MTT SCIENCE 17

Physical properties of a boreal clay soil under differently managed perennial vegetation

Doctoral Dissertation

Kimmo Rasa

MTT

MTT SCIENCE 17

Physical properties of a boreal clay soil under differently managed perennial vegetation

Doctoral Dissertation

Kimmo Rasa

Academic Dissertation:

To be presented, with the permission of the Faculty of Agriculture and Forestry of the University of Helsinki, for public criticism in Arppeanum, Snellmaninkatu 3, Helsinki, on November 25th, 2011, at 12 o´clock.
Finland 2011

ISBN 978-952-487-335-2 (Print)
ISBN 978-952-487-336-9 (Electronic)
ISSN 1798-1824 (Printed version)
ISSN 1798-1840 (Electronic version)
www.mtt.fi/mtttiede/pdf/mtttiede17.pdf

Distribution and sale
MTT Agrifood Research Finland,
Media and Information Services,
FI-31600 Jokioinen,
phone +358 3 41881, e-mail julkaisut@mtt.fi
Printing year 2011
Cover photo by Kimmo Rasa: Shrinkage
cracks in a clayey surface soil
Printing house Tampereen Yliopistopaino
Juvenes Print Oy

Supervisors:
Professor Markku Yli-Halla
University of Helsinki, Finland

Professor Rainer Horn
Christian-Albrechts-Universität zu Kiel,
Germany

Doctor Liisa Pietola
University of Helsinki, Finland

Pre-examiners:
Professor Trond Børresen
Norwegian University of Life Sciences,
Norway

Professor Wim Cornelis
Ghent University, Belgium

Opponent:
Professor Nicholas Jarvis
Swedish University of Agricultural
Sciences, Sweden

Custos:
Professor Markku Yli-Halla
University of Helsinki, Finland

Physical properties of a boreal clay soil under differently managed perennial vegetation

Kimmo Rasa

MTT Agrifood Research Finland, Plant Production Research, FI-31600 Jokioinen
kimmo.rasa@mtt.fi

Abstract

The physical properties of surface soil horizons, essentially pore size, shape, continuity and affinity for water, regulate water entry into the soil. These properties are prone to changes caused by natural forces and human activity. The hydraulic properties of the surface soil greatly impact the generation of surface runoff and accompanied erosion, the major concern of agricultural water protection.

The general target of this thesis was to improve our understanding of the structural and hydraulic properties of boreal clay soils. Physical properties of a clayey surface soil (0–10 cm, clay content 51%), with a micaceous/illitic mineralogy subjected to three different management practices of perennial vegetation, were studied. The study sites were vegetated buffer zones located side by side in SW Finland: 1) natural vegetation with no management, 2) harvested once a year, and 3) grazed by cattle. The soil structure, hydraulic properties, shrinkage properties and soil water repellency were determined at all sites.

Two distinct flow domains were evident. The surface soil was characterized by subangular blocky, angular blocky and platy aggregates. Hence, large, partially accommodated, irregular elongated pores dominated the macropore domain at all sites. The intra-aggregate pore system was mostly comprised of pores smaller than 30 μm, which are responsible for water storage. Macropores at the grazed site, compacted by hoof pressure, were horizontally oriented and pore connectivity was poorest, which decreased water and air flux compared with other sites.

Drying of the soil greatly altered its structure. The decrease in soil volume between wet and dry soil was 7–10%, most of which occurred in the moisture range of field conditions. Structural changes, including irreversible collapse of interaggregate pores, began at matric potentials around -6 kPa indicating, instability of soil structure against increasing hydraulic stress. Water saturation and several freeze-thaw cycles between autumn and spring likely weakened the soil structure.

Soil water repellency was observed at all sites at the time of sampling and when soil was dryer than about 40 vol.%. (matric potential < -6 kPa). Therefore, water repellency contributes to water flow over a wide moisture range. Water repellency was also observed in soils with low organic carbon content (< 2%), which suggests that this phenomenon is common in agricultural soils of Finland due to their relatively high organic carbon content.

Aggregate-related pedofeatures of dense infillings described as clay intrusions were found at all sites. The formation of these intrusions was attributed to clay disper-

sion and/or translocation during spring thaw and drying of the suspension in situ. These processes generate very new aggregates whose physical properties are most probably different from those of the bulk soil aggregates. Formation of the clay infillings suggested that prolonged wetness in autumn and spring impairs soil structure due to clay dispersion, while on the other hand it contributes to the pedogenesis of the soil.

The results emphasize the dynamic nature of the physical properties of clay soils, essentially driven by their moisture state. In a dry soil, fast preferential flow is favoured by abundant macropores including shrinkage cracks and is further enhanced by water repellency. Increase in soil moisture reduces water repellency, and swelling of accommodated pores lowers the saturated hydraulic conductivity. Moisture- and temperature-related processes significantly alter soil structure over a time span of 1 yr. Thus, the pore characteristics as well as the hydraulic properties of soil are time-dependent.

Key words:
clay soil, boreal climate, buffer zone, shrinkage, water repellency, hydraulic properties, thin sections, micromorphology

Suomalaisen savimaan fysikaaliset ominaisuudet eri tavoin hoidetuilla suojavyöhykekasvustoilla

Kimmo Rasa

MTT (Maa- ja elintarviketalouden tutkimuskeskus), Kasvintuotannon tutkimus, 31600 Jokioinen, kimmo.rasa@mtt.fi

Tiivistelmä

Maan pintakerroksen fysikaaliset ominaisuudet, kuten huokosten koko, muoto ja jatkuvuus sekä maan vedenhylkivyys, säätelevät veden imeytymistä syvempiin kerroksiin. Nämä luonnonvoiminen ja ihmistoiminnan vaikutuksesta herkästi muuttuvat maan ominaisuudet vaikuttavat pintavirtauksen ja edelleen eroosion syntyyn. Maatalouden vesiensuojelussa eroosion ehkäiseminen on yksi tärkeimmistä tavoitteista.

Tämän tutkimuksen tavoitteena oli parantaa käsitystämme boreaalisen savimaan rakenteellisista ja hydraulisista ominaisuuksista. Tutkimus tehtiin Jokioisten savimaalla (savespitoisuus 51 %) sijaitsevalla koekentällä, jonne oli perustettu eri tavoin hoidettuja monivuotisen kasvillisuuden peittämiä suojavyöhykkeitä: 1) hoitamaton luonnontilainen, 2) vuosittain niitetty ja 3) laidunnettu suojavyöhyke. Keväällä 2005 otetuista maanäytteistä (0-5 ja 5-10 cm) tutkittiin maan rakennetta, hydraulisia ominaisuuksia, kutistumisominaisuuksia ja vedenhylkivyyttä.

Pintamaan rakenne muodostui kulmikkaista, hieman pyöristyneistä tai litteistä muruista. Murujen väliin jäi suuria epäsäännöllisiä pitkänomaisia huokosia, joita pitkin vesi liikkuu. Murujen sisällä oli pääasiassa pieniä huokosia (alle 30 µm), jotka varastoivat vettä maahan. Laidunnetulla suojavyöhykkeellä maa oli tiivistynyttä: isot huokoset olivat vaakatason suuntaisia ja huokosten jatkuvuus oli heikoin, minkä seurauksena veden- ja ilmanjohtavuus oli käsittelyistä huonoin.

Savimaan kuivuminen vaikutti merkittävästi sen rakenteeseen. Maan tilavuus pieneni 7-10 %, kun kyllästynyt maa kuivattiin täysin. Valtaosa tilavuuden muutoksesta tapahtui kuitenkin jo luonnossa esiintyvissä kosteusolosuhteissa. Rakenteelliset muutokset, mm. huokosrakenteen romahtaminen, alkoivat maan ollessa vielä hyvin kosteaa (matriisipotentiaali -6 kPa), eli heikotkin hydrauliset voimat aiheuttivat pysyviä muutoksia maan rakenteeseen. Kasvukauden ulkopuolinen märkyys sekä jäätyminen ja sulaminen ovat oletettavasti heikentäneet maan rakennetta.

Maa oli näytteenottohetkellä vettä hylkivää (R>1,95) kaikilla suojavyöhykkeillä. Vedenhylkivyyttä esiintyi maan vesipitoisuuden ollessa alle 40 % maan tilavuudesta (matriisipotentiaali < -6 kPa), joten vedenhylkivyys vaikuttaa veden liikkeisiin laajalla kosteusalueella. Vedenhylkivyyttä esiintyi myös maan orgaanisen aineen pitoisuuden ollessa alhainen (<2 %). Koska suomalaisten maiden orgaanisen aineksen pitoisuus on suhteellisen korkea, on vedenhylkivyys oletettavasti yleinen ilmiö maatalousmaissamme.

Ohutleikkeiden (paksuus 30 µm) mikromorfologisessa tarkastelussa havaittiin

maamatriisista poikkeavia tiheitä saveskeskittymiä. Nämä muodostumat ovat voineet syntyä keväällä dispergoituneen ja/tai kulkeutuneen saveksen kuivuessa vielä syvemmältä routaisen maan pintakerrokseen. Prosessin seurauksena syntyy uusia maamurusia, joiden fysikaaliset ominaisuudet oletettavasti poikkeavat aiempien murujen ominaisuuksista. Murujen synty osoittaa, että kasvukauden ulkopuolinen märkyys heikentää maan rakenteen kestävyyttä (dispersio), mutta toisaalta se synnyttää uusia rakenneyksiköitä.

Väitöskirjan tulokset osoittavat savimaan fysikaalisten ominaisuuksien olevan dynaamisia. Kun maa on kuivaa, kutistumisen seurauksena syntyneet halkeamat ja vedenhylkivyys lisäävät veden oikovirtausta. Maan kostuessa vedenhylkivyys vähenee, mutta osa halkeamista sulkeutuu, mikä heikentää hydraulista johtavuutta. Vuotuiset kosteuden ja lämpötilan muutoksiin liittyvät prosessit muuttavat maan fysikaalisia ominaisuuksia merkittävästi. Siten maan huokosrakenne ja hydrauliset ominaisuudet vaihtelevat vuodenajan mukaan.

Avainsanat:
savimaa, boreaalinen ilmasto, suojavyöhyke, kutistuminen, vedenhylkivyys, hydrauliset ominaisuudet, ohutleikkeet, mikromorfologia

Acknowledgements

I began my work on soil physics in the spring of 2005, when I joined a project focusing on the functioning of buffer zones. The work was done at the University of Helsinki under the supervision of Professor Markku Yli-Halla and Docent Liisa Pietola. I extend my warmest thanks to both of you for guiding me through these years.

International co-operation had very important role in this thesis. I am grateful to my supervisor abroad, Professor Rainer Horn, who guided me in the field of soil physics. The time I spent in Kiel was hectic, productive, and interesting. I wish to thank Professor Rolf Tippkötter for introducing me to soil micromorphology and the beauty of the soils. During my several visits to Bremen, I was fortunate to work together with Dr Thilo Eickhorst, the best "young micromorphologist in 2010". Long days in the laboratory were compensated by good company and food, as well as tasty German beer.

I wish to thank the pre-examiners of my thesis, Professor Trond Børresen and Professor Wim Cornelis, for the constructive criticism on my work.

This thesis would not exist without Professor Helinä Hartikainen. Her inspiring courses changed my focus from multidisciplinary environmental protection to soil science toward the end of my MSc studies. I wish to thank all of my co-workers in the Division of Environmental Soil Science. The hilarious breaks in the famous coffee room without any windows are unforgettable. Warmest thanks to Mari Räty for fellowship and support throughout the years of our PhD studies.

In the course of my PhD-studies, I had an opportunity to work for two years for the TEHO-project in the Southwest Finland Regional Environment Center. There, I was able to combine my knowledge of soil science with practical questions concerning agricultural water protection. Thanks TEHO girls and boys. My subsequent position at MTT Agrifood Research Finland enhanced the work of this thesis. I wish to thank Helena Soinne, Riikka Keskinen, Jaana Uusi-Kämppä, Jari Hyväluoma, Risto Uusitalo, and Eila Turtola for encouragement during the final stages of this project.

For financial support, I express my sincere appreciation to the Finnish Ministry of Agriculture and Forestry, Maj and Tor Nessling Foundation, Maa- ja Vesitekniikan Tuki ry, The Finnish Cultural Foundation, Salaojituksen tukisäätiö and the University of Helsinki Funds.

Finally, I am grateful to my family for unceasing support and encouragement during the years of my academic studies. Dear Dr Outi Vesakoski, I would like to thank you for mental sparring, which made the writing of this thesis so much easier.

List of original publications

This thesis is based on the following publications:

I Rasa, K., Eickhorst, T., Tippkötter, R. & Yli-Halla, M. Structure and pore system in a differently managed clayey surface soil as described by micromorphology and image analysis. Manuscript.

II Rasa, K. & Horn, R. Structure and hydraulic properties of a boreal clay soil under differently managed buffer zones. Manuscript.

III Rasa, K., Horn, R., Räty, M., Yli-Halla, M. & Pietola, L. 2009. Shrinkage properties of differently managed clay soils in Finland. Soil Use and Management 25: 175-182.

IV Rasa, K., Horn, R., Räty, M., Yli-Halla, M. & Pietola, L. 2007. Water repellency of clay, sand and organic soils in Finland. Agricultural and Food Science 16: 267-277.

Docent Liisa Pietola, Professor Markku Yli-Halla and Kimmo Rasa originally planned the experimental design. Professor Rainer Horn planned the laboratory work related to soil shrinkage (III) and water repellency (IV) and Kimmo Rasa refined the plans. Kimmo Rasa did the fieldwork, as well as most of the laboratory work. Measurements of hydraulic conductivity, bulk density and macroporosity (II) were done by research technician M.Sc. Olga Nikolenko. Dr. Thilo Eickhorst guided the laboratory work related to soil thin sections (I). The qualitative description of the thin sections was prepared together with Dr. Thilo Eickhors and Professor Rolf Tippkötter (I). Kimmo Rasa calculated and interpreted the results and prepared the manuscripts. The co authors commented on the manuscripts.

Reprints of the original articles are published with the kind permission of John Wiley & Sons Ltd. (III) and The Scientific Agricultural Society of Finland and MTT Agrifood Recearch Finland (IV).

Abbreviations

BZ	Buffer zone
COLE	Coefficient of linear extensibility
EPD	Equivalent pore diameter
K_a	Air permeability
K_{fs}	Field-saturated hydraulic conductivity
K_{sat}	Satureated hydraulic conductivity
WR	Water repellency

Contents

1 Introduction

Soil structure, essentially the size, shape and continuity of the pores, is a fundamental soil property. It controls the transport of water, gases, heat and solutes into the soil. Soil structure affects root growth and soil workability, and it defines the habitat of soil fauna. In agricultural soils, the interaction between soil structural properties and water contributes significantly to soil productivity, as well as to the environmental impacts of agriculture.

The soils of Finland are subjected to a boreal climate. These soils experience dry spells in summer. After the growing season, they are exposed to autumn rains and prolonged wetness, freezing in winter and finally snowmelt waters and melting of frost in spring. These conditions leave a substantial imprint on our soils. For example, in clay soils freezing, thawing and seasonal variation in soil moisture rework pores and aggregates (Heinonen & Pukkala, 1954; Kivisaari, 1979). Soil cracking is a well-known phenomenon in practice and often visible to the naked eye. However, some other soil moisture-related phenomena, such as water repellency (WR), may play important, though less evident, roles in the interaction of soil and water. WR reduces soil affinity for water and therefore affects the infiltration processes. These phenomena make soil structure and hydraulic properties highly dynamic by nature.

