuated at the Tammiste test site within 1 m (Figure 13 and 18) and relatively coherently. Spearman's rank order correlation analysis between pore pressure datasets gave the lowest R=0.56. Cluster analysis divides pore pressure datasets into two first order groups but none of them forms first or second-order group with meteorological or hydrological data. The latter forms

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the second cluster, which is relatively weakly linked with the pore pressure group (Figure 14, Paper I).

Correlation analysis between pore pressures, meteorological and hydro-logical data revealed good correlation with temperature (R>–0.60) that probably caused by enhanced evapotranspiration during the snow-free period. Precipitation has generally small effect to the pore pressures during the snow-free period, even relatively rainy August (123 mm) and following September (99 mm) in 2010 didn't raise notably the pore pressures. Similar phenomenon is reported from Italy by Picarelli et al. (2004). The Tammiste test site has relatively thick unsaturated, buffering zone, that infiltrating water has to pass before the pore pressures starts to rise. The precipitation starts to influence the slope more profoundly then the 14 day average temperature drops below 5 ℃. Spring thaw, again controlled by the temperature, is the second period of rapid pore pressure rise. Pore pressures gradually lower between those high-stand periods due to seepage and evapotranspiration (Paper I).
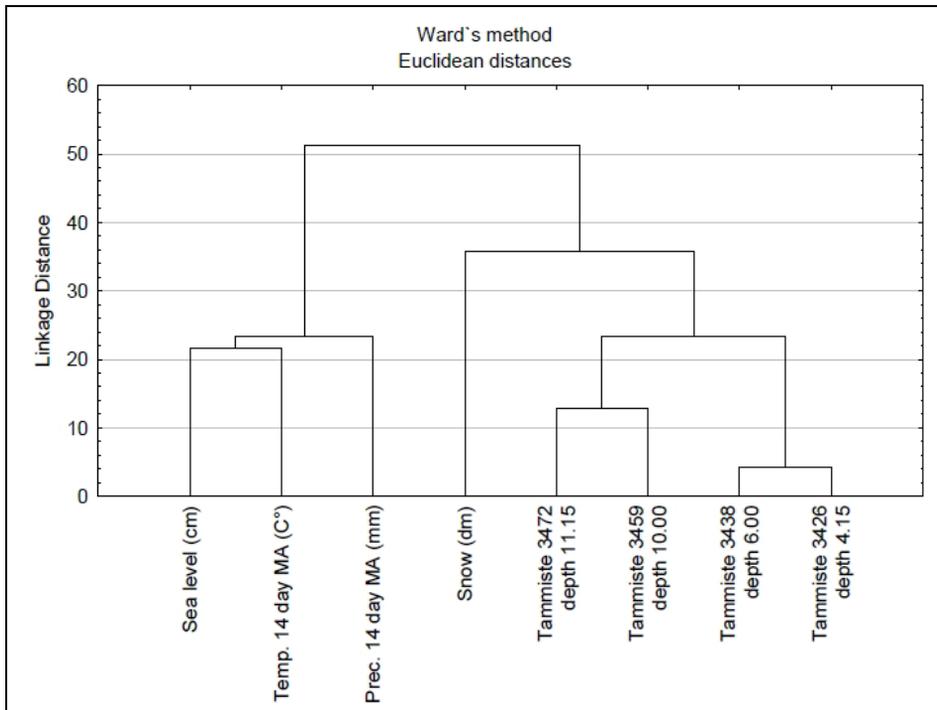

Figure 14. Hierarchical tree, showing relationships between pore pressures and meteorological and hydrological datasets at the Tammiste test site. MA – moving average. Shorter linkage distance means larger similarity (Paper I).

3.4.2.2. Sauga test site

Location of the Sauga test site is shown in Figure 2 and the geological setting together with the piezometers set-up in Figure 15. Piezometers were installed at different depths in order to cover sediments with different properties.
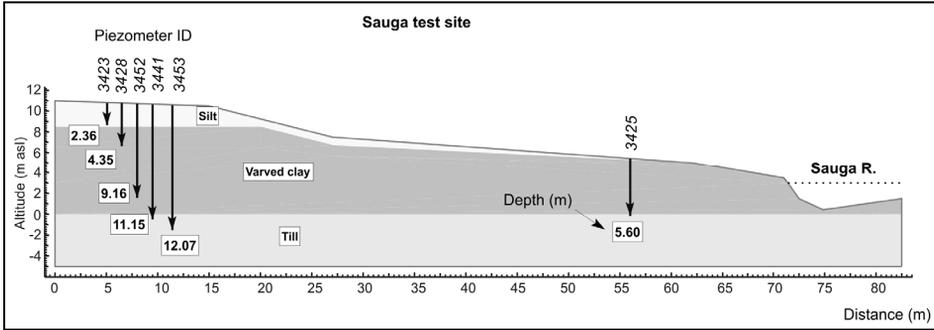

Figure 15. Geological setting and piezometers set-up at the Sauga test site. For location of the test site see Figure 2.
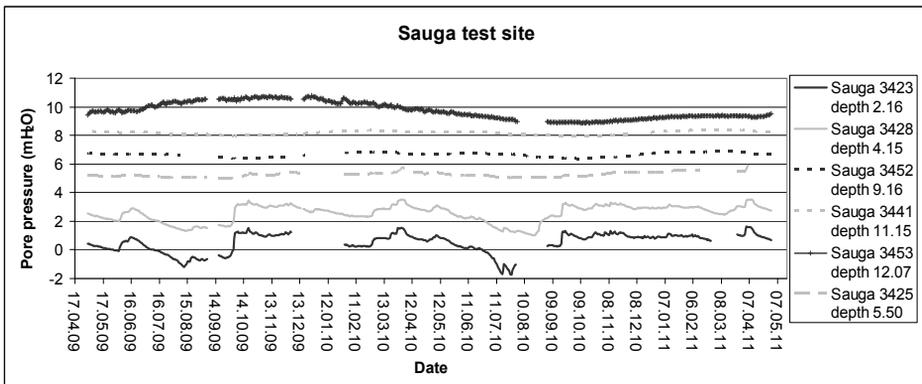
Figure 16. Graph of pore pressures expressed as water column height above the measuring point at the Sauga test site. For location of the test site see Figure 2, for geological setting and piezometers set-up Figure 15. Gaps in the data series were caused by failed batteries.

General pore pressure distribution loosely follows the hydrostatic distribution, with mild downward gradient (Figure 16 and 18). Seasonal variations are greatest (up to 2.6 m) and almost synchronous within the capping desiccated clay and the upper part of the varved clay layer (piezometers 3423 and 3428). Pore pressures in the lower part of the varved clay were very stable during the measuring period. Pore pressure variations in till (3453) are relatively large (ca 1.5 m) but changes are much slower than close to the ground surface (Paper I, Figure 16 and 18).