The seasonal variation in the distribution of water into surface runoff and subsurface drainage flow has been recognized in long-term field experiments in Finland (Puustinen et al., 2005; Turtola et al., 2007; Uusi-Kämppä & Jauhiainen, 2010). Without doubt, climatic factors are of great importance, but soil moisture related behaviour strongly affects water distribution as well. The focus has not been strong on shrinkage properties and WR, although their importance in soil function has been commonly addressed internationally (e.g. Doerr et al., 2007; Jarvis, 2007; Köhne et al., 2009). In this thesis, soil structure, shrinkage characteristics, WR and hydraulic properties are examined in the context of Finnish clay soil in boreal climate.

1.1 Clay soils of Finland

At the end of the Weichselian glaciation 11 500 years ago, glacial and postglacial clays were deposited. Currently, these clayey soils (> 30% of clay) are being used effectively in agriculture especially in southern and southwestern Finland, where their proportion exceeds 50% (Frosterus, 1922). The percentage of fine mineral (clay and silt) soils is close to 70% in the most southern crop cultivation zone of Finland (Sippola & Tares, 1978). Clayey soils contribute to around one third to one fourth of the total agricultural land area of 2.3×10^6 ha in Finland (Viljavuuspalvelu, 2008). Although the clay content in the plough layer of these soils is often between 30% and 60%, it may increase to 80% or even over 90% in the deeper horizons (e.g. Turtola & Paajanen, 1995; Yli-Halla & Mokma 2001).

The common pedon properties of clay soils in Finland can be found in the studies of Yli-Halla and Mokma (1998, 2001), Peltovuori et al. (2002) and Yli-Halla et al. (2009). These studies showed that noncalcareous clay soils are often formed of more than one parent material, e.g. having thin strata of silt incorporated into clay horizons. These young soils are cold, having a cryic soil temperature regime. The mean summer soil temperature measured in Anjalankoski (1971–1990), southern Finland at a depth of 50 cm, was 13.7 °C and mean annual soil temperature 6.2 °C. Weathering has been only slight in the uppermost part of the soil profile and is even less significant in the deeper horizons. Artificial drainage has influenced the pedogenesis in agricultural fields, allowing development of the soil structure at least to the depth of the drainage pipes (app. 1 m). The structure of the surface horizon is often described as sub-angular to angular blocky, changing to prismatic in the deeper horizons.

Turtola et al. (2007) reported some weather data collected from Jokioinen, southwestern Finland. The annual precipitation between 1989 and 2001 was in the range of 423–777 mm (mean 643 mm). In this same period, the average duration of the snow cover was from 24 November to 14 April and dates of frost from 17 November to 21 April. The maximum frost depth was 39 cm and the number of total snowmelts before the final snowmelt varied from zero to two. In Jokioinen, the daily mean soil temperatures may reach 20°C at a depth of 20 cm (Heikinheimo & Fougstedt, 1992), which suggests much higher temperatures in the daytime in the topmost part of the soil. The topsoil experiences several wetting-drying cycles annually caused by frost in winter and drought in summer.

Installation of subsurface drainage pipes increased extensively in the 1960s and the majority of cultivated clay soils has since been artificially drained. Despite extensive artificial drainage, these soils tend to have low hydraulic conductivity in the subsoil, resulting in fluctuation of the oxidation state. Thus, they have abundant redox concentrations and an aquic soil moisture regime, as well as often a stagnic soil colour pattern (FAO, 2006). According to Soil Taxonomy they are classified as Cryaquepts (Yli-Halla & Mokma, 2001). Since the soil is wet several months annually, soil compaction is a major threat to soil quality during agricultural operations and grazing (Alakukku, 1996; Pietola et al., 2005; Räty et al., 2010a).

Fluctuating climate and frost in winter are characteristics of boreal soils. In an international context, clayey soils subjected to such conditions occur, e.g. in Scandinavia and Canada. In the agricultural soil of central and eastern Sweden and southeastern Norway, heavy clays and silty clays are dominant (Ulén et al., 2007). As in Finland, high clay contents of up to 60% also occur in Sweden (Rasmussen, 1999). Mineralogy rich in illites characterizes glacial and postglacial clays both in Sweden and in Finland (Sippola, 1974; Messing & Jarvis, 1990; Ulén & Snäll, 2007) and clays containing mainly illite occur in Norway as well (Olsen & Haugen, 1998). In Canada, soils with high clay content (up to 80%) having micaceous mineralogy (Ross, 1978) are present under relatively similar climatic conditions as in Fennoscandia. In contrast, smectite, micaceous and mixed mineralogy classes are dominant in British clayey soils (Reeve et al., 1980)

Soils rich in clay play a dual role. On one hand, these soils are important for food production, but on the other they contribute significantly to diffuse nutrient loading to surface waters and further to the Baltic Sea. The agricultural history of southwestern Finland, the major area of clay soils, dates back several hundred years. During this time, the majority of clay soils has been reclaimed for agriculture. Soil erosion is a natural process in these soils. However, intensification of cultivation, heavier machines and lack of wintertime vegetation cover are conducive to increased erosion (Puustinen et al., 2005), because these agricultural practices deteriorate soil structure and expose the soil surface to raindrop impacts. At the same time, the amount of nutrients potentially transported by eroded soil particles has increased, due to intensive phosphorus fertilization and build-up of phosphorus in the plough layer, particularly from the 1960s to the 1990s (Uusitalo et al., 2007).

The agri-environment support scheme used in Finland includes several measures to reduce nutrient discharge, from which establishment of vegetated buffer zones (BZs) is of interest in the context of the present study. BZs are established on agricultural soil between arable fields and bodies of water to reduce erosion and runoff of particle-bound nutrients (Syversen, 2005; Hoffmann et al., 2009; Uusi-Kämppä & Jauhiainen, 2010). Although the function of a BZ is based on several processes described by Hoffmann et al. (2009), the structure of the surface soil is fundamental because it contributes to water infiltration. Fast and even water infiltration of well-structured soil reduces surface runoff and accompanied soil erosion, decreases water bonding and clay dispersion, and allows interaction between water, solutes and soil matrix (sorption processes) as water flows through the soil profile. Therefore, soil structure conducive to water infiltration is considered as a prerequisite for a well-functioning BZ (Hoffmann et al., 2009).

1.2 Soil structure and water flow

Soil is traditionally described as a three-phase system, in which the solid phase is formed of organic and inorganic material, while liquid or gaseous compounds fill the pore space between the solids. Arrangement of the solids determines the quality and quantity of soil porosity. In soil physics, this complex dynamic three-dimensional architectonic system is termed the soil structure. In fine-textured soils, flocculation of clay particles forms coalitions of different sizes, i.e. soil aggregates. In the agricultural soils of Finland, aggregation is enhanced by relatively high clay content, abundant cementing by organic matter and biological activity (Heinonen, 1951). In addition, changes in moisture are important in the generation of soil structure (Keso, 1930).

Water transport is dependent on the properties of the entire soil profile as well as the drainage system (ditches, subsurface drains). Conductive macropores are essential to prevent water bonding in the surface horizon and generation of surface runoff. The contribution of macropores to soil hydraulic processes has gained increased focus in the scientific literature (e.g. Aura, 1983; Youngs & Leeds-Harrison, 1990; Øygarden et al., 1997; Jarvis, 2007; Hintikka et al., 2008; Alakukku et al., 2010). Even though macropores allow fast infiltration, they may also facilitate transport of agrochemicals, including inorganic and organic fertilizers and particle-bound nutrients to subsurface drains (Øygarden et al., 1997; Uusitalo et al., 2001; Jarvis, 2007). These preferential flow routes act as main pathways of water to the subsurface drainage pipes (Hintikka et al., 2008; Warsta, 2008). Pores, especially those larger than 300 μm in equivalent diameter, are considered to participate in these processes (Jarvis, 2007). In addition to pore-size distribution, a crucial factor determining the ability of the pore system to conduct solutions is the pore geometry. Rounded earthworm burrows (channels) and continuous elongated pores (planes) are conducive to water flow (Pagliai et al., 2004).

The structure of surface soil regulates water entry into the soil. A permeable surface soil allows infiltration and acts as a distribution zone, where water seeks its way to vertical macropores. In the agricultural soils of Finland, these horizons are commonly characterized by different degrees of aggregation (Heinonen, 1951). The aggregates participate in soil water-flow, since they start to imbibe water in the early stages of rain events falling on dry soil. Water is further distributed into deeper soil layers via contact points between the aggregates. When the water potential in the soil surface increases (soil becomes wet) or the infiltration capacity of the aggregates is exceeded, water enters the macropores (Bouma & Dekker, 1978). The contribution of macropores to flow processes increases hydraulic conductivity significantly. Youngs and Leeds-Harrison (1990) described these principles of water movement in aggregated soils, addressing the importance of the bimodal nature of the pore system of the aggregated soils and the degree of saturation in water flow.

The surface horizon is the most active part of the soil, in terms of both biological and physical factors. Cultivated soils are annually loosened to maintain a structure favourable for water infiltration, gas exchange and root growth. Artificial loosening of the structure is restricted in soils under permanent vegetation, including BZs and pastures, or no-till fields. In these soils, biological activity, such as burrowing animals and plant roots, is an important factor forming soil structure. The structure of these surface soils is prone to artificial stresses, such as tractor wheeling and cattle trampling. Moreover, the abiotic stresses, including shrink-swell and freeze-thaw cycles, are most intense and frequent in the topsoil. Therefore, the structural properties of surface horizons are of special interest when the hydraulic properties and environmental effects of soils without annual till are considered.

1.2.1 Study of soil structure

Field observation of soil properties is the first step in the investigation of soil structure. Further studies are often based on the analysis of undisturbed core samples. Commonly measured indirect attributes, such as bulk density, total and macroporosity and saturated hydraulic conductivity (K_{sat}), are affected by soil structure. Such studies have shown the importance of macropores on hydraulic properties and crop growth in clay soils of Finland (Aura, 1983; Alakukku, 1996; Pietola et al., 2005). However, the need for more fundamental understanding of soil structure and porosity has been emphasized in to better interpret the results gained from indirect measurements (Ringrose-Voase, 1991; Alakukku, 1998; Bouma, 2006). Methods applicable to yielding more specific data on pore characteristics include image analysis of soil thin sections, computed tomography, magnetic resonance images (Kodešová, 2009) and electrical resistivity tomography (Samouëlian et al., 2004).

Soil micromorphology is a method for characterizing solids, pores and many other soil features observed in thin sections. The terminology used to describe these features was developed in the early 20th century and has been updated and reformulated since then (Kubiena, 1938; Brewer, 1976; Fitzpatrik, 1984; Stoops, 2003). To study the soil structure, micromorphology provides a method for classifying pores based on their shape and size appearing in two dimensions. However, the technique is limited to pores larger than 30–50 µm in equivalent diameter.

The origin of the main pore types can be related to specific soil processes. For example, biological activity, plant roots and burrowing animals generate rounded pores (channels), whereas planar flat voids (planes) are often associated with soil shrinkage. Soil compaction results in massive structures or generates elongated pores (platy aggregates) oriented rectangularly to the stress direction applied (Bullock & Murphy, 1980). On the other hand, recognition of pore types allows assessment of their contribution to water movement in soil. Continuous elongated planes and cylindrical channels are especially conducive to water flow (Pagliai et al., 2004).

Even though the terminology describing soil features has been developed for scientific purposes to maintain unambiguousness and comparability of the observations, science has an inherent tendency towards quantification. In the 1970s, technological development allowed quantitative detection of soil porosity with computer-aided digital image analysis of soil thin sections (e.g. Murphy et al., 1977a). In this approach, thin sections are photographed and measurements of void properties, including area, number, perimeter (Pe) and convex perimeter (Pe_{co}) are done with image analysis software available commercially. From the basic measurements, parameters related to pore characteristics (e.g. shape, irregularity) can be derived (Murphy et al., 1977a; Pagliai et al., 1983; Ringrose-Voase, 1991).

Combination of qualitative and quantitative methods for studying thin sections has proven to be an efficient approach for studying differences in soil structure under various management practices, treatments or natural conditions (Murphy et al., 1977b; Ringrose-Voase, 1991). For example, the effects of zero and conventional tillage (Pagliai et al., 1984; VandenBygaart et al., 1999), influence of earthworms (VandenBygaart et al., 2000) and application of organic matter (Pagliai & Vittori Antisari, 1993) on soil structure have been studied. Sveistrup et al. (2005) studied the morphological and physical properties of Norwegian silt/silt loam soils. They addressed the importance of freezing and desiccation on structure formation of these soils. Yli-Halla et al. (2009) presented a detailed description of a clay soil of Finland, including micromorphology of vertically oriented thin sections. The study focused on pedogenesis, and therefore the emphasis was on subsoil horizons. In a context of clay soils of Finland, qualitative and quantitative characterization of the pore morphology is still lacking.

1.2.2 Shrinkage

Soil shrinkage upon drying is a fundamental process in the formation of soil structure and therefore it affects numerous processes occurring in soil. In the international literature, shrinkage properties were commonly discussed in the early 20th century (Tempany, 1917; Haines, 1923). The importance of soil shrinkage on drainage practices was recognized in Finland as well (Keso, 1930). The development and improvement of models and mathematical procedures for describing more precisely this phenomenon is often justified by the importance of shrinkage in water and solute transport. Hydrological modelling especially benefits from better understanding of shrinkage-related processes (Kankaanranta, 1996; Baumgartl & Köck, 2004; Peng et al., 2006; Warsta, 2008).

Shrinkage properties are affected by many soil properties, such as mineralogy, chemical composition, structural state, bulk density, degree of aggregation, aggregate strength, predrying intensity, as well as soil biota (Reeve et al., 1980; Peng & Horn, 2005; Peng et al., 2007). Fine clay material (< 0.2 μm) in clay soils of Finland is dominated by micaceous mineralogy or illites (Sippola, 1974). The coefficient of linear extensibility (COLE) of these soils is high enough to justify use of the Vertic qualifier (FAO, 2006; Yli-Halla et al., 2009).

However, the COLE provides information on the total potential of soil shrinkage. To assess the dynamic process of soil shrinkage, decrease in soil volume upon drying is traditionally described as a shrinkage curve (Fig. 1). The development of shrinkage models and mathematic methods allow a quantitative definition of shrinkage zones and their contribution to the loss of water and volume. More specifically, the water pools active in different shrinkage zones, i.e. condensed and swelling water in the micro- and macropore systems, can be defined (Braudeau et al., 2004). Soil physical properties are reflected in the shape of the shrinkage curve as well (Groenevelt & Grant, 2001; Cornelis et al., 2006b).

The shrinkage curve can be presented in terms of the soil moisture ratio (ϑ) and void ratio (e) (Bronswijk, 1991; Peng & Horn, 2005).