Almost total unison (R=0.99) in pore pressure fluctuations with a ca 5–6 hour time lag within the upper 4 m of sediment points to a water-conducting fissure system in the upper desiccated clay layer. Without this system, unison behaviour in the data series would have been very unlikely as the intact desiccated varved clay has very low hydraulic conductivity: ca 1^{-10} m/s (Kalm et al., 2006). A similar fissure system that brings surface water deeper within sediment has been described in French Alp overconsolidated glaciolacustrine clays (van Asch et al., 1996) and in Italian argillitic shales (Simon et al., 2004). Fissures do not extend deeper into the varved clay complex due to flowing consistency (the water content is close to or exceeds the liquid limit) of clay below the capping desiccated layer. This statement is supported by a very stable pore pressure in the lower clay complex, reflecting also very low (if at all) water exchange/ movement within this part. This is similar to findings by Kenney and Lau (1984) who stated that in overconsolidated clay slopes in Canada ground-water pressures remained essentially constant at depths greater than approximately 7 m below the ground surface. Till with a better hydraulic conductivity (10^{-5}–10^{-6} m/s) lies on top of water-bearing limestones and has correspondingly larger pore pressure variation. The aquifer bounded to till is isolated from the upper aquifer and therefore their fluctuation patterns are different. This grouping of pore pressures is supported by cluster analysis (Figure 17), as pore pressure datasets are spread between three subgroups (Paper I).

Three subgroups of piezometers also respond differently to changes in the meteorological and hydrological parameters. The first piezometer subgroup (3428, 3423, 3425) has strong (R~0.8) positive correlation with water level in the Sauga River through the entire measuring period. It also has strong (R~−0.8) negative correlation with temperature during the snow free period. Precipitation has only mild correlation (R~0.25) with pore pressures but its influence is not so strongly controlled by the temperature as it is at the Tammiste test site. At least 4 mm of average precipitation in form of rain during the 14 day period is needed to cause rise of the Sauga River water level and pore pressures in the upper part of the slope. The second subgroup (3441 and 3452) shows inverted correlations with temperature, the Sauga River and precipitation compared to the data from the upper piezometers. The precipitation has the strongest correlation (R~−0.35). Those differences between subgroups probably indicate considerable, smoothing time lag between external input and response of the deeper pore pressures but longer data series are needed to confirm this suggestion. The third subgroup (3453) has only very weak correlations (R<0.13) with all analysed parameters. The signs of correlations are the same as for the first subgroup, indicating a faster response time to the external para-meters compared to the second subgroup. On the other hand, very weaker correlations point to a larger time lag compared to the first subgroup and most likely the influence of the underlying, limestone-bound aquifer (Paper I).
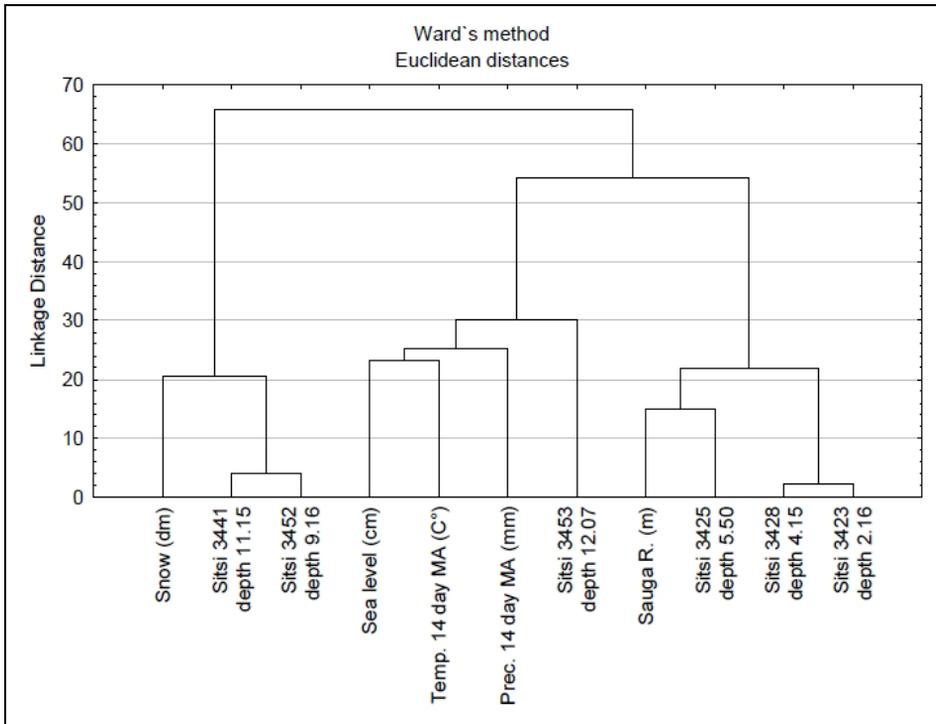
Figure 17. Hierarchical tree, showing relationships between pore pressures and meteorological and hydrological datasets at the Sauga test site. MA – moving average. Shorter linkage distance means larger similarity (Paper I).

The strength and direction of the correlations between analysed variables change as the snow cover develops. In the presence of the snow cover, the first pore pressure subgroup is pretty well correlated with temperature (R~0.75) and sea level (R~0.68). Good correlation between sea level and pore pressures is not most likely casual as those parameters were simply very stable in the presence of the snow cover. Pore pressures deeper within varved clay (second subgroup) had inverted, very weak or statistically insignificant correlation coefficients compared to the first subgroup. Generally correlations were weak or even statistically insignificant. Pore pressure in till (third subgroup) showed similar rather weak and inverted correlations as the second subgroup, the strongest one with precipitation (R~–0.4). Those correlations are hard to explain as the relations are probably blurred by different time lags and the influence of the underlying aquifer; some may not be casual at all (Paper I).

Pore pressures in till measured from the top (3453) and foot (3425) of the slope differ consistently (Figures 16 and 18). This disparity is probably mostly caused by influence of the Sauga River. The first piezometer is also located close to the secondary rupture surface penetrating the Sauga landslide body (Paper II). Both reasons may increase the influence of precipitation and evapotranspiration to

the pore pressure in till. The Sauga River influence to piezometer 3425 is also evident in Figure 17, as they form first-order class (Paper I).