$\vartheta = V_w V_s^{-1}$ Equation 1

$e = V_f V_s^{-1}$ Equation 2

where V_w, V_s and V_f are volumes of water, solids and pores, respectively. The shrinkage curve consists of four distinct shrinkage zones. These zones, starting from saturated soil and ending in dry soil (Fig. 1), are as follows: structural shrinkage zone (i) is associated with macropores (i.e. voids between structural units, peds and aggregates, and biopores), from which water is removed and replaced by air without significant volume loss. In the proportional shrinkage zone (ii), volume loss and water loss are thought to be equal (the 1:1 line), or at least proportional to each other, forming a straight line. In the residual shrinkage zone (iii) air enters the finer pores of the soil structural units and the volume loss is smaller than the loss of water. The dry end of the shrinkage curve is the zero shrinkage zone (iv). The volume loss in this zone is negligible, since the rest of the water between the primary particles is removed. These zones can be distinguished by mathematical procedures (Groenevelt & Grant, 2001; Peng & Horn, 2005). The fitting parameters of the soil shrinkage curve seem to have a physical meaning as well, but this approach is not yet well established (Peng & Horn, 2005).

The contribution of shrinkage to the formation of soil structure was described, e.g., by Horn and Smucker (2005) and is referred to here only briefly. When a homogenized soil dries, removal of water from the pore system results in meniscus forces. These forces pull adjacent soil particles closer to each other. The consequent decrease in soil volume results in rearrangement of soil particles, generation of shrinkage cracks and soil aggregation. Increase in contact points between the construction units of the aggregates makes them denser and strengthens them. Consequently, shrinkage contributes to development of inter- and intra-aggregate pore systems.

Shrinkage greatly influences water movement in soil. Shrinkage cracks provide routes for preferential flow when open in a dry soil. At the soil surface, these cracks may participate in distribution flow, since they provide a highly interconnected pore system between cracks and biopores (Shipitalo et al., 2004; Jarvis, 2007). Shrinkage cracks affect soil temperature. In windy weather the relative humidity drops, heat flux becomes faster and evaporation increases in the shrinkage cracks (Adams & Hanks, 1964). Drying enhances cracking and formation of the soil structure of deeper soil horizons otherwise having low hydraulic conductivity.

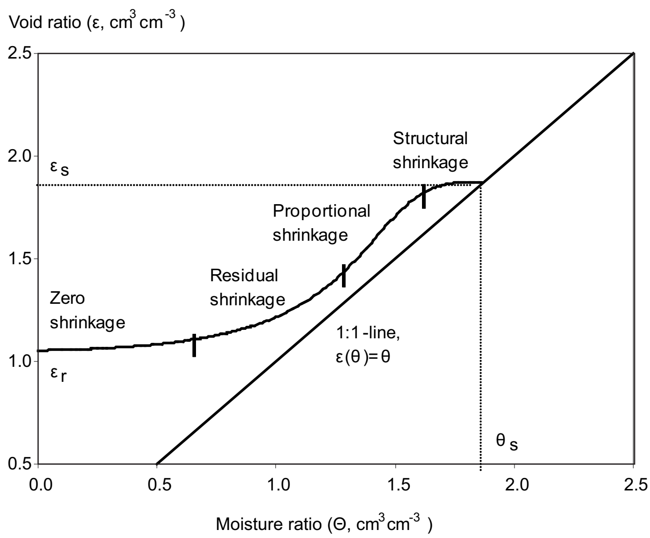

Figure 1. Schematic presentation of the shrinkage curve and four different shrinkage zones according to Peng and Horn (2005). The subscript r stands for residual (dried at +105 °C) and s stands for water saturation. On the 1:1 line water and volume losses are identical.

Yli-Halla and Mokma (2001) documented 1–cm-wide cracks between prisms below the plough layer and angular blocky structure down to a depth of 112 cm in a clay soil of Finland. Considering the importance of shrinkage in soil properties and processes, as well as agricultural practices, this phenomenon in clay soils of Finland is poorly documented and little is known of how soil management affects shrinkage characteristics.

1.2.3 Water repellency

Water entry into the soil occurs via inter-aggregate and intra-aggregate pores. The hydraulic properties of the aggregates are defined by their pore structure, but the surface properties may significantly alter the affinity of aggregates for water in the form of soil WR, i.e. hydrophobicity (Fig. 2). A soil is critically water-repellent when water does not enter the soil (contact angle of soil and water > 90°). Tillman et al. (1989) called a soil subcritically water-repellent when wetting was not fully restricted but delayed because of hydrophobicity.

WR, factors affecting its occurrence and implications of WR on water flow are extensively studied topics. For example, an overview of studies on WR written by DeBano (2000) was based on 500 papers published since the 19th century. In 2005, Dekker et al. published a paper entitled Exponential increase of publications related to soil water repellency. Indeed, at that time more than 1200 papers related to WR were published. These papers were produced by scientists of several backgrounds, including hydrologists, soil scientists, biologists, chemists and physicists, to name a few of those listed by Doerr et al. (2007). In this paper, the authors addressed the fact that the work is not yet done, but that there remain several open questions regarding, e.g., the effect of temporally variable environmental conditions on WR. Regarding the number of papers written on WR, it may be appropriate to suggest the reader go through these review papers to gain a comprehensive insight into the phenomenon at issue. Here, a few important aspects of soil WR regarding boreal climates and clayey soils were selected

Figure 2. Water droplet on hydrophobic soil.

Soil WR is commonly related to soil organic matter, although the existence and severity of WR cannot be unambiguously explained by the amount or chemical composition of organic matter (Ma'shum & Farmer, 1985; Franco et al., 1995; Doerr et al., 2000; Harper et al., 2000). Microbiota may either decompose or generate organic substances involved in WR (Hallett & Young 1999). Land use affects WR, often being high under permanent vegetation, whereas cultivation reduces the degree of WR (Dekker & Ritsema, 1996; Hallett et al. 2001). Traditionally, WR is considered to be common in coarse soils and warm climates (Dekker, 1998; Ritsema, 1998; Doerr & Thomas, 2000), although it was later recognized in soils of humid climates and clayey soils as well (Jaramillo et al., 2000; Pietola et al., 2005; Doerr et al., 2006; Jarvis et al., 2008). In clayey soils, WR is attributed to aggregates covered by hydrophobic substances. The degree of WR is moisture-dependent, being usually higher in dry soils. Therefore, It is practical to define the critical soil water content limit, which divides soils into water-repellent and wettable (Dekker & Ritsema 1994, Dekker et al., 2001; Doerr et al., 2006).

The interest in WR stems from its importance in numerous soil processes, including plant growth, hydrology and erosion (Doerr et al., 2000). WR delays wetting of the soil matrix, enhancing preferential flow, and results in an uneven wetting pattern (Dekker & Ritsema 1996). WR may also expose soil to surface runoff and consequent erosion (Osborn et al., 1964). However, predicting the influence of WR on water and solute transport is complicated and no appropriate method has been introduced (Shakesby et al., 2000; Jarvis et al., 2008). Doerr et al. (2007) addressed the importance of quantifying and modelling the hydrological effects of WR on plot, field and catchment scales, to assess the contribution of WR to the generation of surface runoff. Despite the fact that WR is recognized as a widespread phenomenon, only one previous study suggested occurrence of WR in Finland (Pietola et al., 2005). Therefore, more comprehensive documentation of the occurrence and behaviour of WR in agricultural soils of Finland is needed.

1.3 Objectives of the study

The general target of this study was to improve our understanding of the structural and hydraulic properties of boreal clay soils. The focus was set on the surface horizon, because it is prone to natural and artificial stresses affecting soil physical properties. The structural properties of clayey surface soil subjected to different management practices of perennial vegetation were examined, using traditional indirect methods, such as determination of bulk density and macroporosity. To interpret these results more precisely and consider pore functions, pore characteristics, e.g. shape and size, were defined. Therefore, soil thin sections were described qualitatively and studied quantitatively, using image analysis. Soil hydraulic properties were studied to allow conclusions of the implications of the above-mentioned soil properties on air and water flux. It was assumed that soil subjected to cattle trampling is less favourable for infiltration than undisturbed soil and that these differences are reflected in both the quantity and quality of the pores. These results will improve our understanding of soil physical properties, especially pore morphology, that affect the water flow in boreal clay soils.

The seasonal dynamic behaviour of boreal clay soils is commonly recognized at a practical level, but the moisture-related phenomena are scarcely quantified. The other main objective of this thesis was therefore the study of two moisture-dependent properties, namely soil shrinkage and WR. The aim was to determine the occurrence and degree of WR in a clayey surface soil under differently managed perennial vegetation and to assess the effect of soil moisture content on WR. Soil shrinkage curves were determined to assess the degree of shrinkage in surface soil rich in illites, as well as to relate the shrinkage characteristics to soil physical properties. The results benefit soil hydraulic modellers, since understanding of moisture-related phenomena is considered as prerequisite for modelling of transport processes occurring in clay soils.

The aims of the study were:

- To characterize the soil macropore system of a boreal clayey surface soil qualitatively and quantitatively, and to assess the effect of different management practices on the soil pore system.
- To examine soil hydraulic properties and relate pore characteristics to these observations.
- To determine the degree of total soil shrinkage and assess the implications of soil physical properties for shrinkage characteristics.
- To determine whether WR occurs in boreal clayey surface soil and to what extent, assess the effects of soil moisture content on the degree of WR and define the moisture regime in which WR may contribute to water flow.

2 Material and methods

2.1 Study area and sampling

The Lintupaju experimental field at Jokioinen, southwestern Finland (60° 48' N, 23° 28' E), is located on the premises of MTT Agrifood Research Finland. In 1991, an experiment on vegetated BZs was initiated in the field, described in further detail by Uusi-Kämppä (2005) and Uusi-Kämppä and Jauhiainen (2010). The 18-m-wide and 10-meter-long BZs were established at the edge of the field, sloping (12–18%) to a natural gully. The cropped field continued to be used, mostly for growing cereals and silage grass in rotation. Part of the slope was cropped until 2002, when the youngest BZs were established.

The texture of the soil (0–6 cm) was determined by a pipette method (clay 51%, silt 42% and sand 7%) and the cation exchange capacity (CEC) was determined with 1 M ammonium acetate (pH 7) extraction (37 cmol (+) kg^{-1}). The silty clay soil was classified as a Vertic Stagnic Cambisol (Eutric) (FAO, 2006) or as a Typic Cryaquept (Soil Survey Staff, 2010). Soil pH was measured in water suspension (soil:H_2O = 1:2.5 volume/volume). Soil organic carbon (C%) was determined by dry combustion with the Leco CNS 1000 apparatus and soil moisture at the time of sampling was derived from the core samples (Table 1). The clay mineralogy determined for similar soils less than 1 km away was dominated by mica, vermiculite, chlorite and mixed-layer clay minerals, with feldspars and quartz as accessory minerals (Peltovuori et al., 2002; Yli-Halla et al., 2009).

Three adjacent experimental areas differing in management of perennial vegetation were sampled (names given in parentheses refer to the names of the site used in the original publications):

- natural: 14-yr-old natural vegetation with grass species, thin stand of shrubs and some hardwood trees in a natural state (I and II = natural, III = old natural, IV = 14-yr natural)
- harvested: 14-yr-old annually harvested vegetation (light lawn mower) with grass species (I and II = harvested, III = old harvested, IV= 14-yr harvested)
- grazed: 3-yr-old vegetations with grass species grazed by cattle (I and II = grazed, III = young grazed, IV = 3-yr grazed) The grazing intensity was 72 and 234 cow grazing days ha^{-1} yr^{-1} in 2003 and 2004, respectively. Before 2003, this area had been cultivated.

The management practices (harvesting and grazing) are commonly used to remove vegetation from BZs. The site harvested with the light lawn mower is considered to represent soil not affected by tractor tyres, i.e. low degree of disturbance. Grass grew at all sites, mainly timothy (*Phleum pratense L.)* and meadow fescue *(Festuca pratensis Huds.),* and the nonmanaged natural site had wild hays and flowers in addition (Räty et al. 2010a; Räty et al. 2010b).

Sampling was performed in May 2005, when the soil was drying after the snowmelt period, but before substantial vegetative growth. The field-saturated hydraulic conductivity (K_{fs}) was measured at the time of sampling and additionally in November 2005 after a long wet period, but before freezing of the soil.

Undisturbed 100–cm^3 core samples were taken (pushed by hand or gently hammered) from each site at 0–5–cm and 5–10–cm depths. At the same time, another set of 250–cm^3 core samples were collected (using an open ring holder, Eijkelkamp) from the same depths. For the preparation of soil thin sections, three replicates of undisturbed soil samples about 10 cm × 10 cm were collected. The samples were excavated carefully, using a knife or spatula, the orientation was recorded and the samples were sealed in aluminium foil and plastic to minimize changes in the soil structure due to evaporation and consequent shrinkage. The samples were stored at +4 °C until measurements. All the samples were taken and field measurements carried out along the sampling line of 2 m from the upper end of the BZ.

The 100–cm^3 and 250–cm^3 core samples were saturated with water in the laboratory. Thereafter, the 100–cm^3 samples were dried stepwise on a sand bed at matric potentials of -3, -6, -10 kPa, on a ceramic plate at -15, -30 and -50 kPa, and thereafter, at temperature of +40, +70 and +105 °C. The 250–cm^3 samples were dried at matric potentials of -1 and -10 kPa, and thereafter the temperature of +105 °C.

2.2 Laboratory and field measurements

2.2.1 Thin sections

In total, nine vertical (soil surface to about 5–cm depth) and nine horizontal (depth of 2.5 cm) thin sections (25–30 µm) were produced from the samples taken from the natural, harvested and grazed sites. Thin-section preparation is described in further detail elsewhere (I). Water was removed from the soil by wet dehydration, applying a graded series of acetone (Tippkötter et al., 1986). Unsaturated polyester resin with acetone as a thinner was used for embedding these samples (Tippkötter & Ritz, 1996). The micrographs were observed using bright-field, dark-field, and crossed-polarized light illumination microscopy, and the micromorphological descriptions were made according to Bullock et al. (1985), FitzPatrick (1984) and Stoops (2003).

For the image analysis, the binary images were generated (Photoshop CS, Adobe) from pictures taken by a digital camera (Canon EOS 350D; 8 megapixels). The images were analysed with AnalySIS 3.2 (Soft Imaging) software for pore area (A), Pe and Pe_{co} for pores larger than 50-µm equivalent pore diameter (EPD).

The pores were classified according to their EPD into five size classes (50-100 µm, 100–300 µm, 300–500 µm, 500–1000 µm, >1000 µm). The form factor (F) was calculated with the equation $F = A\ Pe^{-2}$. According to Bouma et al. (1977), elongated pores have a form factor < 0.015, irregular pores 0.015–0.04 and rounded pores > 0.04. Furthermore, the elongated pores were classified into four shape classes (Fig. 3) according to the Pe_{co}:Pe ratio as follows: < 0.3 very irregular pores, 0.3–0.5 moderately irregular pores, 0.5–0.7 moderately regular pores and > 0.7 regular pores (Pagliai et al., 1984).

Table1. Volumetric water content (W) at the time of sampling, content of soil organic carbon (C) and pH of the experimental sites.

	W(v/v), %		C, %		pH	
Site	0–5 cm	5–10 cm	0–5 cm	5–10 cm	0–5 cm	5–10 cm
Natural	42	39	5.1	2.7	5.9	5.9
Harvested	36	36	3.8	2.3	5.6	5.7
Grazed	29	34	2.5	2.1	5.7	5.7
SE	1	1	0.3	0.4	-	-

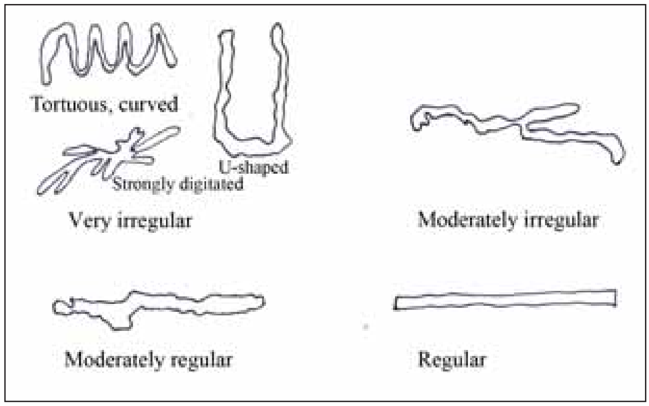

Figure 3. Schematic presentation of the shape classes of elongated pores, according to Pagliai et al. (1984).

2.2.2 Core samples

The K_{sat} of 250–cm^3 core samples (n = 10) were measured with the constant head method (Klute and Dirksen., 1986), using a laboratory permeameter (Eijkelkamp). The results were calculated according to Darcy's equation (3):

$$K_{sat} = V \times L / (A \times t \times (H_2 - H_1)) \quad \text{Equation 3}$$

in which V is the volume of water passing through the sample per unit time (t), L the length of the sample and H_2-H_1 the hydraulic head difference imposed across the sample.

The dry bulk density was calculated, dividing the weight of the soil dried at 105 °C (n = 10) by the volume of the cylinder (250 cm^3). The amount of water removed at matric potentials of -10 kPa and -1 kPa was used to determine pores > 30 µm (macropores) and > 300 µm (large macropores), respectively.

Air permeability (K_a) measurements were carried out at various moisture contents (corresponding to -3, -6, -10, -15, -30, and -50 kPa), using 100–cm^3 core samples with six replicates. The volume of air (V_a) passing through the sample per unit time (t) was determined, using an apparatus described by Horn et al. (2004) and air conductivity (K_l) was calculated, using Equation (4):

$$K_l = \rho_a \times g \times (V_a \times l)/(t \times \Delta_p \times A) \quad \text{Equation 4}$$

where ρ_a is air density at 20 °C, g the gravitational acceleration, Δ_p the air pressure applied during measurement (1 cm) and l and A are the length and area of the sample, respectively. The K_a was calculated from K_l, using Equation 5:

$$K_a = K_l \times (\eta_a / (\rho_a \times g)) \quad \text{Equation 5}$$

where η_a is the viscosity of the air. The organization index (O) was calculated, dividing K_a by the air-filled porosity at a matric potential of -3 kPa (Groenevelt et al., 1984; Blackwell et al., 1990) according to Equation 6:

$$O = K_a / \varepsilon \quad \text{Equation 6}$$

The small-scale water sorptivity (S_w), i.e. capacity of the soil to take up water by capillarity (Philip, 1957), was measured at various soil moisture contents, which correspond to the values for the K_a measurement. The infiltration of water was measured, using a mini-infiltrometer described in Chapter 2.2.4 (n = 6). The device samples a small area of the soil surface, and therefore the results are considered to represent the sorptivity of the aggregates.

2.2.3 Shrinkage

Soil height change was measured with a vernier calliper, using the 100–cm^3 core samples (n = 3) at nine defined locations after each drying step (see Chapter 2.1). The moisture ratio and void ratio were calculated according to Equations 1 and 2 (average of three replicates), and soil shrinkage was assumed to be isotropic (Bronswijk, 1990). The shape of the shrinkage curve is opposite to that defined by the water retention curve of van Genuchten's formula (Equation 7), as suggested by Peng and Horn (2005):

$$e(\vartheta) = 1 / (1 + (\acute{\alpha}\vartheta)^{-n})^{m} \qquad \text{Equation 7}$$

where ά, n and m are the fitted parameters. The soil moisture ratio and void ratio have values between those of saturated soil (ϑ_s, e_s) and soil dried at +105 °C (ϑ = 0 $cm^3\ cm^{-3}$, e_r). At saturation, the void and moisture ratios are equal and the curve ends in the residual void ratio (e_r, the dry end ϑ = 0). A schematic illustration of a shrinkage curve is presented in Figure 1. The shrinkage zones were determined using the mathematical approach (Peng and Horn, 2005) by which the inflection point, dry- and-wet side extreme curvatures were obtained by repeated differentiation of Equation 7 and then finding its roots (Mathcad software). The slope of the proportional shrinkage zone was determined, using the slope between the two endpoints of this shrinkage zone although this may have not been a straight line (Peng et al., 2005).

2.2.4 Water repellency

The infiltration of water and ethanol (95%) at 20 °C was measured, using the apparatus (Fig. 4) described by Leeds-Harrison et al. (1994) and modified by Hallet and Young (1999). The measurements were performed at various moisture contents (see Chapter 2.1), using the 100–cm^3 core samples with six replicates. Infiltration was detected manually within 75 s at 15-s intervals. The sorptivity (S) of water (S_w) and Ethanol (S_e) were calculated ac-

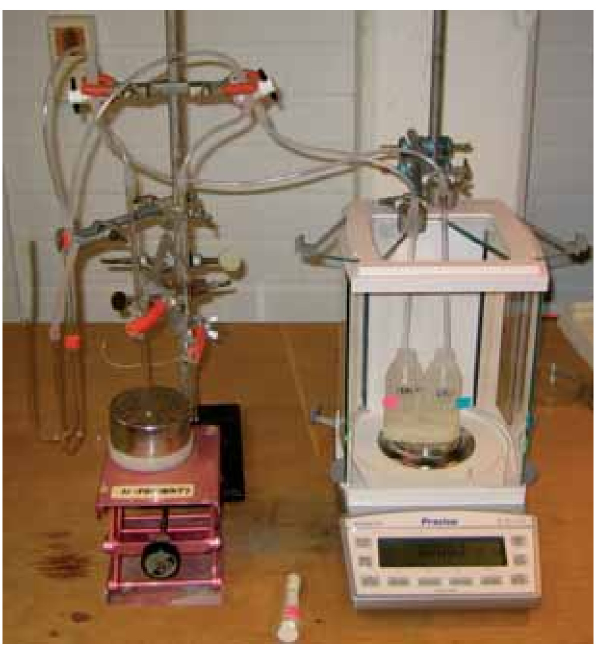

Figure 4. Mini-infiltrometer.

cording to Leeds-Harrison et al. (1994), as presented in Equation 8:

$$S = \sqrt{Q \times f \, / \, (4b \times r)} \qquad \text{Equation 8}$$

where Q is the stationary infiltration rate of liquid flow, f is the air-filled porosity and r is the radius of the infiltration tip: 1.5 mm. The value for parameter b, which is dependent on the soil-water diffusion function, is 0.55 (White & Sully, 1987). A pressure head of -2 cm was used. The WR index (R) was calculated from S_w and S_e according to Equation 9 (Tillman et al., 1989):

$$R = 1.95 \times (S_e / S_w) \qquad \text{Equation 9}$$

In Equation 9, the parameter 1.95 originates from differences of viscosity and surface tension of liquids (Hallett & Young 1999). Tillman et al. (1989) showed that an R index > 1.95 represents the transition from non-repellent to subcritically water-repellent soil. For example, an R index of 5 represents a situation in which water infiltration has decreased by a factor of 5 due to WR. The term *actual WR* refers to the results measured for moisture content at the time of sampling and *potential WR* refers to R-indices measured after drying at +40 °C.

2.2.5 Field measurements

The K_{fs} was measured, using a single-ring infiltrometer (cylinder Ø 30 cm, height 30 cm) connected with a Mariotte reservoir (Ø 20 cm, height 1.2 m). The cylinder was installed to a depth of 10 cm and the ponding height inside the cylinder was adjusted to 10 cm by the Mariotte siphon. The loss of water from the Mariotte reservoir was manually recorded at 30-s intervals. The K_{fs} was calculated from the steady-state infiltration rate (v_i) according to Darcy's law (Equation 10):

$$v_i = V/(A \times t) = K_{fs}(H_1 + L)/L \qquad \text{Equation 10}$$

where V is the volume of infiltrated water, A the cylinder area, H1 the ponding height and L the assumed profile height. Here, the profile height was 10 cm, corresponding to the depth of the cylinder bottom from the soil surface. The number of replicated measurements in spring was three and in autumn two. Bouwer (1986) listed several possible sources of errors and uncertainties concerning the approach at issue, from which the lateral divergence of flow below the cylinder and lack of information of the wetting front are probably the most relevant ones in this case. Thus, the field infiltration results are considered as comparable relative values between the sites.

2.3 Statistics

The results of image analysis of thin sections are presented with their standard errors of the means (SE) and least significant difference values (LSD, p = 0.05 with 6 degrees of freedom). Statistical analyses for the results obtained from the core samples as well as for the results of field-saturated hydraulic conductivity and water repellency were run, using SAS statistical software. The assumptions for analysis of variance (ANOVA) were tested from the residuals with the Kolmogorov-Smirnov test for normality and Levene's test for heteroskedasticity. To meet the normality assumptions, K_a, aggregate sorptivity and K_{fs} data were log-transformed. The values reported were back-transformed. In each case, pairwise comparisons of significant interactions were conducted with Tukey-Kramer tests. The p-values after multiple comparisons were Bonferroni-corrected. The K_{sat} was a response variable with a substantial number of results equal to zero, i.e. the samples had no conductivity at all or it were below the detection limit, and tests based on normality were not valid. Therefore, descriptive statistics are presented.

3 Results and discussion

3.1 Soil structure

3.1.1 Morphology

The surface horizon of the experimental soil was characterized by a high clay content (51%). The soil matrix at depth of 0–5 cm was composed of moderately to poorly sorted silty clay, including a relatively sparse distribution of larger mineral particles (Fig. 5, Fig. 2 in I). These mineral particles were mostly quartz and biotite, reflecting the composition of the parent rock from which they were crushed during the last glacial period more than 10 000 yr ago. In that matrix, there were abundant pseudomorphic nodules (iron and manganese) with sharp boundaries. The conditions in the surface horizon, i.e. the moisture regime varying from saturation to very dry, sufficient iron and manganese content, and extensive macroporosity, are conducive to formation of redox concentrations, as suggested by Zhang and Karathanasis (1997). Similar concentrations were also documented in deeper horizons of nearby pedons (Yli-Halla & Mokma, 2001; Yli-Halla et al., 2009). These studies suggested active formation of the nodules as well as the existence of relic nodules formed before human impact, i.e. artificial drainage in the first place, on soil. In the present study (I), the nodules mostly were homogeneously distributed and no differences in abundance, size (100–500 µm) or shape of nodules were observed between differently managed sites.

The dominant vegetation consisted of grass species with mostly even distributions of plant roots, as revealed by qualitative analysis of the thin sections. Few larger root channels at the natural site were attributed to a multispecies community, which according to Räty et al. (2010a) included common bent (*Agrostis capillaries* L.), *meadow vetchling (Lathyrus pratensis* L.) and dandelion (*Taraxacum officinale* F.H. Wigg.) in addition to the grass species timothy and meadow fescue growing in all BZs. Fibrous tissue type plant residues at all experimental sites were incorporated into clay aggregates close to the soil surface (Fig. 3b in I). Excrements of soil macrofauna, such as enchytraeidae or mites, were observed as loose continuous excremental infillings in cracks between aggregates (Fig. 3a in I). However, the plant residues and excrements were most abundant at the natural site. The highest biological activity at the natural site can be related to the abundant decaying plant material, which elevated the soil organic matter content up to 5%. In addition, the soil has not been disturbed for 14 yr, and therefore this site has attained a more coherent structure than the other sites, as was observed in the thin sections (Fig. 5, Figs. 1 and 3 in I).

3.1.2 Aggregates

Relatively high clay content, abundant organic matter, biological activity and soil desiccation are conducive to coalition of clay particles and formation of aggregates (Heinonen, 1951). Therefore, the experimental surface soil was characterized by angular blocky to subangular blocky aggregates and by abundant platy aggregates (i.e. the horizontal axis of the aggregate is longer than the vertical) at the grazed site (Fig. 5, Fig. 1 in I). In undisturbed soils, aggregates tend to have spherical shape (Horn & Smucker, 2005), granules and porous crumbs, the latter often related to excrements (Stoops, 2003). Such aggregates were hardly evident in this soil, probably due to the previous agricultural history of the soil or the natural physical stresses altering soil structure.

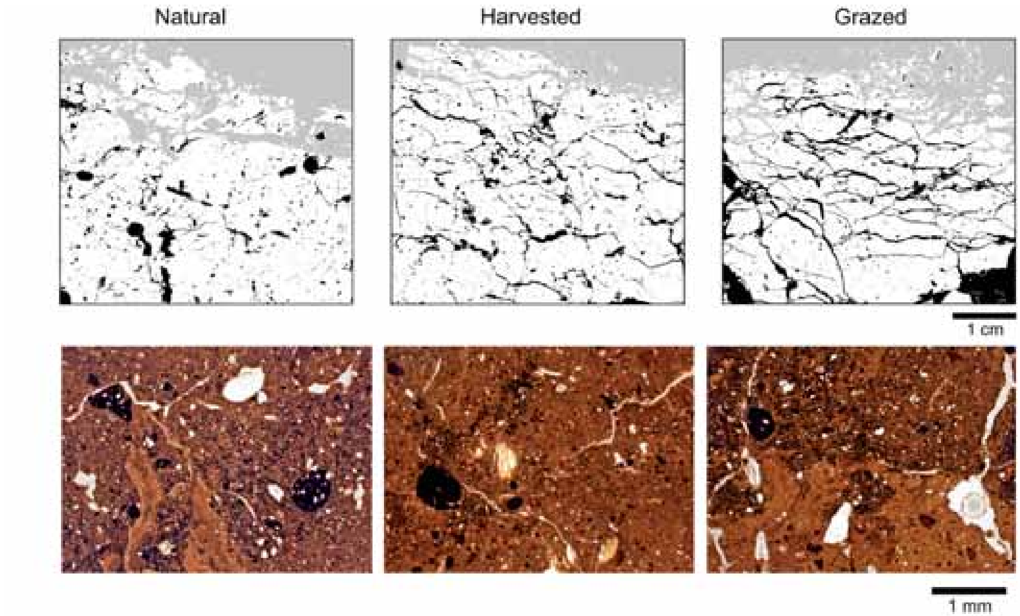

Figure 5. Examples of vertical thin sections at depths of 0–5 cm (above): black = pore space, white = soil matrix, grey = pore space connected to the surface. Selected micrographs of vertical thin sections with bright-field illumination (below).