3.4.3.3. Comparison of the two test sites

The pore pressures measured at two test sites show a different distribution and response dynamics to the same meteorological and hydrological changes. Median values, annual variation and distribution of the pore pressures measured at the test sites are shown in Figure 18.
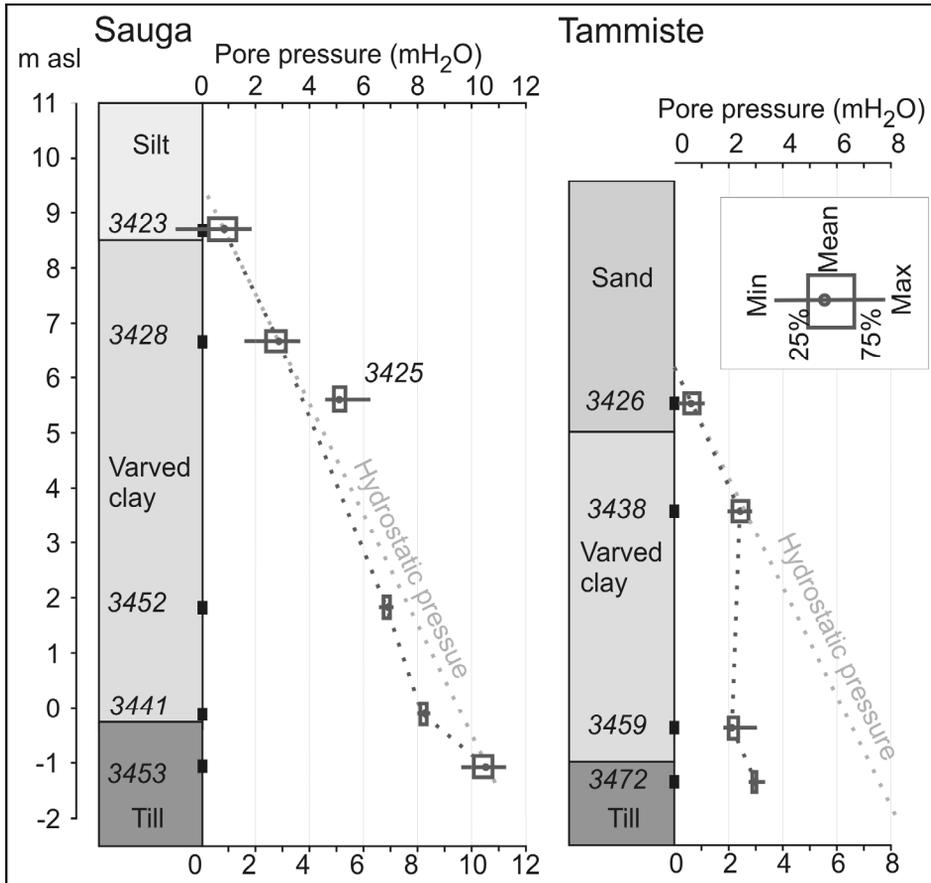
Figure 18. Box and whisker diagram showing the distribution, variance and median values of pore pressures at the Sauga and Tammiste test sites within a two-year period. Piesometer 3425 was installed in till, 45 m towards the river from the main piesometer cluster (Figure 15). For location of the test sites see Figure 2 (Paper I).

Pore pressures in the clayey slope at the Sauga test site have a much larger seasonal variability than the sandy slope in Tammiste. This can be explained by

drainage conditions as the slope at the Tammiste test site consists mainly of water-conducting sand and has steeper slope angle, allowing faster drainage of infiltration water without notable influence on pore pressure. Also, the slope at the Tammiste test site has roughly 3 m thick unsaturated vadose zone above saturated sediments whereas at the Sauga this zone is only 1 m thick. The unsaturated zone buffers the infiltrating water, therefore smoothing pore pressure response to the changing meteorological conditions. For example, pore pressures, expressed as water column height in the upper part of the clayey Sauga slope rose 1.4 m in two days in response to the autumn rains in 2009, while it took 25 days for pore pressure to raise 0.7 m rise at the sandy Tammiste site. On the other hand, the influence of the rainy autumn in 2009 is evident even in the deepest pore pressure measurements at Tammiste, while varved clays shielded completely its effect at the Sauga site. Extended periods (probably years) of increased precipitation are needed to raise pore pressure in the whole slope at Sauga, whereas slope as a whole at the Tammiste site is more sensitive to single extreme rainstorms or shorter rainy periods (weeks). Various authors point out that extended wet periods are needed to cause deeper landslides whereas shallow slope failures are triggered by single rainstorms (van Asch et al., 1999; Picarelli, 2007). Soil permeability, functioning either through interconnected pores or crack systems, plays an important role in controlling propagation of pore pressure within the slopes, especially in the clayey slope at the Sauga test site (Paper I).

The piezometric level of the artesian groundwater below varved clay is lower at the Tammiste test site (ca 7 m) probably because of its relative proximity to the Pärnu River channel and to Pärnu town. The former drains till and the latter generates a local groundwater depression that lowers the piezometric level by ca 2 m (Figure 11).

3.5. Failure mechanism and changes in slope stability

Slope stability modelling was used as a tool to analyse the failure mechanism (Papers II–IV) and variation in the slope stability due to changing pore pressures (Paper I). Additional knowledge was obtained from the morphological features, position and timing of the landslides. Retrogressive failures are characteristic of group A landslides (Paper IV, Table 2) in clayey slopes near to the Sauga and Audru Rivers. Six stages of failures occurred in Sauga landslide (Figure 19, Paper III).
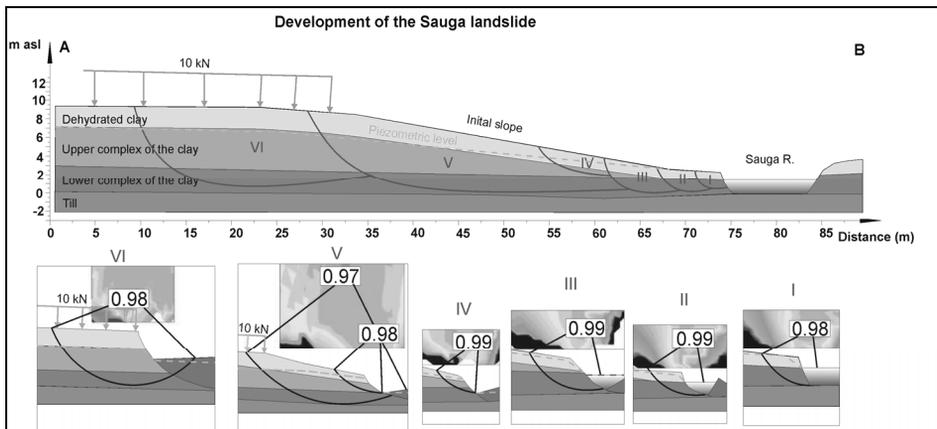
Figure 19. Development of the Sauga landslide according to slope stability modelling (Paper III). For location of the landslide see Figure 2.

Two general mechanisms have been proposed for retrogressive landslides (group A landslides in the current study) in sensitive clays (Quinn et al., 2007). One approach describes the formation of the retrogressive complex via a series of rotational slumps developing retrogressively to the eventual headscarp and the other suggests nearly simultaneous liquefaction beneath the entire landslide, followed by subsidence and lateral spreading of the upper un-liquefied slide mass. Haug et al. (1977) explained the development of the retrogressive slide complex in Canada, within glaciolacustrine clays and covering sands, with series of individual failures that exploit a common failure surface. The leading blocks offer reduced restraining force on the following ones as they move faster and the reduction of the lateral support decreases upslope. The final scheme is very similar to the one presented in Figure 19 (Paper III). Van Asch et al. (1984) also explain the formation of the retrogressive landslide complex in glaciolacustrine clays of the French Alps with a series of retrogressing individual failures. Our stability analyses (Papers II–IV) and morphological features of the landslides primarily support the first mechanism, the formation of the retrogressive complex through successive failures of individual blocks. The rupture surfaces are linked tangentially, "exploiting" former failure zones. There are some evidences of partial liquefaction in the ruptured zone, especially in case of the Audru 1 landslide. Varve correlation between the drill cores indicate that up to 2.8 m of varved clay was liquefied and pressed out during the multiple failures from the lower part of the landslide body (Paper II).

Sandy slopes (group B failures) inherit a different failure mechanism. The first slope stability models (Paper IV) indicated that those slopes should have remained stable at the angles up to 30°, which is much higher than the minimum ruptured slope angles (ca 20°) measured in the field. As sandy slopes have always failed after heavy rains or thaw, the water movement inside the slope

was considered to be of crucial importance. Therefore, slope stability was analysed together with the finite element groundwater modelling that allowed inclusion of the additional shear stress generated by moving groundwater. When additional shear stress was included into the model, the critical slope angle for this type of landslides decreased to 20º, being thus in accordance with the field data (Paper IV). Besides the viscous drag of flowing groundwater, the seepage may cause erosion at the slope base and ultimately lead to slope undercutting (Budhu and Gobin, 1996). Both mechanisms were simulated in laboratory by Chu-Agor et al. (2008), who found that seepage erosion often precedes the seepage-induced failure, undercutting the slope until it becomes unstable. Seepage erosion is not reported from the study area so far and remains a subject for further investigations.

Rise in pore pressures reduces effective normal stress within the slope and causes therefore reduction of the frictional component of the soil strength (Iverson et al. 1997). This mechanism, in turn, destabilizes slopes and therefore the factor of safety (FoS) of the slope fluctuates together with pore pressures. Those changes at both test sites are displayed in Figures 20 and 21. The entire analysed slope is divided into bins, 2 m (Figure 20) or 4 m (Figure 21) wide. Their location on the slope profiles is given in the upper parts of Figures 20 and 21. All the analysed slip surfaces are allocated between the bins according to the intersection points between the individual slip and ground surface. The FoS of the weakest slip surface within each bin is plotted in the lower parts of Figures 20 and 21. Four FoS curves along the slope characterize both seasonal fluctuations and long-term changes in slope stability at the Tammiste and Sauga test sites (Paper I).

Seasonal slope stability variations at the Tammiste test site (Figure 20) are far smaller than at the Sauga test site (Figure 21). Median seasonal variation of the FoS is only 0.08 at the Tammiste site and the largest changes occur close to the Pärnu River channel, indicating the importance of the river water level. The two lowermost points of the FoS curve, within bins 4 and 16 in Figure 20, mark intersections between the global minimum slip surface (FoS=1.3) and the ground surface. The piezometric level of the underlying artesian aquifer has also a small effect on the slope stability at the Tammiste test site. The slope was only marginally more stable during the low-stand during the 1980s (Figure 11); then the piezometric level was about 4 m lower than now. Hypothetical increase in the piezometric level (+ 2 m) has a minor effect on slope stability (Figure 20). The pore pressures generally little affect the slope stability at the Tammiste test site. The slope failures near the Tammiste site are mainly controlled by the river undercutting and water level in the channel. High water level in the river affects the slope by lessening of the slope stabilizing capillary forces during the growth of the saturated zone (Krahn et al., 1989; Rinaldi and Casagali, 1999). Theoretical calculations are supported by experiments with artificial sand slope made by Jia et al. (1996). These revealed initial failures during the rise in the tank water level, explained by loss of negative pore pressures and a second series of failures during the drawdown, caused by excessive pore pressures and seepage forces.

Figure 20. Slope stability changes due to pore pressure fluctuations at the Tammiste test site together with the slope profile (Paper I). Gw – artesian groundwater level, FoS – factor of safety. For location of the test site see Figure 2.
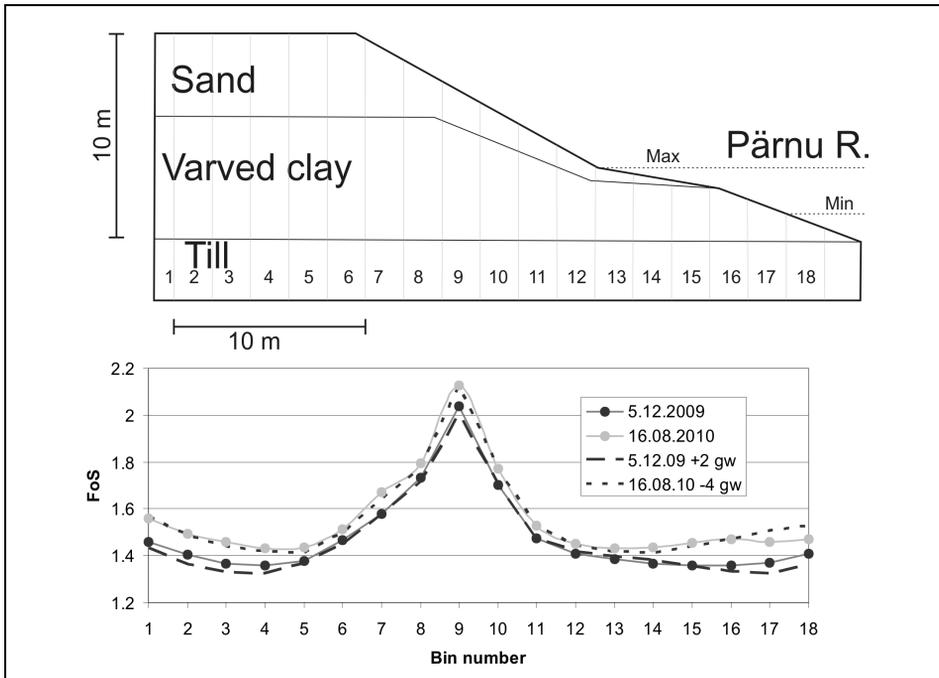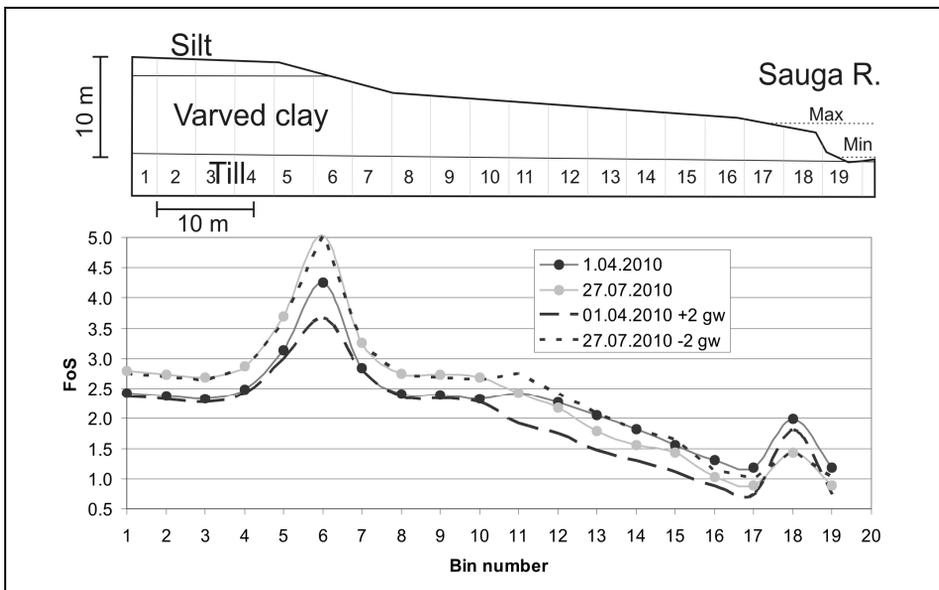