Low intra-aggregate macroporosity was visible in all thin sections. The pore system inside the aggregates is often comprised of smaller and more tortuous pores than in bulk soil (Horn et al., 1994). The porosity due to pores smaller than 30 µm was from 43% to 52%, with the grazed site having significantly lower ($p < 0.05$) porosity in the surface horizon than the other sites. These pores are considered mostly as intra-aggregate pores, contributing especially to water storage capacity (VandenBygaart et al., 1999). However, the aggregates participate in water transport, since they act as sinks of water in the early stage of rain events on dry soil (Youngs et al., 1994). Aggregates of the surface horizons of the present soil imbibed water relatively rapidly (0.13 mm $s^{-½}$, Fig. 5 in II) compared with values presented previously (Youngs et al., 1994; Lipiec et al., 2006; Ferrero et al., 2007). Frequent and most intense wetting-drying in the soil surface were expected to favour formation of pores conducive to aggregate sorptivity, as previously suggested by Czarnes et al. (2000). Aggregate sorptivity at the grazed site did not differ from the other sites, regardless of differences in the amount of pores smaller than 30 µm in the surface horizon or differences in the shape the of aggregates. In contrast, the sorptivity was significantly lower in the deeper soil horizon (0.05 mm $s^{-½}$; Fig. 5 in II), which according to Lipiec et al. (2006), indicates denser, less porous aggregates than at the soil surface. Reduced sorptivity of the aggregates in the deeper horizon enhances bypass flow.

Aggregate-related pedofeatures of dense clay infillings (Fig. 5, Fig. 2 in I) were found in the soil studied. These aggregates were observed at all sites and were described as clay intrusions. The infillings were incorporated into the clay matrix, and they had different sizes and shapes ranging from smooth to undulating with medium sphericity or with sharp edges. They seemed to be altered after formation, since they were pervaded by shrinkage cracks. Several aggregates showed relatively sharp angled shapes, although the shape of the aggregates at the natural site was more rounded than in the grazed BZ. Possible reasons include physical stress during freezing, shrinking and also hoof pressure at the grazed site.

Dispersion of clay is a prerequisite for formation of clay-rich aggregates. During the spring thaw there is an excess of snow melt water with low electrical conductivity. The topsoil remains saturated since the deeper horizons are still frozen, creating conditions conducive to clay dispersion. When frozen soil eventually thaws, the clay suspension infiltrates through the soil into subsurface drainage pipes or is attached to ped surfaces. Illuvial clay observed in thin sections (Yli-Halla et al., 2009) and clay transport from surface soil to drainage pipes (Uusitalo et al., 2001) have both been documented in soils close to the present site. Clay intrusions may have formed when the infiltration of clay suspension was restricted and suspension became dried/sedimented in situ. It is also possible, and would be in accordance with the functional idea of the BZ, that transport of the sediment from the above field (i.e. surface runoff) has acted as an additional source of clay particles.

These clay intrusions, however, directly indicate dispersion of clay in the surface horizon. Clay dispersion contributes to erosion and nutrient transport from agricultural fields, but it seems to play a role in the development of soil structure as well. The physical properties of the bulk soil aggregates and clay intrusions are likely to be distinct.

3.1.3 Macropores

Mutual arrangement of aggregates defines the interaggregate pore system and soil structure (Horn & Smucker, 2005). Weak or moderate degrees of pedality observed in the thin sections indicate that the aggregates were not entirely surrounded by pores (Fig. 5, Fig. 1 in I). The main pore types were vughs and channels with mamillate surface characteristics and planes showing straight or zig-zag crack patterns. Generation of such pore types is attributed to burrowing animals, plant roots and soil shrinkage (Stoops, 2003). Accordingly, biological activity was evident in thin sections (Chapter 3.1.1), as well as contribution of shrinkage to formation of soil structure (Chapter 3.3).

Large, irregularly elongated pores, occurring as voids between subangular blocky, angular blocky and platy aggregates, were the most common at all sites (I). These pores were partially accommodated and corresponded to over 90% of the pores observed in thin sections (Tables 2 and 3). The pore pattern dominated by irregular and elongated pores has been commonly documented in clayey soils (Murphy et al., 1977b; Pagliai & Vittori Antisari, 1993; Servadio et al., 2005; Constantini et al., 2006), although most of these studies were done in soils and climates very different from those of the present study.

The rounded and irregular pores were most numerous, but due to their small size (mostly smaller than 300 µm) these pores consisted of less than 10% of the macropore area observed in the thin sections (Table 2 and 3). Based on the size and shape (sometimes mammillated), they were probably root channels and/or burrows of enchytraeids, as suggested by Murphy et al. (1977b) and VandenBygaart et al. (2000).

In the vertical thin sections (Fig. 5, Fig. 1 in I), horizontally oriented macropores increased from the natural to the harvested site and further to the grazed site. Applied external stress tends to compact soil and alters pore morphology. In the present study, the soil of the grazed site was subjected to external stress by cattle. In the study of Räty et al. (2010a), this pressure was estimated to be approximately 125 kPa per hoof. In Italy, Servadio et al. (2005) studied silty clay soil subjected to compaction by tractor. They found generation of platy structure and decrease in macroporosity, especially the large (> 500 μm) elongated pores. Such pore characteristics are associated with soil compaction, decreased hydraulic conductivity as well as poor vertical root growth (Murphy et al., 1977b; Servadio et al., 2005). Indeed, adverse effects of compaction and platy soil structure on hydraulic properties were also noted in the present study (Chapter 3.2).

The results obtained using indirect physical methods and measured hydraulic properties (Table 4) supported the previous findings. In the surface horizon (0–5 cm) of the grazed site, bulk density was significantly higher (1.24 g cm^{-3}) than at the natural and harvested sites (1.04 and 0.99 g cm^{-3}, respectively). At the depth of 5–10 cm, the bulk density at all sites varied between 1.20 and 1.23 g cm^{-3}. These results indicate: i) significantly different bulk density of natural and harvested sites between the two horizons, ii) homogenous bulk density profile at the grazed site and iii) equal bulk densities at all sites in the deeper soil horizon. Similar trends were observed for total porosity (53–63%), macroporosity > 30 μm (6–12%, no significant difference between the natural and the grazed sites) and large macro pores > 300 μm (2–6%) (Fig. 1 in II).

The results of bulk density and macroporosity seem to be representative of the clay soils of Finland. For example, in the study of Pietola et al. (2005) the range of bulk density (0.88–1.16 g cm^{-3}) and large macropores (1–5%, 0–15 cm) of a nearby heavy clay soil under variable intensities of cattle trampling were essentially the same. The range of macropores also corresponded to those presented by Regina and Alakukku (2010) for clayey soils of Finland under no till or annual ploughing (6–12%, 0–20 cm). The porosity of app. 60% measured in the surface layer of the natural and harvested sites was high, close to the values found after spring harrowing of clay soil in the study of Aura (1983). High porosity of the natural and harvested sites was considered as an indication of well-developed soil structure, since these sites have been under no or only slight disturbance for 14 yr.

Earthworm burrows are regarded as important drainage routes in boreal clay soils (e.g. Alakukku et al., 2010). In this study, root-related pore structures as well as indications of burrowing animals (faecal pellets), which tend to produce biopores, were observed in the thin sections (Figs. 3a and 3c in I). Yet, rounded pores larger than 300 μm were not detected by image analysis. The lack of these pores was partly attributed to shrinkage cracks joining with initially rounded pores, which disturbs the rounded shapes. This finding is in line with the suggestion of Shipitalo et al. (2004) that soil shrinkage, freezing or compaction could have disrupted earthworm burrows in the plough layer of a nearby field of sandy clay. Cracks connected to well-conductive biopores may facilitate distribution flow in the surface horizon and, therefore, increase preferential flow.

The results presented in this chapter show that indirect methods complemented with data from additional study techniques, such as study of thin sections, provide good insight into the soil pore system. In the present soil, the results revealed the complexity and bimodality of the pore system, as well as differences in the quantity and quality of the pore systems between management practices. Grazing seems to compact soil, reduce macroporosity and generate platy soil structure. Deterioration of pore geometry and continuity adversely affects soil hydraulic properties (Dörner & Horn, 2006).

Table 2. Soil porosity (> 50 µm) expressed as a percentage of the investigated area of vertical thin sections and the percentage of the pore shape classes.

Management	Porosity (%)	% of total porosity		
		Rounded	Irregular	Elongated
Natural	13.6	1	4	95
Harvested	12.4	2	6	92
Grazed	17.4	1	4	95
SE	3.9	0.5	1.4	1.9
LSD	13.4	1.7	4.9	6.6

Table 3. Soil porosity (> 50 µm) expressed as a percentage of the investigated area of horizontal thin sections and the percentage of the pore shape classes.

Management	Porosity (%)	% of total porosity		
		Rounded	Irregular	Elongated
Natural	19.2	1	3	95
Harvested	17.6	2	4	94
Grazed	13.0	3	6	91
SE	3.8	0.6	1.4	2.0
LSD	13.0	2.1	4.7	6.8

3.2 Hydraulic properties

The active role of macropores was evident in the soil studied. At a matric potential of -3 kPa, K_a attained levels (0–5 cm, 43.6 µm^2 and 5–10 cm, 19.0 µm^2, Fig. 3 in II) from which it increased only slightly upon further soil drying. This outcome indicated that large, irregularly elongated pores observed in thin sections were emptied and began to dominate airflow. These macropores were responsible for the high K_{fs} levels measured in spring, as well as high K_{sat} values of some samples, regardless of the long wetting period in the laboratory (Table 4). In accordance with our results, Pagliai et al. (2004) suggested that continuous elongated pores, i.e. the dominant pore type of the present soil, are conducive to water flow.

However, a substantial drop in K_{fs} was observed between the measurements in spring and autumn (Table 4). Although the number of replicates was low, the results suggest remarkable seasonal variation in the soil pore system and hydraulic properties. The lower K_{fs} in late autumn was attributed to the long wet period before measurements. These conditions allow the (partially) accommodated macropores observed in the thin sections to swell and close, which increased pore tortuosity and decreased conductivity (Bullock & Murphy, 1980). The high seasonal variation in K_{fs} of clay soils in Sweden was attributed to shrinkage-induced structural changes as well (Messing & Jarvis, 1990).

As was shown in the previous chapter (3.1.3), grazing adversely affected the pore system in the surface horizon. These findings were reflected in pore functioning as well, which was evident in the several variables measured. The levels of K_{sat} tended to decrease as follows: grazed < har-

Table 4. Saturated hydraulic conductivity (K_{sat}) at depth of 0–5 cm and 5–10 cm (median, first and third quartile), and field-saturated hydraulic conductivity (K_{fs}, SE in parentheses) measured in spring (n = 3) and autumn. In autumn n = 2, both values are presented. Means with the same letter do not differ at p = 0.05.

	K_{sat} (cm h^{-1})			K_{fs} (cm h^{-1})
Site	Median 0–5 cm	Q_1 0–5 cm	Q_3 0–5 cm	Spring
Natural	3	0	11	49 a (+13, -10)
Harvested	13	8	54	82 a (+22, -17)
Grazed	0	0	1	11 b (+3, -2)
Site	5–10 cm	5–10 cm	5–10 cm	Autumn
Natural	32	0	175	2 and 8
Harvested	56	5	241	1 and 3
Grazed	8	0	174	1 and 1

vested < natural site (Table 4). The K_{fs} (in spring) was significantly lower at the grazed site than at the others (Table 4). The same pattern was observed in K_a, since significantly lower K_a was detected at the grazed site than at the natural and harvested sites (Fig. 3 in II). Calculated organization indices (Fig. 6) suggested that poor air and water flux resulted not only from differences in the air-filled porosity, but also from poorer vertical continuity of the pores at the grazed site (Fig. 4 in II). Consequently, the formation of platy soil structures due to grazing adversely effected pore functions, since the solution was forced to go through horizontally oriented pores, which makes the flow route longer. It is also possible that compaction increases narrow pore necks that hinder the flow.

A high number of samples (22 of the 60 samples) had K_{sat} values below the detection limit, regardless of the abundant macroporosity observed. This finding, i.e. discrepancy in capacity and functionality, emphasizes the high importance of pore characteristics and connectivity of the pores in swollen soils. On the other hand, the upper quartile of K_{sat} presented in Table 4 with some high values indicates high spatial variability in hydraulic conductivity. These conditions favour preferential flow, as rainwater falling on the less conductive patches of the soil seeks its way to more conductive spots (Jarvis, 2007).

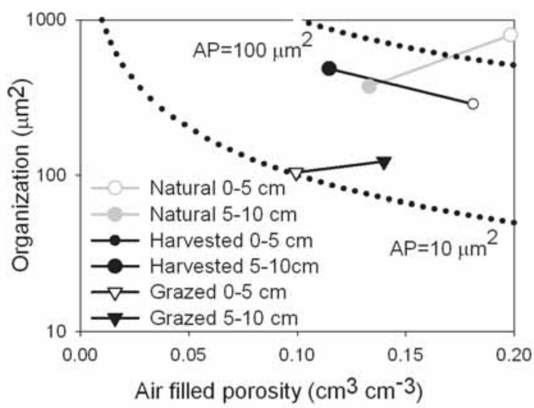

Figure 6. Organization and air-filled porosity at matric potential of -3 kPa. Isolines for air permeability (AP) are shown. Blank symbols stand for depth of 0–5 cm and filled symbols for depth of 5–10 cm.

3.3 Shrinkage properties

The total shrinkage from saturation to dryness (at +105 °C) was from 6.7% to 10.5% of the total soil volume (III), which are relatively low values compared with the volume reduction of as much as 21% in British clayey soil containing smectite (Reeve et al., 1980), as well as tropical soils containing smectite (Cornelis et al., 2006a). Yli-Halla et al. (2009) attributed the low COLE values (0.03–0.07 cm cm^{-1}) of a Finnish clay soil to a micaceous-vermiculitic mineralogy. Sippola (1974) showed that nonswelling minerals such as quartz and feldspar are relatively abundant, even in the coarse clay fraction, while smectite is virtually absent in these soils.

The majority of the volume changes occurred upon drying the soil from saturation to +40 °C (III). Based on the data of Heikinheimo and Fougstedt (1992), it could be assumed that such high temperatures may also occasionally be reached in surface soils in Finland. However, significant volume loss was already observed at a matric potential of only -50 kPa (2.1–3.8% of the total soil volume, III). These findings suggest that even relatively small moisture changes may cause changes in soil volume, but intense cracking requires soil drying close to the air-dry state, as suggested by Peng and Horn (2007).