Figure 21. Slope stability changes due to pore pressure fluctuations at the Sauga test site with slope profile on top (Paper I). Gw – pressurized groundwater level, FoS – factor of safety. For location of the test site see Figure 2.

Seasonal changes in the FoS at the Sauga test site can be as large as 0.77, the median being around 0.34 (Figure 21). A similar fluctuation range was calculated for Canadian clay slopes by Kenney and Lau (1984). The largest fluctuations occur in the upper part of the slope, together with the most variable pore pressures (Figure 18). The water level in the Sauga River influences the slope from bin 10 to 19 where the high-stand FoS curve (01 April 2010) crosses the low-stand one (27 July 2010). This equals the distance of up to 40 m upslope from the river channel. The Sauga River bank is unstable (FoS>1) during the low-stand of the river up to 15 m upslope from the channel. The ongoing slope failures (Paper III) support this statement. The modelled slope configuration is stable during the high-stand of the Sauga River (01 April 2010), even if it is concurrent with elevated pore pressures (Figure 21). The turning point occurs when the river level falls below 2.5 m asl. It should be noted that the pressurized aquifer did not have its maximum piezometric level at the same time as the upper one. Data from monitoring wells show general seasonal fluctuation within 1–2 m, but timing of the peaks is scattered within the seasons. The slopes would have an (slightly) lower FoS than the high-stands of both aquifers would occur at the same time (Paper I).

The underlying pressurized aquifer influences slope stability up to 40 m upslope from the Sauga River (bin 10 in Figure 21). The lower part of the Sauga slope had a 0.2–0.3 higher FoS than now during the aquifer low-stand in the 1980s (Figure 11), but the upper part of the slope was not affected at all. Hypothetical rise in the piezometric level by 2 m would also influence mostly the lower part of the slope. Its FoS would decrease as much as 0.4–0.6. This phenomenon would cause larger landslide complexes as clay slopes near Pärnu are known to fail in a retrogressive manner (Papers I–IV).

As fluctuations of pore pressure may cause slope failures, it is important to consider the exogenic factors that influence pore pressures. Climatic analyses have shown that precipitation has an approximately 30-year cycle in Estonia and currently (2011) we are close to the peak of one cycle (Jõgi and Tarand, 1995). Annual precipitation in the Pärnu area has increased by 25% during the past 50 years – from ca 600 mm/y in the 1960s to ca 780 mm/y in the 2000s (Jaagus and Raudsepp, 2008). Both trends should lead to increasing pore pressures in the Pärnu area. The influence of these trends could also be mani-fested in groundwater levels, but it is unfortunately well masked by ground-water withdrawal (Paper I). There is only one monitoring well (No. 19270) opening the uppermost aquifer in the central part of Pärnu town which has a sufficiently long measurement series (1965–1986). Data from this well show an indeed mild (+0.25 m), statistically significant rise in the groundwater level during the measurement period, but this may also be caused by some local anomaly like leaking water pipes or unaccounted subsidence of the ground level, known to exist in Pärnu town (Vilo, 1986) (Paper I).

The general high precipitation background amplifies the influence of extre-me events like severe storms. Concurrently the number of both extremely dry and wet days has risen during the period 1957–2006 (Tammets and Jaagus,

2007). Because of the increasing cyclonic activity (Rajasalu and Keevallik, 2001; Tomingas, 2002; Jaagus, 2006), low coastal plains in NW Europe have suffered from an increasing number of extreme storms (Alexandersson et al., 1998; Paciorek et al., 2002) and corresponding flood surges (Suursaar et al., 2006; Jaagus et al., 2008). The accompanying strong winds may concentrate precipitation even further at the slopes (Guthrie et al., 2010). Extremely rainy days will cause faster rise in pore pressures in clayey slopes near Sauga River. Extended dry periods may cause further dehydration of the uppermost part of varved clay. Dehydration in turn causes decrease in the clay volume (shrinkage) and therefore the formation or deepening of the water-conductive fissure system (Paper I).

Ongoing climate change manifests itself in Estonia mostly by rising temperatures and increasing precipitation during spring and winter (Jaagus, 2003). Both trends are also predicted to continue (Graham et al., 2008). Thus mild, snowy winters, followed by early spring with warm, rainy westerly cyclones, will probably be more common in the future. Spring storms that bring rain and strong warm wind will melt snowpack most effectively. An up to 25% increase in infiltration rates was measured by Harr (1981) in Oregon during the snow on rain events. Elevated spring runoff will also lead to more intense river bank undercutting, which is the most prominent destabilizing geological process nowadays. On the other hand, rising temperatures cause increase in evapotranspiration, which may almost completely compensate the effect of increasing precipitation on pore pressures (Persson, 2008). Still, evapotranspiration is highest during the vegetation period, clearly evident in Figures 13 and 16 as a steady drop in the pore pressures at the upper parts of the slopes during the summers of 2009 and 2010. Pore pressures during high-stand periods, early spring and late autumn, are less influenced as temperatures are usually below 5 °C and therefore evapotranspiration is inhibited. Combined effect to of the above-mentioned climatic trends and increasing human activities on slope activity has not yet fully investigated but is most likely destabilizing (Paper I).

3.6. Landslide susceptibility mapping

Detailed DEMs for the investigated valley sections (Figure 2) served as a basis for landslide susceptibility analysis (Paper IV). Comparison of the collected geological data and the existing slope angles with the critical values derived from the geotechnical modelling and field surveys allows allocation of stable and potentially unstable sections of river valleys. According to susceptibility zonation, slopes in marine sand are seldom unstable but most of the river valley slopes in glaciolacustrine varved clay may fail, particularly if human activities are involved. Clay slopes are least unstable in the valley of the Sauga River and most stable in the Audru River valley (Table 3). As the investigated area is currently facing fast urban development, possible human impact on the slopes was considered in potential landslide susceptibility evaluation. As human im-

pact can be very many-sided and includes different factors, a simplification was used and the safety coefficient 1.4 was applied as a divisor to the modelled critical slope angles. This is the coefficient that, according to Estonian standards for geotechnical investigations (EVS 1997-1:2003), should be applied to slope stability calculations in order to secure constructions design. After this modification we may conclude that the slopes in glaciolacustrine clay and in marine sand can be considered stable if the slope angles are less than 7º and 14º, respectively (Paper IV).

Table 3. Landslide susceptibility mapping results as percentage of the slope type from the investigated valley section. Critical slope angles were derived from the slope stability modelling discussed in the text.

Type of the slope		In % of the length of the slope type from the studied valley section			
		Audru River W bank/E bank	**Pärnu River W bank/E bank**	**Sauga River W bank/E bank**	**Reiu River W bank/E bank**
Glaciolacustrine clay	> 7°	84.4/88	97.6/79.2	100/97.1	88.6/90.7
	>10°	46.4/33.6	79.2/13.2	89.5/75.5	42.9/53.1
Marine sand	>14°	9.6/8.8	0/0	11.7/3.4	4.2/1.6
	>20°	1.6/0.8	0/0	1.1/0	1.2/0

Most of the published landslide susceptibility maps deal with vast and/or remote areas and are often based on remote sensing methods like ortho- or satellite image analysis (Geertsema et al., 2006), combined with general geological information. The most used technique for susceptibility maps is multivariate analysis. Basic workflow is as follows: mapping of landslides and correlating factors (slope angle, aspect, rainfall intensities, etc.), establishing relationships between correlating factors and instability and finally classifying terrain units into domains of different landslide susceptibility (Clerici et al., 2002). In our study the second step in the workflow was completed with the "worst case" slope stability modelling – probably the most direct approach possible. However, it is suitable only in small-scale, relatively uniform susceptibility mapping as failure types and controlling parameters tend to vary too much in case of regional analysis with rigid terrain.

The landslide susceptibility mapping method presented in Paper IV assumes that in the future slope failures will occur under those conditions which led to past instability. Therefore this method needs updating as controlling factors change. Clerici et al (2002) brought out five factors considered to have the closest bearing to slope instability: lithology, slope angle, slope-bedding relation, rainfall and land use. Land use is the most varying factor. It may be

even so erratic, either naturally or human activity induced, that some authors do not consider it as a reliable representation of those past conditions under which landslides were generated (Clerici et al., 2002). Therefore in the "worst case scenario" presented in Paper IV, critical slope angles with human influence should be a baseline for planning any kind of activities close to river valleys in the study area.

The analysis in Paper IV offers a simple methodology for high-resolution (4 × 4-m grid) landslide susceptibility zonation. Detailed background data with at least the same resolution are needed for this task, covering the geological setting and topography of the area. As so accurate geological data sets covering larger areas are most probably unavailable, extensive field work is needed to obtain geological background information. Geological field mapping also enables more detailed landslide inventories than orthophoto-based approaches, as only two out of nine landslides investigated in 2002 can be identified in high-resolution (pixel 0.4 m) orthophotos. Apparently, in the conditions of relatively small slides with low scarps and rapid revegetation, orthophoto-based landslide mapping is not sufficient and should be accompanied with an extensive field survey.

Recently published LIDAR based digital terrain model from Estonian Land Board, may also reveal currently unknown sites. LIDAR data has been successfully used in many cases around the world for landslide mapping and as a base for landslide susceptibility maps (Schulz, 2006; van den Eeckhaut et al., 2007, Razak et al., 2011). Nowadays, it is the density and accuracy of geological, hydrogeological and meteorological information that is likely limiting the resolution of the susceptibility map not the quality of the DEM (Carrara et al., 1999; van Western et al., 2006). As our initial landslide susceptibility mapping was based on slope angle analysis, some cells that were considered "safe" actually may fail because retrogressive failures can reach even entirely flat areas up to 70 m away from the river (Paper IV). This problem with retrogressive slides is also pointed out by Quinn et al. (2010).

3.7. Monitoring the slope development

The Sauga landslide, covered in Paper III, was monitored after the major slope failure in December 2005 in order to follow post-failure slope development. Repeated measurements of fixed points and comparative photos show ongoing movements due to erosion of the landslide body that completely filled the river channel after the first failure in 2005. The results presented in Figure 22 include only those markers that have moved more than the measuring error during the monitoring period (in practice >10 cm horizontally and >2 cm vertically). The initial marker setup can be found in Paper III, Figure 9.
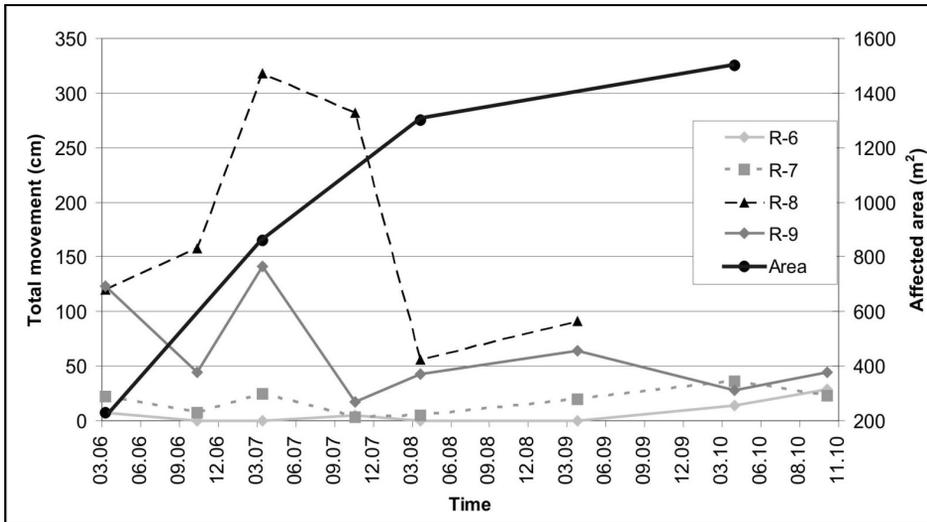
Figure 22. Summarized vertical and horizontal movements of the markers and re-activated area in the Sauga landslide. For location of the landslide see Figure 2.

New, smaller-scale movements occurred close to the river channel and progressed upslope due to river undercutting. This process is particularly clear in the NW part of the landslide where marker R-8 is located. One marker (R-10) was eroded away a month after its installation. A second marker (R-11) was destroyed by earthworks in March 2008. Marker R-8 shows the largest measured movement: 2.44 m of vertical and 5.07 m of horizontal displacement. The upper part of the slope has so far been stable, as markers located in that part have not moved more than the measurement error (Paper III).