All soil samples studied, regardless of the depth, showed clear structural shrinkage zones (Fig. 7), being responsible for 18–28% of water loss (Table 5). The structural shrinkage observed was consistent with abundant interaggregate macroporosity of the soils, i.e. abundance of large, irregularly elongated pores (see Chapter 3.1.3). The rest of the water loss, over 70%, occurred in the other shrinkage zones, in which swelling and condensation water were removed from the micropore regime (Braudeau et al., 2004). This result is in agreement with the previous statement that intra-aggregate pores with relatively fast sorptivity contribute significantly to water flow and water storage in the present soils (Chapter 3.1.2). Moreover, this result emphasizes the importance of intra-aggregate pores in the structural changes of soil,

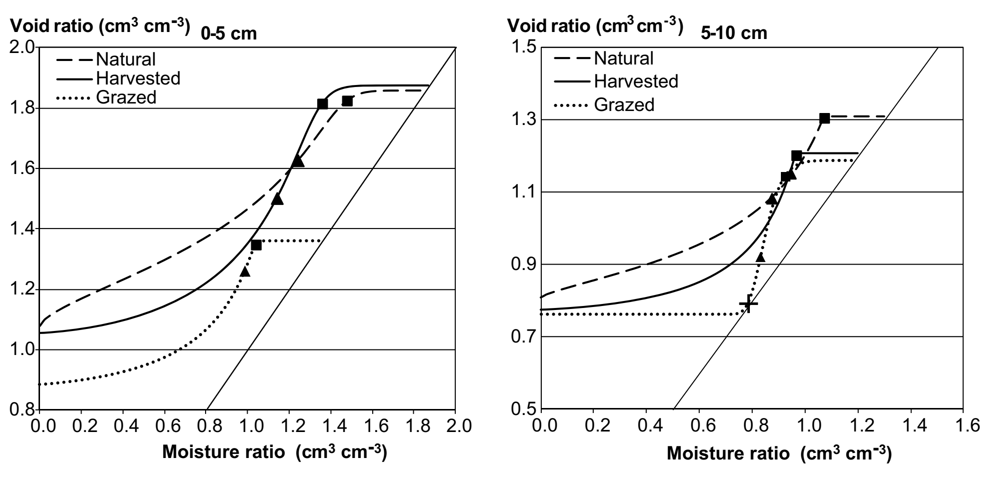

Figure 7. Shrinkage curves of the natural, harvested and grazed sites (mean of three replicates) for depths of 0–5 cm and 5–10 cm. The symbols show the transition points between 1) the structural and proportional zone (■), 2) proportional and residual zone (▲) and 3) the residual and zero-shrinkage zone(+). The straight line stands for the 1:1 line

Table 5. Loss of water and volume change in the four shrinkage zones and the slope of the proportional shrinkage zone.

Site	Loss of water (%)				Loss of volume (%)				Slope
0–5 cm	Structural	Proportional	Residual	Zero	Structural	Proportional	Residual	Zero	Proportional
Natural	20	13	67	0	4	25	70	0	0.8
Harvested	28	12	61	0	7	38	55	0	1.4
Grazed	24	4	73	0	2	19	79	0	1.7
Site 5–10 cm									
Natural	18	15	67	0	1	44	55	0	1.1
Harvested	20	2	78	0	1	12	87	0	2.3
Grazed	22	8	4	66	10	52	30	7	2.3

because the loss of soil volume is attributed to aggregates.

The proportional shrinkage zones of the shrinkage curves were relatively small. (Fig. 7, Table 5). In homogenized soil, proportional shrinkage plays a dominant role as water is removed from the soil as a whole (Gusli et al., 1994). In contrast, aggregates tend to restrict proportional shrinkage (Stirk, 1954). The lack of a zero-shrinkage zone (except at grazed site 5–10 cm) indicates that water removal even under very dry conditions still rearranges aggregates and/or single soil particles. Although it could be argued that the transition points between the proportional – residual and residual – zero shrinkage zones (III) could have been defined more precisely if additional data points between moisture ratios of 0.2 and 0.8 were measured, the main reasons for the differing shrinkage behaviour between the sites studied lies in the differences in soil structural stage, initial bulk density and organic carbon content, as earlier suggested by Reeve et al. (1980) and Peng et al. (2007). For example, in the surface horizon the lowest total shrinkage was measured at the grazed site (0–5 cm), which had the highest bulk density, lowest macroporosity and lowest organic carbon content (Chapter 3.1.3 and Table 1).

Soils having an extensive structural shrinkage zone with high water loss over a wide suction range are considered to be well-structured and rigid. Although water loss in the present study was substantial in the structural shrinkage zone (Table 5), only a slight increase in hydraulic stress caused significant changes in the pore system. The transition point between the structural and proportional shrinkage zones occurred at a matric potential close to -6 kPa, indicating rather unstable macropores. An observation that is even more striking was the steep slope of the proportional shrinkage zone (Table 5). A slope larger than 1 indicates collapse of the interaggregate pore system (Groenevelt & Grant, 2001), and such values were observed at all sites (1.1–2.3) except the surface horizon at the natural site (0.8). The natural site has been undisturbed for 14 yr and it had the most coherent structure (Chapter 3.1.1, Fig. 5) and the highest organic matter content (Table 1), suggesting a well-developed structure.

The collapse of the interaggregate pore system can be associated with freezing and thawing in the previous winter and water saturation in autumn and spring. These processes weaken soil structure (Kay et al., 1985; Bullock et al., 1988; Gusli et al., 1994). The collapse of the interaggregate pores was observed in the results of K_a (II). The permeability tended to decrease between the drying step from -3 to -6 kPa, which coincides with the beginning of the proportional shrinkage zone. These findings indicate the immediate effect of shrinkage on soil flux processes.

After collapse and rearrangement of the aggregates, they do not return to their original organization upon reswelling. Consequently, the shrinkage properties detected will not apply to subsequent swelling and shrinkage cycles. This conclusion is supported by the results of Peng and Horn (2007), who found that in mineral soils the maximum void ratio did not return to its initial level after complete shrinkage. Moreover, frequent shrink-swell cycles tend to increase large pores and strengthen soil structure (Peng et al., 2007) and maximum stability of the aggregates is reached in autumn (Bullock et al., 1988). Therefore, it can be assumed that the shrinkage characteristics presented in this thesis are time-dependent as well. Soil structure develops towards the lowest free energy stage in summer, and is altered again by weather conditions in autumn and winter. This process may partly explain the differences in K_{fs} measured in spring and autumn (Table 4).

According to Warsta (2007) the performance of a dual-permeability, three-dimensional process model run in a clay field in southern Finland could be improved if a dynamic cracking and swelling component were included. In the heavy clay soil of Finland, Knisel and Turtola (2000) found that transport of particles into subsurface drains was unsatisfyingly simulated when shrinkage was not adequately included in the Groundwater Loading Effects of Agricultural Management Systems (GLEAMS) model. The results presented in this thesis support the previous suggestion, since shrinkage causes a significant volume change in the clay soils of Finland. However, shrinkage properties vary, depending on soil management, suggesting that different parameters are required, e.g. in the modelling of ploughed and no-till soils.

3.4 Water repellency

The R indices, indicating the degree of WR, are presented in Figure 8. These results indicate that all sites were subcritically water-repellent (R > 1.95) at the time of sampling and that these sites showed potential WR as well. This result is not surprising since in Sweden, existence of WR has been documented in Gyttja soil (mixture of organic and minerogenic material), loamy sand and clayey soils (Berglund & Persson, 1996; Dekker et al., 1999; Jarvis et al., 2008).

The behaviour of WR in the soil studied was in agreement with the trends described in the literature. Both actual and potential WR were significantly higher at the soil surface than in the lower layer, which could be attributed to accumulation of organic material at the soil surface (Table 1). WR is commonly observed under permanent nondisturbed vegetation, because in cultivated soils oxidation of organic matter and disturbance of aggregates lower WR (Dekker & Ritsema, 1996; Harper et al., 2000; Hallett et al., 2001). Grazing is conducive to WR, which, according to Hallett and Young (1999), can be attributed to enhanced microbiological activity, due to cattle manure. However, differences between the sites were rather small, which could be partly explained by the high spatial variation in WR. Higher WR in the clayey surface soil studied compared with WR in sandy soils of Finland (IV) supports the suggestion of Doerr et al. (2006) that fine-textured aggregated soils are prone to WR as well.

To assess the effect of soil moisture on WR, critical soil water content was defined (IV). This water content is considered to divide soil into repellent (below) and non-repellent (above), as suggested by Dekker (1998). For soil samples taken from the clayey surface soil, this value was 43 vol.%. Below this moisture content (or matric potential < -6 kPa) all sites were subcritically water-repellent. In Sweden, significant WR was observed in a clay soil under grass cover at a moisture content of 36% (Jarvis et al., 2008). Dekker and Ritsema (1996) assumed water-repellent aggregates and prism faces to be the reasons for lowered water uptake in clayey soils in the Netherlands at a moisture content below 42%. These results are in relatively good agreement with the findings of this thesis and indicate that WR contributes to water flow over a wide moisture range.

Due to the moisture-dependent nature of the existence and degree of WR, it is appropriate to assess WR at various moisture contents (De Jonge et al., 1999). In the present study, the samples were pretreated in the laboratory, i.e. saturated with water and dried stepwise to assess the effect of moisture change on WR. The potential WR was measured afterwards, which may have causes some bias between the actual and potential WR (Czarnes et al., 2000; Doerr et al., 2007). The potential WR of the soil was detected after drying of the samples at +40 °C, +70 °C and +105 °C, because there is no commonly accepted temperature used to define potential WR (IV). The temperature of +40 °C faded out differences in soil moisture content (Fig. 8). The higher drying temperatures may alter hydrophobic coatings (Dekker et al., 1998), and a temperature of +40 °C can occur occasionally at the soil surface in Finland, as discussed earlier. Therefore, detection of potential WR after drying the field-moist samples at +40 °C is an appropriate procedure for the surface soil. In contrast, lower drying temperatures should be considered to determine potential WR in the lower horizons (which remain cooler) to avoid artefacts.

Rapid preferential flow was observed in the field measurements conducted in spring (Table 4). The results of the actual WR (Fig. 8) represent the state of soil hydrophobicity at that time, indicating the existence of WR. However, the effect of WR on infiltration cannot be distinguished from the data. It could be assumed that WR enhances bypass flow and, thus, reduces the ability of a soil to store water for plants in the case of intensive showers falling on initially dry and water-repellent soil. This example demonstrates that even if WR is recognized, its implications for water flow are difficult to isolate, as suggested by Shakesby et al. (2000).

The results presented above indicate the transient nature of WR. WR affects soil hydraulic properties over a wide moisture range and its existence and degree vary seasonally. It seems unlikely that WR contributes to the generation of surface runoff in wet periods, i.e. in autumn and early spring. In contrast, WR is most severe during dry spells, which coincides with soil shrinkage and cracking, as was shown in the previous chapter (3.3). Dekker and Ritsema (1996) suggested that drying and opening of shrinkage cracks are conducive to the generation of water-repellent aggregate faces along crack walls if hydrophobic substances are present. Therefore, it is probable that WR enhances preferential vertical water flow in dry periods rather than affecting surface runoff. Jarvis et al. (2008) came to a similar conclusion in their study of clay soils in Sweden.

In the case of severe WR or very slow aggregate sorptivity (due to poor intra-aggregate porosity), the initiation of preferential flow may be instantaneous, as suggested by Clothier et al. (2008). They calculated a monetary value of soil ecosystem services that are affected by preferential flow and estimated the amount to be several hundred billion US dollars per year globally. The existence of WR may increase in

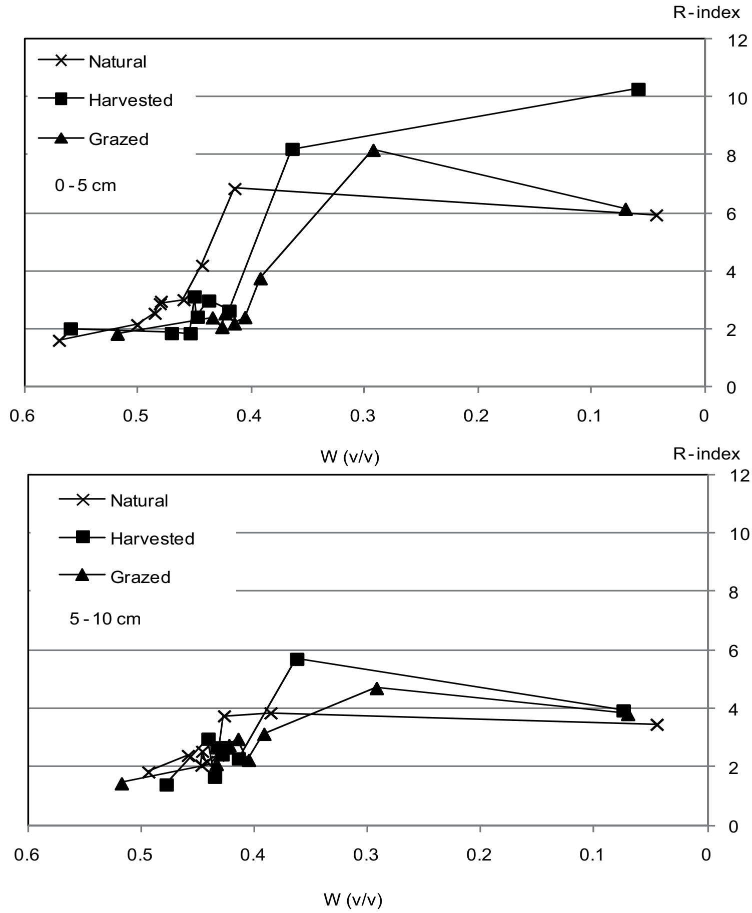

Figure 8. Water-repellency (R-index) of clayey surface soil at different moisture contents (W, v/v). Drying treatments from left to right: 0, -3, -6, -10, -15, -30, and -50 kPa ,field moisture (actual WR), +40 °C (potential WR).

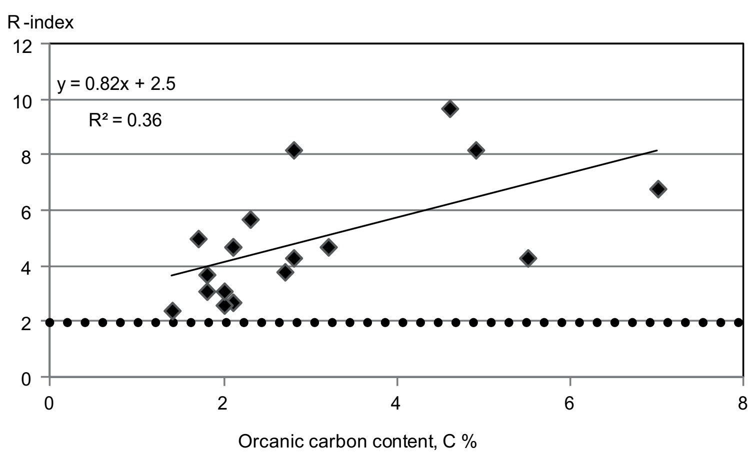

Figure 9. Actual water repellency (at the time of sampling) in relation to the content of soil organic carbon (C%) in clayey and sandy surface soils (IV). Dotted line stands for the R index 1.95.

agricultural soils, because direct drilling and conservation tillage tend to increase the content of organic matter in surface horizons and reduce soil disturbance in comparison to ploughing (see discussion above). Another relevant, but very unambiguous, factor affecting WR is climate change. Changes in temperature, rainfall and precipitation may affect the degree and occurrence of WR, which pose further challenges to the predictability of this phenomenon. Due to the highly dynamic nature of WR and its hardly distinguishable effect on water flow, the development of hydraulic models accounting for the effects of WR seems to be an important but complicated task.

Bouma (2006) suggested that screening of the existence of WR should be made before simulating soil-water movement of the soil concerned. In the present study, I attempted to facilitate this task. In Figure 9, the actual R indices (at the time of sampling) of the experimental sites are plotted against the degree of organic carbon content. Additional WR results were included (IV). These results were obtained from three adjacent sites at the Lintupaju experimental field and from three sites on a sandy soil (Haplic Regosol) situated in Maaninka, central Finland (for further details, see IV). The organic carbon content did not unambiguously explain the degree of WR, which is in agreement with the study of Harper et al. (2000). However, WR seems to occur in soils with wide ranges of organic carbon contents. Only in soils with very low contents of organic carbon, clearly less than 2%, may WR have an insignificant effect on soil hydraulic properties. Such low contents of organic carbon are seldomly found in our agricultural soils (Sippola & Tares, 1978). Although the data are relatively small, it seems that WR is a common feature in the agricultural soils of Finland.

4 Conclusions

In this thesis, the physical properties of a boreal clayey surface (0–10 cm) soil subjected to different management practices of perennial vegetation were studied. The loss of soil volume upon shrinkage from water saturation to dry soil was up to 10%, indicating that changing moisture conditions greatly impact the physical properties of the soil. Although drying in the field is usually less intense, the relatively high content of clay (> 50%) and frequent wetting-drying cycles favour the development of aggregates. Indeed, the structure of the surface soil was characterized by subangular to angular blocky aggregates with weak or moderate degrees of pedality. In addition, root-related pore structures, animal burrows and faecal pellets were observed in thin sections, indicating the importance of biological activity on the generation of soil structure.

Large, partially accommodated, irregularly elongated pores characterized the inter-aggregate macropore system. The largest macropores became empty at low suction (-3 kPa) and allowed soil aeration since then. Some of the pores maintained their ability to conduct water when saturated, although swelling tends to close accommodated pores. In the field, hydraulic conductivity was greater in spring than after a long wet period in autumn. Therefore, soil moisture conditions greatly affect the functioning of these macropores.

A bimodal pore system prevailed in the soil studied. Surface soil aggregates were characterized by low macroporosity, abundant pores smaller than 30 µm and relatively fast water sorption. Such aggregates contribute to water storage as well as to water flow. They may hinder generation of surface runoff or preferential flow, because they act as sinks of water during the first moments of rain events. However, the sorptivity of the aggregates decreased in the lower soil horizon, which may enhance bypass flow.

The differences in soil properties between the management practices were evident in the shallow top layer (0–5 cm) of the soil. The rather coherent structure developed in 14 yr at the nondisturbed natural site. The horizontally oriented macropores increased from the natural site to the harvested site and further to the grazed site. The compacted platy structure at the grazed site adversely affected the soil hydraulic properties. Both description of the soil thin sections and calculated organization indices suggested that the altered hydraulic properties resulted from the poorest continuity of the pores combined with compaction.

Considering soil structure, pore characteristics, pore functions and hydraulic properties, parallel use of indirect physical methods and thin section methodology proved to be successful. Although three-dimensional methods for describing the soil pore system are readily available, the use of thin sections is supported by the fact that they provided valuable information on soil forming processes. For example, aggregate-related pedofeatures of dense infillings described as clay intrusions were found at all sites. The formation of these intrusions was attributed to clay dispersion and/or translocation during the spring thaw and drying of the suspension in situ. These processes generate very new aggregates with physical properties most probably different from those of the bulk soil aggregates.

The results suggested that subcritical WR is a common feature in the agricultural soils of Finland, unless the soil has a very low content of organic carbon (< 2%). WR occurred over a wide moisture range, because all sites showed WR before they were dried to field capacity (-10 kPa). In models dealing with water flow during the wet seasons (autumn and spring), disregarding of WR may be reasonable. In contrast, models developed to describe soil hydrology during the growing season should account for WR. In dry summers, interaction of WR and soil shrinkage may significantly enhance preferential flow and reduce the capability of the surface soil to store water during rain events.

Abundant macropores were reflected in the soil shrinkage curve in which the structural shrinkage zone was evident. The macropore system, however, proved to be unstable against increasing hydraulic stress caused by soil drying. The instability was observed as a collapse of inter-aggregate pores before the soil was dried to field capacity. The weak structure of the soil in spring was attributed to soil freezing in winter and wetness in autumn and at the time of snowmelt. After structural collapse, soil would not reach the same volume during swelling and, thus, the subsequent shrinkage curve would be different. These subsequent shrink-swell cycles, however, convert the soil structure towards a more stable state in the course of summer. Thus, the pore characteristics as well as the hydraulic properties of soil are time-dependent.

References

Adams, J.E. & Hanks, R.J. 1964. Evaporation from soil shrinkage cracks. Soil Science Society of America Proceedings 28: 281-284.

Alakukku, L. 1996. Persistence of soil compaction due to high axle load traffic. I. Short-term effects on the properties of clay and organic soils. Soil & Tillage Research 37: 211-222.

Alakukku, L. 1998. Properties of compacted fine-textured soils as affected by crop rotation and reduced tillage. Soil & Tillage Research 47: 83-89.

Alakukku, L., Nuutinen, V., Ketola, E., Koivusalo, H. & Paasonen-Kivekäs, M. 2010. Soil macroporosity in relation to subsurface drain location on a sloping clay field in humid climatic conditions. Soil & Tillage Research 106: 275-284.

Aura, E. 1983. Soil compaction by the tractor in spring and its effect on soil porosity. Journal of the Scientific Agricultural Society of Finland 55: 91-107.

Baumgartl, T.H. & Köck, B. 2004. Modelling volume change and mechanical properties with hydraulic models. Soil Science Society of America Journal 68: 57-65.

Berglund, K. & Persson, L. 1996. Water repellence of cultivated organic soils. Acta Agriculturae Scandinavica. Section B, Soil and Plant Science 46: 145-152.

Blackwell, P.S., Ringrose-Voase, A.J., Jaywardane, N.S., Olsson, K.A., MCKenzie, D.C. & Mason, W.K. 1990. The use of air-filled porosity and intrinsic permeability to air to characterize structure of macropore space and saturated hydraulic conductivity of clay soils. Journal of Soil Science 41: 215-228.

Bouma, J. 2006. Hydropedology as a powerful tool for environmental policy. Geoderma 131: 275-286.

Bouma, J., Jongerius, A., Boersma, O., Jager, A. & Schoonderbeek, D. 1977. The function of different types of macropores during saturated flow through four swelling soil horizons. Soil Science Society of America Journal 41: 945-950.

Bouma, J. & Dekker, L.W. 1978. A case study on infiltration into dry clay soil I. Morphological observations. Geoderma 20: 27-40.

Bouwer, H. 1986. Intake rate: Cylinder infilrometer. In: Klute, A (ed.). Methods of soil analysis. Part 1. Physical and mineralogical methods. Second Edition. American Society of Agronomy, Inc. Soil Science Society of America, Inc. Madison, Wisconsin USA, pp. 825-844.

Braudeau, E., Frangi, J-P. & Mohtar, R.H. 2004. Characterizing nonrigid aggregated soil-water medium using its shrinkage curve. Soil Science Society of America Journal 68: 359-370.

Brewer, R. 1976. Fabric and mineral analysis of soils. Krieger, New York, USA. 482p.

Bronswijk, J.J.B. 1990. Shrinkage geometry of heavy clay soil at various stresses. Soil Science Society of America Journal 54: 1500-1502.

Bronswijk, J.J.B. 1991. Drying, cracking, and subsidence of a clay soil in a lysimeter. Soil Science 152: 92-99.

Bullock, M.S., Kemper, W.D. & Nelson S.D. 1988. Soil cohesion as affected by freezing, water content, time and tillage. Soil Science Society of America Journal 52: 770-776.

Bullock, P., Fedoroff, N., Jongerius, A., Stoops, G. & Tursina, T. 1985. Handbook for soil thin section description. Waine Research Publications., Wolverhampton, 152 p.

Bullock, P. & Murphy, C.P. 1980. Towards the quantification of soil structure. Journal of Microscopy 120: 317-328.

Clothier, B.E. Green, S.R. & Deurer, M. 2008. Preferential flow and transport in soil: progress and prognosis. European Journal of Soil Science 59: 2-13.

Constantini, E.A.C., Pellegrini, S., Vignozzi, N. & Barbetti, R. 2006 Micromorphological characterization and monitoring of internal drainage in soils of vineyards and olive groves in central Italy. Geoderma 131: 388-403.

Cornelis, W.M., Corluy, J., Medina, H., Díaz, J., Hartmann, R., Van Meirvenne, M. & Ruiz, M.E. 2006a. Measuring and modeling the soil shrinkage characteristic curve. Geoderma 137: 179-191.

Cornelis, W.M., Corluy, J., Medina, H., Hartmann, R., Van Meirvenne, M. & Ruiz, M.E. 2006b. A simplified parametric model to describe the magnitude and geometry of soil shrinkage. European Journal of Soil Science 57: 258-268.

Czarnes, S., Hallett, P.D., Bengough, A.G. & Young, I.M. 2000. Root- and microbial-derived mucilages affect soil structure and water transport. European Journal of Soil Science 51: 435-443.

DeBano, L.F. 2000. Water repellency in soils: a historical overview. Journal of Hydrology 231-232: 4-32.

De Jonge, L.W., Jacobsen, O.H. & Moldrup, P. 1999. Soil water repellency: effects of water content, temperature and particle size. Soil Science Society of America Journal 63: 437-442.

Dekker, L.W. 1998. Moisture variability resulting from water repellency in Dutch soils. Doctoral thesis, Wageningen Agricultural University, The Netherlands, 240 p.

Dekker, L.W., Doerr, S.H., Oostindie, K., Ziogas, A.K. & Ritsema, C.J. 2001. Water repellency and critical soil water content in a dune sand. Soil Science Society of America Journal 65: 1667-1674

Dekker, L.W. & Ritsema, C.J. 1994. How water moves in a water reprellent sandy soil. 1. Potential and actual water repellency. Water Resources Research 30: 2507-2517.

Dekker, L.W. & Ritsema, C.J. 1996. Preferential flow paths in a water repellent clay soil with grass cover. Water Resources Research 32: 1239-1249.

Dekker, L.W., Ritsema, C. J., Oostindie, K. & Boersma, O.H. 1998. Effect of drying temperature on the severity of soil water repellency. Soil Science 163: 780-796.

Dekker, L.W., Ritsema, C.J., Wendroth, O., Jarvis, N., Oostindie, K., Pohl, W., Larsson, M. & Gaudet, J-P. 1999. Moisture distributions and wetting rates of soils at experimental fields in the Netherlands, France, Sweden and Germany. Journal of Hydrology 215: 4-22.

Dekker, L.W., Oostindie, K. & Ritsema, C.J. 2005. Exponential increase of publications related to soil water repellency. Australian Journal of Soil Research 43: 403-441.

Doerr, S.H., Ritsema, C.J., Dekker, L.W., Scott, D.F. & Carter, D. 2007. Water Repellence of soils: new insights and emerging research needs. Hydrological Processes 21: 2223-2228.

Doerr, S.H., Shakesby, R.A., Dekker, L.W. & Ritsema, C.J. 2006. Occurrence, prediction and hydrological effects of water repellency amongst major soil and land-use types in a humid temperate climate. European Journal of Soil Science 57: 741-754.

Doerr, S.H., Shakesby, R.A. & Walsh, R.P.D. 2000. Soil water repellency: its causes, characteristics and hydro-geomorphological significance. Earth-Science Rewievs 51: 33-65.

Doerr, S.H. & Thomas, A.D. 2000. The role of soil moisture in controlling water repellency: new evidence from forest soils in Portugal. Journal of Hydrology 231: 134–147.

Dörner, J. & Horn, R. 2006. Anisotropy of pore functions in structured Stagnic Luvisols in the Weichselian moraine region in N Germany. Journal of Plant Nutrition and Soil Science 169: 213-220.

FAO, 2006. World reference base for soil resources 2006. A framework for international classification, correlation and communication. World Soil Resources Reports 103. Rome, Italy. 128 p.

Ferrero, A., Lipiec, J., Turski, M. & Nosalewics, A. 2007. Stability and Sorptivity of soil aggregates in grassed and cultivated sloping vineyards. Polish Journal of Soil Science 40: 1-8.

FitzPatrick, E.A. 1984. Micromorphology of soils. Chapman & Hall, London, UK. 433 p.

Franco, C.M.M., Tate, M.E. & Oades, J.M. 1995. Studies on non-wetting sands. I. The role of intrinsic Particulate organic matter in the development of water-repellency in non-wetting sands. Australian Journal of Soil Research 33: 253-263.

Frosterus, B. 1922. Suomen maaperä (in Finnish). Geologinen komissioni. Geoteknisiä tiedonantoja, 34, 37 p.

Groenevelt, P.H. & Grant, C.D. 2001. Re-evaluation of the structural properties of some British swelling soils. European Journal of Soil Science 52: 469–477.

Groenevelt, P.H., Kay, B.D. & Grant, C.D. 1984. Physical assessment of a soil with respect to rooting potential. Geoderma 34: 101-114.

Gusli, S., Cass, A., Macleod, D.A. & Blackwell, P.S. 1994. Structural collapse and strength of some Australian soils in relation to hardsetting: Structural collapse on wetting and draining. European Journal of Soil Science 45: 15-21.

Haines, W.B. 1923. The volume changes associated with variations of water content in soil. The Journal of Agricultural Science 13: 296-310.

Hallett, P.D., Baumgartl, T. & Young, I.M. 2001. Subcritical water repellency of aggregates from a range of soil management practices. Soil Science Society of America Journal 65: 184-190.

Hallett, P.D. & Young, I.M. 1999. Changes to water repellence of soil aggregates caused by substrate-induced microbial activity. European Journal of Soil Science 50: 35-40.

Harper, R.J., Mckissock, I., Gilkes, R.J., Carter, D.J. & Blackwell, P.S. 2000. A multivariate framework for interpreting the effects of soil properties, soil management and landuse on water repellency. Journal of Hydrology 231-232: 371-383.

Heikinheimo, M. & Fougstedt, B. 1992. Statistic of soil temperature in Finland 1971-1990. Meteorological publications No 22. Finnish Meteorological Institute. Helsinki. 75 p.

Heinonen, R. 1951. Maanrakenteesta, kirjallisuuskatsaus. English summary: On the structure of the soil, review of the literature. Agrogeologisia julkaisuja 57, 32 p.

Heinonen, R. & Pukkala, H. 1954. Seasonal variation in soil structure. The Journal of the Scientific Agricultural Society of Finland 26: 116-119.

Hintikka, S., Koivusalo, H., Paasonen-Kivekäs, M., Nuutinen, V. & Alakukku, L. 2008. Role of macroporosity in runoff generation on sloping subsurface drained clay field – a case study with MACRO-model. Hydrology Research 39: 143-155.

Hoffmann, C.H., Kjaergaard., Uusi-Kämppä, J., Hansen, H.C.B. & Kronvang., B. 2009. Phosphorus retention in riparian buffers: review of their efficiency. Journal of Environmental Quality 38: 1942-1955.

Horn, R. & Smucker, A. 2005. Structure formation and its consequences for gas and water transport in unsaturated arable and forest soils. Soil & Tillage Research 82: 5-14.

Horn, R., Taubner, H., Wuttke, M. & Baumgartl, T. 1994. Soil physical properties related to soil structure. Soil & Tillage Research 30: 187-216.