Some authors have suggested large-scale creep within varved clays in western Estonia (Olli and Martin, 1961; Kaljund and Vilo, 1967). This process would lower soil strength towards the residual level, possibly leading to slope failures (Mesri and Shahien, 2003). According to monitoring data, the process is currently going on at the Sauga landslide site. Measurements showed soil movement during three years at marker R-7 before new tension cracks opened and a small scarp formed in April 2009. The behaviour of the new movements within the Sauga landslide copies the main landslide evolution in a smaller and slower scale. The mechanism is that a small-scale slide occurs close to the river channel due to undercutting and is followed by a larger-scale slide that exploits the newly created weak zone (i.e. a slip surface or zone of the initial sliding stage) as the area affected by displacement enlarges progressively (Paper III). Wang and Sassa (2003) reproduced a similar failure mechanism in laboratory.

3.8. Mitigation measures

Mitigation measures can be divided into active (drainage, etc) and passive (various exclusion zones) measures. However, both types must be based on the knowledge of the underlying mechanics and triggering factors of the considered slope failures (Crozier and Glade, 2005). Active measures, either stabilization of an existing landslide or prevention of a pending slope failure, must reduce the driving forces or increase the resisting forces (Popescu and Sasahara, 2008). As an example of the passive measure and result of the current study, an 80 m wide building exclusion zone along the Sauga River valley was approved by the Sauga Parish in 2008 (Paper IV). This distance was derived from field investigations and slope stability modelling which set the outward limit of retrogressive slides (group A) to 70 m from the river channel. A 45 m wide restriction zone would be sufficient in Tammiste village at the Pärnu River, where group B landslides are most abundant.

Active countermeasures have so far been utilized in the Sauga landslide. The applied methods, modification of the slope profile and drainage, are the first and the second most used stabilizing methods in engineering practice (Popescu and Sasahara, 2008). By implementing those methods, the FoS of the once failed slope were raised from initial 1.2 to around 2 (Kalm et al., 2007). The third proposed countermeasure was reinforcement and reshaping of the Sauga River channel in order to reduce the main preparatory process of the landslide(s) – the river erosion. This part of the project was not implemented due to high cost. As a result, next spring flood in 2006 eroded most of the channel-filling landslide body and new failures started to develop (Paper III).

A long-term and cost-effective active countermeasure would be afforestation of river banks to reduce erosion and increase evapotranspiration. Relatively small-scale slides are, among other factors, conditioned by the strength of the vegetation root network. This factor is difficult to estimate as it depends on factors like vegetation type and density, plant species or age (Ekanayake and Phillips, 2002). According to Roering et al. (2003) root cohesion can be as high as 15.2 kPa and may significantly increase slope stability in case of small-scale, shallow slides. Shrubs or trees on the river banks could decrease the susceptibility to initial failure that may lead to large retrogressive landslide complexes in case of the Sauga River. The stability of the sandy slopes near Pärnu and Reiu Rivers, where single rotational failures prevail, would benefit from trees and shrubs via enhanced evapotranspiration.

CONCLUSIONS

The following conclusions can be drawn from the current thesis:

- Altogether 26 landslide records from the study area were included into the landslide database. The largest one measures 200 x 100 m and occurred in glaciolacustrine varved clays. Most of the landslides take place on convex slopes of the river valleys. Landslide inventory revealed clear clustering of the slope failures.
- Landslides form two distinctive groups: larger, retrogressive landslide complexes affecting clayey slopes of the Sauga and Audru River valleys and smaller, single rotational slope failures affecting sandy slopes in the Pärnu and Reiu River valleys.
- The frequency of landslides has grown since 1900, partly due to deforestation of the river valleys and human-induced changes in the groundwater regime.
- River erosion is the main preparatory process of landslides. It is accompanied by strength loss of glaciolacustrine varved clay. Therefore critical state (fully softened) strength parameters should be used in initial slope stability analysis in case of clayey slopes. Water-related phenomena, like excess pore pressures, seepage, loss of negative pore pressure, etc., are the main triggers for slope failures.
- The Sauga River valley is the most unstable region in the study area according to landslide susceptibility mapping. Building exclusion zone (80 m) along most of the Sauga River valley was proposed by researchers and approved by Sauga Parish.
- Slope stability varies seasonally due to changing pore pressures within the slopes. Pore pressure fluctuations have a strong effect on the stability of clayey slopes. The stability of the sandy slopes is less affected.
- Temperature and related thaw are two most important factors controlling pore pressures. Predicted climate change will have notable destabilizing effect to clayey slopes.
- The underlying artesian groundwater level and its human-induced fluctuations affect slope stability far outside the settlements. Clayey slopes are much more sensitive to changes in the piezometric level than sandy slopes.
- Reforestation and drainage of river banks would decrease river undercutting and stabilize valley slopes most effectively.

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SUMMARY IN ESTONIAN

Maalihked Lääne-Eesti savipinnases

Käesolev doktoritöö annab ülevaate Pärnu ümbruses toimunud maalihete levikust ja selgitab nende tekkepõhjuseid ning mehhanismi.

Sageli seostatakse maalihkeid mägiste ning sademeterohkete aladega, kuid ka tasase pinnamoega Eestis toimuvad gravitatsioonist tingitud pinnase liikumised. Esimesed teated maalihetest ulatuvad mitmesaja aasta taha, kui Vigalasse ehitatud vastne mõisahoone jõkke libises. Teaduslikult hakati maalihkeid uurima 1960. aastatel, kui Pärnus ja selle lähemas ümbruskonnas toimusid mitmed küllaltki suured nõlvade purunemised, mis põhjustasid ka märkimisväärset varalist kahju. Pärnu linnas toimunud lihete puhul leidsid tollased uurijad, et peamine maalihkeid vallandav tegur on jõekallaste ülekoormamine täitepinnasega inimese poolt. Nende uuringute tulemusena nimetati Pärnu ümbrust Eesti kõige maalihkeohtlikumaks piirkonnaks. Peale suhteliselt vaikseid 1980.–90. aastaid sagenesid teated Pärnu lähikonnas toimunud maalihetest taas järsult 2000. ning sellele järgnevatel aastatel. Neist suurim oli 2005. aasta lõpus Eametsa külas toimunud lihe, mis liigutas ühekorraga ligikaudu 60 000 m^3 pinnast ning ohustas vägagi tõsiselt lähedal paiknevaid eramuid. Tollal hoogu koguva kinnisvarabuumi taustal süvenes kohalike omavalitsuste mure nõlvade püsivuse osas, sest puudusid andmed maalihete leviku, omaduste jne kohta ning samal aja suurenes surve uute eramute ehitamiseks jõgede kallastele.