Horn, R., Vossbrink, J. & Becker, S. 2004. Modern forestry vehicles and their impacts on soil physical properties. Soil & Tillage Research 79: 207-219.

Jaramillo, D.F., Dekker, L.W., Ritsema, C.J. & Hendrickx, J.M.H. 2000. Occurrence of soil water repellency in arid and humid climates. Journal of Hydrology 231-232: 105-111.

Jarvis, N.J. 2007. A review of non-equilibrium water flow and solute transport in soil macropores: principles, controlling factors and consequences for water quality. European Journal of Soil Science 58: 523-546.

Jarvis, N., Etana, A. & Stagnitti, F. 2008. Water repellency, near-saturated infiltration and preferential solute transport in a macroporous clay soil. Geoderma 143: 223-230

Kankaanranta, J. 1996. Veden virtaus ja ravinteiden huuhtoutuminen savimaassa (In Finnish). Diploma thesis. Helsinki University of Technology, Finland, 79 pp.

Kay, B.D., Grant, C.D. & Groenevelt, P.H., 1985. Significance of ground freezing on soil bulk density under zero tillage. Soil Science Society of America Journal 49: 973-978.

Keso, L. 1930. Kulttuuriteknisiä maaperätutkimuksia erikoisesti ojaetäisyyttä silmälläpitäen. I osa. Viljelyksellisesti tärkeät maalajimme ojaetäisyyksien määräämisperusteet (In Finnish). Doctoral thesis, Suomen teknillinen korkeakoulu, Valtioneuvoston kirjapaino Helsinki, Finland. 269 p.

Kivisaari, S. 1979. Effects of moisture and freezing on some physical properties of clay soils from plough layer. Journal of the Scientific Agricultural Society of Finland 51: 245-326.

Klute, A. & Dirksen, C, 1986. Hydraulic conductivity and diffusivity: Laboratory methods. In: Klute, A (ed.). Methods of soil analysis. Part 1-Physical and mineralogical methods. Second Edition. American Society of Agronomy, Inc. Soil Science Society of America, Inc. Madison, Wisconsin USA, pp. 687-734.

Knisel, W.G. & Turtola, E. 2000. Gleams modell application on a heavy clay soil in Finland. Agricultural Water Management 43: 285-309.

Kodešová, R. 2009. Soil micromorphology use for modeling of a non-equilibrium water and solute movement. Plant, Soil and Environment 55: 424-428.

Kubiena, W.L., 1938. Micropedology. Collegiate Press, Ames, USA. 242 p.

Köhne, J.M., Köhne, S. & Šimůnek, J. 2009. A review of model applications for structured soils: a) Water flow and tracer transport. Journal of Contaminant Hydrology 104: 4-35.

Leeds-Harrison, P.B., Youngs, E.G. & Uddin, B. 1994. A device for determining the sorptivity of soil aggregates. European Journal of Soil Science 45: 269-272.

Lipiec, J., Kuś, J., Nosalewicz, A. & Turski, M. 2006. Tillage system effects on stability and sorptivity of soil aggregates. International Agrophysics 20: 189-193.

Ma´shum, M. & Farmer, V.C. 1985. Origin and assessment of water repellency of a sandy South Australian soil. Australian Journal of Soil Research 23: 623-626.

Messing, I. & Jarvis, N.J. 1990. Seasonal variation in field-saturated hydraulic conductivity in two swelling clay soils in Sweden. Journal of Soil Science 41: 229-237.

Murphy, C.P., Bullock, P. & Turner, R.H. 1977a. The measurement and characterization of voids in soil thin sections by image analysis. Part I. Principles and techniques. Journal of Soil Science 28: 498-508.

Murphy, C.P., Bullock, P. & Biswell, K.J. 1977b. The measurement and characterization of voids in soil thin sections by image analysis. Part II. Applications. Journal of Soil Science 28: 509-518.

Olsen, P.A. & Haugen, L.E. 1998. A new model to the shrinkage characteristic applied to some Norwegian soils. Geoderma 83: 67-81.

Osborn, J.F., Pelishek, R.E., Krammes, J.S. & Letey, J. 1964. Soil wettability as a factor in erodibility. Soil Science Society of America Proceedings 28: 294-295.

Øygarden, L., Kværner, J. & Jenssen, P.D. 1997. Soil erosion via preferential flow to drainage system in clay soils. Geoderma 76: 65-68.

Pagliai, M., La Marca, M. & Lucamante, G. 1983. Micromorphometric and micromorphological investigations of a clay loam soil in viticulture under zero and conventional tillage. Journal of Soil Science 34: 391-403.

Pagliai, M., La Marca, M, Lucamante, G. & Genovese, L. 1984. Effects of zero and conventional tillage on the length and irregularity of elongated pores in a clay loam soil under viticulture. Soil & Tillage Research 4: 433-444.

Pagliai, M., Vignozzi, N. & Pellegrini, S. 2004. Soil structure and effects of management practices. Soil & Tillage Research 79: 131-143.

Pagliai, M. & Vittori Antisari, L. 1993. Influence of waste organic matter on soil micro- and macrostructure. Bioresource Technology 43: 205-213.

Peltovuori, T., Uusitalo, R. & Kauppila, T. 2002. Phosphorus reserves and apparent phosphorus saturation in four weakly developed cultivated pedons. Geoderma 110: 35-47.

Peng, X. & Horn, R. 2005. Modeling soil shrinkage curve across a wide range of soil types. Soil Science Society of America Journal 69: 584-592.

Peng, X. & Horn, R. 2007. Anisotropic shrinkage and swelling of some organic and inorganic soils. European Journal of Soil Science 58: 98-107

Peng, X., Horn, R., Deery, D., Kirkham, M.B. & Blackwell, J. 2005. Influence of soil structure on the shrinkage behaviour of a soil irrigated with saline–sodic water. Australian Journal of Soil Research 43: 555-563.

Peng, X., Horn, R., Peth, S. & Smucker, A. 2006. Quantification of soil shrinkage in 2D by digital image processing of soil surface. Soil & Tillage Research 91: 173-180.

Peng, X., Horn, R. & Smucker, A. 2007. Pore shrinkage dependency of inorganic and organic soils on wetting drying cycles. Soil Science Society of America Journal 71: 1095-1104.

Philip, J.R. 1957. The theory of infiltration: 4. Sorptivity and algebraic infiltration equations. Soil Science 84: 257-264.

Pietola, L., Horn, R. & Yli-Halla, M. 2005. Effects of trampling by cattle on the hydraulic and mechanical properties of soil. Soil & Tillage Research 82: 99-108.

Puustinen, M., Koskiaho, J. & Peltonen, K. 2005. Influence of cultivation methods on suspended solids and phosphorus concentrations in surface runoff on clayey sloped fields in boreal climate. Agriculture, Ecosystems and Environment 105: 565-579.

Rasmussen, K.J. 1999. Impact of ploughless soil tillage on yield and soil quality: A Scandinavian review. Soil & Tillage Research 53: 3-14.

Regina, K. & Alakukku, L. 2010. Greenhouse gas fluxes in varying soils under conventional and no-tillage practices. Soil & Tillage Research 109: 144-152.

Reeve, M.J., Hall, D.G.M. & Bullock, P. 1980. The effect of soil composition and environmental factors on the shrinkage of some clayey British soils. Journal of Soil Science 31: 429-442.

Ringrose-Voase, A.J. 1991. Micromorphology of soil structure: description, quantification, application. Australian Journal of Soil Research 29: 777-813.

Ritsema, C.J. 1998. Flow and transport in water repellent sandy soils. Doctoral thesis, Wageningen Agricultural University, the Netherlands. 215 p.

Ross, G.J. 1978. Relationships of specific surface area and clay content to shrink-swell potential of soils having different clay mineralogical compositions. Canadian Journal of Soil Science 58: 159-166.

Räty, M., Horn, R., Rasa, K., Yli-Halla, M. & Pietola, L. 2010a. Compressive behaviour of the soil in buffer zones under different management practices in Finland. Agricultural and Food Science 19: 160-172.

Räty, M., Uusi-Kämppä, J., Yli-Halla, M., Rasa, K. & Pietola, L. 2010b. Phosphorus and nitrogen cycles in the vegetation of differently managed buffer zones. Nutrient Cycling in Agroecosystems 86: 121-132.

Samouëlian, A., Righard, G., Cousin, I., guérin, R., Bruand, A. & Tabbagh, A. 2004. Three-dimensional crack monitoring by electrical resistivity measurement. European Journal of Soil Science 55: 751-762.

Servadio, P., Marsili, A., Vignozzi, S., Pellegrini, M. & Pagliai, M., 2005., Effects of soil qualities in central Italy following the passage of four wheel drive tractor fitted with single and dual tires. Soil & Tillage Research 84: 87-100.

Shakesby, R.A., Doerr, S.H. & Walsh, R.P.D. 2000. The erosional impact of soil hydrophobicity: current problems and future research direction. Journal of Hydrology 231-232: 178-191.

Shipitalo, M.J., Nuutinen, V. & Butt, K.R. 2004. Interaction of earthworm burrows and cracks in a clayey, subsurface-drained, soil. Applied Soil Ecology 26: 209-217.

Sippola, J. 1974. Mineral composition and its relation to texture and to some chemical properties in Finnish subsoils. Annales Agriculturae Fenniae 13: 169-234.

Sippola, J. & Tares, T. 1978. The soluble content of mineral elements in cultivated Finnish soils. Acta Agriculturæ Scandinavica, Supplementum 20: 11-25.

Soil Survey Staff 2010. Keys to Soil Taxonomy. 11th Ed. United States Department of Agriculture, Natural Resources Conservation Service. 338 pp.

Stirk, G.B. 1954. Some aspect of soil shrinkage and the effect of cracking upon water entry into the soil. Australian Journal of Agricultural Research 5: 279 -290.

Stoops, G. 2003. Guidelines for analysis and description of soil and regolith thin sections. Soil Science Society of America, Inc. Madison, USA. 184 p.

Sveistrup, T.E., Haraldsen, T.K., Langohr, R., Marcelino, V. & Kvaerner, J. 2005 Impact of land use and seasonal freezing on morphological and physical properties of silty Norwegian soils. Soil & Tillage Research 81: 39-56.

Syversen, N. 2005. Effect and design of buffer zones in the Nordic climate: the influence of width, amount of surface runoff, seasonal variation and vegetation type on retention efficiency for nutrient and particle runoff. Ecological Engineering 24: 483-490.

Tempany, H.A. 1917. The shrinkage of soils. The Journal of Agricultural Science 8: 312-330.

Tillman, R.W., Scotter, D.R., Wallis, M.G. & Clothier, B.E. 1989. Water-repellency and its measurement by using intrinsic sorptivity. Australian Journal of Soil Research 27: 637-644.

Tippkötter, R. & Ritz, K. 1996. Evaluation of polyester, epoxy and acrylic resins for suitability in preparation of soil thin sections for in situ biological studies. Geoderma 69: 31-57.

Tippkötter, R., Ritz, K. & Darbyshire, J.F. 1986. The preparation of soil thin-sections for biological studies. Journal of Soil Science 37: 681-690.

Turtola, E. & Paajanen, A. 1995. Influence of improved subsurface drainage on phosphorus losses and nitrogen leaching from a heavy clay soil. Agricultural Water Management 28: 295-310.

Turtola, E., Alakukku, L., Uusitalo, R. & Kaseva, A., 2007. Surface runoff, subsurface drainflow and soil erosion as affected by tillage in a clayey Finnish soil. Agricultural and Food Science 16, 332-351.

Ulén, B., Bechmann, M., Fölster, J., Jarvie, H.P. & Tunney, H. 2007. Agriculture as a phosphorus source for eutrophication in the north-west European countries, Norway, Sweden, United Kingdom and Ireland: a review. Soil Use and Management 23: 5-15.

Ulén, B. & Snäll, S. 2007. Forms and retention of phosphorus in an illite-clay soil profile with a history of fertilisation with pig manure and mineral fertilisers. Geoderma 137: 455-465.

Uusi-Kämppä, J., 2005. Phosphorus purification in buffer zones in cold climates. Ecological Engineering 24: 491-502.

Uusi-Kämppä, J. & Jauhiainen, L. 2010. Long-term monitoring of buffer zone efficiency under different cultivation techniques in boreal conditions. Agriculture, Ecosystems and Environment 137: 75-85.

Uusitalo, R., Turtola, E., Grönroos, J., Kivistö, J., Mäntylahti, V., Turtola, A., Lemola, R. & Salo, T. 2007. Finnish trends in phosphorus balances and soil test phosphorus. Agricultural and Food Science 16: 301-316.

Uusitalo, R, Turtola, E., Kauppila, T. & Lilja, T. 2001. Particulate phosphorus and sediment in surface runoff and drainflow from clayey soils. Journal of Environmental Quality 30: 589-595.

VandenBygaart., A.J., Fox, C.A., Fallow, D.J. & Protz, R. 2000. Estimating earthworm-influenced soil structure by morphometric image analysis. Soil Science Society of America Journal, 4, 982-988.

VandenBygaart, A.J., Protz, R. & Tomlin A.D. 1999. Changes in pore structure in a no-till chronosequense of silt loam soils, Southern Ontario. Canadian Journal of Soil Science 79: 149-160.

Viljavuuspalvelu, 2008. http://www.tuloslaari.fi/. Visited 2007.

Warsta, L. 2007. Modelling overland and subsurface drainage runoffs at an agricultural field. A Licenciate thesis, Helsinki University of Technology, Finland, 70 p.

White, I. & Sully, M.J. 1987. Macroscopic and microscopic capillarity length and time scales from field infiltration. Water Resources Research 23: 1514-1522.

Yli-Halla, M. & Mokma, D.L. 1998. Soil temperature regimes in Finland. Agricultural and Food Science in Finland 7: 507-512.

Yli-Halla, M. & Mokma, D.L. 2001. Soils in an agricultural landscape of Jokioinen, south-western Finland. Agriculture and Food Science 10: 33-43.

Yli-Halla, M., Mokma, D.L., Alakukku, L., Drees, R. & Wilding, L.P. 2009 Evidence for the formation of Luvisols/Alfisols as a response to coupled pedogenic and anthropogenic influences in a clay soil in Finland. Agriculture and Food Science 18: 388-401.

Youngs, E.G. & Leeds-Harrison, P.B. 1990. Aspects of transport processes in aggregated soils. Journal of Soil Science 41: 665-675.

Youngs, E.G., Leeds-Harrison, P.B. & Garnett, R.S. 1994. Water uptake by aggregates. European Journal of Soil Science 45: 127-134.

Zhang, M. & Karathanasis, A.D. 1997. Characterization of iron-manganese concretions in Kentucky Alfisols with perched water tables. Clays and Clay Minerals 45: 428-439.

MTT is publishing its research findings in two series of publications:
MTT Science and MTT Growth.

MTT CREATES VITALITY THROUGH SCIENCE

MTT SCIENCE

www.mtt.fi/julkaisut

The MTT Science series includes scientific presentations and abstracts from conferences arranged by MTT Agrifood Research Finland.
Doctoral dissertations by MTT research scientists will also be published in this series.
The topics range from agricultural and food research to environmental research in the field of agriculture.

MTT, FI-31600 Jokioinen, Finland.
Tel. +358 3 41881, email julkaisut@mtt.fi