2002. aastal viis Tartu ülikooli geoloogia osakond koostöös Sauga vallavalitsusega läbi uuringu, et kaardistada Pärnu ümbruses toimunud maalihked ning kogutud info põhjal koostada piirkonna esialgne maalihete ohtlikkust määratlev kaart, mis oleks aluseks uute ehitiste planeerimisele. Suhteliselt detailse lihkeohtlikkuse kaardi koostamiseks kasutati kõrgusmudelit, välitöödel ja kirjandusest kogutud geoloogilist andmestikku ning geotehnilise modelleerimise abil leitud kriitilisi nõlvakaldeid erinevate geoloogiliste situatsioonide jaoks. Liivanõlvad osutusid ohtlikuks alates nõlvakaldest 20º ja savinõlvad alates 10º kaldenurgast. GIS tarkvara abil eraldati geoloogilise kaardi ja kõrgusmudeli põhjal välja potentsiaalselt ohtlikud piirkonnad. Uurimuse käigus tehtud inventuuril määratleti kõige ohtlikumaks piirkonnaks Pärnu ümbruses valdavalt viirsavidesse erodeerunud Sauga jõe org. Samas on paksu (>3 m) merelise liivakattega tasandikesse lõikuvad jõeorud (nt Reiu jõel) üldiselt stabiilsed. Selged maalihete koondumiskohad asuvad Tammiste külas Pärnu jõe ääres, Eametsa külas Sauga jõe orus ning väiksem maalihete kogum Audru jõeorus, Audru alevikust paar kilomeetrit ülesvoolu.

Ühtlasi selgus, et liiva- ning savinõlvadel toimuvad lihked on erinevad. Liivanõlvadel toimuvad tüüpiliselt üksiklihked, mis ei ulatu jõeorust üldiselt välja ja millele iseloomulik esinemiskoht on Tammiste küla Pärnu jõe ääres. Valdavalt viirsavist koosnevatel nõlvadel toimuvad aga mitmest osalihkest koosnevad maalihete kompleksid, mis võivad ulatuda küllaltki kaugele jõeorust väljapoole. Savinõlvadel toimunud maalihke silmapaistev esinemiskoht on Eametsa küla

Sauga jõeorus. Seal toimus ka aastal 1976 siiani suurim teadaolev maalihe, mis liigutas hinnanguliselt 100 000 m^3 pinnast 200 x 100 m suurusel maa-alal.

Inventuurile järgnenud detailsemad uuringud ja modelleerimine keskendusid peamiselt kõige suurematele, Audru ja Sauga jõe orgudes asuvatele savilihetele. Modelleerimise tulemusena selgus, et liivalihete vallandavaks teguriks on põhjavee voolamisest tingitud lisapinge suhteliselt järskudes liivanõlvades. Liivanõlvad on kõige ebastabiilsemad kevadise suurvee ajal. Savinõlvade lihke-kompleksides toimuvad lihked üksteisele järgnevalt, kasutades ära eelneva purunemise käigus tekkinud nõrgemat lihkepinda. Lihetest haaratud ala suureneb iga järgneva purunemise korral kahe-, kolmekordselt ning esimese lihke toimumist vahetult jõe kaldal kontrollib peamiselt jõe poolt pidevalt kulutatava nõlva kallakus ning jõe veetase. Savinõlvad kõige ebastabiilsemad vahetult kevadise suurvee järel, kui veetase jõesängis on juba langenud, kuid nõlva sees on veel kõrge. Viimast väidet kinnitab ka teostatud Sauga lihke järelmoni-tooring, mis näitas kiiret lihkekeha ärakannet jõesängist ning erosiooniga kaasnevat uute maalihete teket ja laienemist.

Erilist tähelepanu on doktoritöö käigus pööratud põhjavee taseme muutustele, sest sellega seotud poorirõhk on nõlva stabiilsuse seisukohalt ajas kõige suurema ja kiirema varieeruvusega komponent. Selleks koguti kokku kõik olemasolevad põhjavee monitooringu andmed (51 kaevu) Pärnu lähikonnast ning nendele vaatlustele tuginedes interpoleeriti maapinnalt esimese, viirsavide all paikneva, aluspõhjalise põhjaveekihi piesomeetriline tase. Andmed pole küll ei ajaliselt kui ka ruumiliselt ühtlased, kuid võimaldasid siiski visualiseerida depressioonilehtri tekkimise, süvenemise, täitumise ning liikumise Pärnu ümbruses perioodil 1960–2009. Vabapinnalise, maapinnale kõige lähema ning viirsavide peale paikneva põhjavee taseme jälgimiseks paigaldati 2009. aasta kevadel Tammiste ja Eametsa külas lihkeohtlikke nõlvade vahetusse lähedusse 10 automaatset piesomeetrit, mis mõõdavad poorirõhke (põhjavee taset) 1-tunnise intervalliga. Käesolevas töös analüüsiti 24 kuu andmeid, poorirõhkude seost meteoroloogiliste ja hüdroloo-giliste näitajatega ning mõju nõlvade stabiilsusele.

Uurimuse käigus selgus, et poorirõhud muutuvad savinõlvades (Sauga) tunduvalt rohkem ja kiiremini kui liivanõlvades (Tammiste). Samuti on poori-rõhkude mõju nõlvade stabiilsusele tunduvalt suurem savinõlvades. Poorirõhku-dest ning piirnevate jõgede veetaseme muutuste tõttu võib keskmine aastasisene varuteguri (FoS) kõikumine savinõlvades ulatuda 0.34-ni, liivnõlvades on mõju tunduvalt väiksem (0.08). Täiendavalt on käsitletud ennustatavate kliima-muutuste mõju nõlvade stabiilsusele. Survelise põhjavee taseme tõusu korral, näiteks ennustatava sademete hulga kasvu tõttu, väheneb ennekõike savinõlvade stabiilsus. Sellise stsenaariumi korral suureneks nii lihete arv kui ka lihke-komplekside ulatus.

Parim ning odavaim tegevus nõlvade stabiliseerimiseks on jõekallaste metsastamine. Sobivad taimed on kiirelt kasvavad, niiskuselembelised puud ja põõsad nagu näiteks pajud (*Salix*) ja lepad (*Alnus*). Nende juurestik vähendaks ühelt poolt erosiooni ning samas alaneks suurenenud evapotranspiratsiooni mõjul pinnases poorirõhud.

PUBLICATIONS

CURRICULUM VITAE

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Research activity

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Honours & Awards

2006, Marko Kohv; Contest of the student scientific studies held by EGOES, first prize.

Dissertations supervised

Helen Hiiemaa, Master's Degree, 2011, (sup) Tiit Hang, Marko Kohv, Selisoo hüdroloogiline uuring (Hydrology of the Selisoo mire), University of Tartu, Faculty of Science and Technology, Institute of Ecology and Earth Sciences, Tartu University, Department of Geography

Maarja Kirt, Master's Degree, 2009, (sup) Tiit Hang, Marko Kohv, Selisoo kujunemise geoloogilised eeldused (Geological presumptions for formation of the Selisoo mire), University of Tartu

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Teaduspreemiad ja -tunnustused

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