PERSISTENCE OF SUBSOILING EFFECTS ON THE SOIL PHYSICAL AND HYDRAULIC PROPERTIES IN A RECONSTRUCTED SOIL

by

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ABSTRACT

Surface mining is one of the most significant forms of anthropogenic disturbance to natural and managed ecosystems. In Alberta, mining disturbs large areas in the Mixedwood Boreal natural region and recovery is often slow because of poor soil quality, specifically the high degree of compaction. Soil compaction, caused by repeated traffic of heavy machinery during soil reconstruction of surface mined lands, hinders the re-establishment of vegetation. Compaction causes changes to soil physical properties such as increased bulk density and reduced macroporosity which reduce soil infiltration capacity, drainage and water holding capacity. The disruption of the soil water balance as a result of these compaction-induced changes to the soil further negatively effects the chemical and biological functioning in the soil because of poor aeration. In compacted forest soils, subsoiling with heavy-duty rip ploughs has been shown to be an effective method at ameliorating compaction by breaking up large compacted layers, into smaller aggregates and peds which significantly increases macroporosity, infiltration, drainage and aeration. The main objective of this research is to quantify any medium-term (~ 4 yrs) benefits of subsoiling with a heavy-duty rip plough on reconstructed soil at a coal mine. In 2010, an experimental research site was established at the Genesee Prairie Mine, 70 km west of Edmonton to investigate the potential for compaction amelioration using a McNabb winged subsoiler D7R XR to a 60 cm depth. Results showed that medium-term effects of ripping are variable with depth. Ripping effects on pore size distribution, saturated hydraulic conductivity, and bulk density were most pronounced in the 15-20 cm depth. Infiltration rates were increased by ripping which is expected to reduce hydraulic barriers at the soil surface. Evidence suggests non-ripped surface layers (5-10 cm depth) showed improvements in soil properties (bulk density, saturated hydraulic conductivity) as a result of natural processes (i.e., plant root expansion and

drying-shrinkage). Over time, it appears that the effects of ripping have decreased, with a simultaneous improvement in non-ripped soils.

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TABLE OF CONTENTS

LIST OF TABLES

Chapter 2:

Chapter 3:

LIST OF FIGURES

Chapter 2:

Chapter 3:

1. CHAPTER 1 - GENERAL INTRODUCTION

1.1. BACKGROUND

Coal is the most plentiful fossil fuel worldwide (Alberta Energy 2015). Coal deposits have formed over extensive periods of time from plant-based residues under extreme heat and pressure. Coal is a non-renewable resource that is mined in two forms in Alberta: bituminous and sub-bituminous. These forms contain twice as much energy as other sources (natural gas, conventional oil and pentanes, bitumen, and synthetic crude) found in the province. Coal is primarily used for electricity generation but can be processed and used in other forms (Alberta Energy 2015).

The coal mining industry is fundamental to the development of Alberta's economy. Approximately 1800 mines have operated since coal mining was first established in 1882 in Lethbridge (Alberta Energy 2014). About 70 % of the coal reserves in Canada are located in Alberta and this equates to approximately 300,000 km^2 or 48 % of Alberta being underlain by coal bearing formations (AESRD 2014). As of 2010, an estimated 31,000 hectares has been disturbed for coal mining exploration in Alberta (AESRD 2012) and, as of 2013, approximately 33.6 billion tonnes of coal reserves remain (Alberta Energy 2014).

Two primary methods of coal extraction are surface mining and in-situ mining. The majority of coal mining in Alberta is done by surface methods since coal deposits are found relatively close to the surface. Surface methods include open-pit mining and strip mining. Open-pit mining involves the removal of overburden materials and construction of a large pit or burrow for coal extraction. Strip mining is the process in which a seam of coal is exposed in a strip. Once the coal is exhausted the overburden material is repositioned and a new adjacent strip is exposed (Alberta Energy 2015).

Coal mining is one of the most significant forms of anthropogenic disturbances, causing degradation to terrestrial ecosystems (Shrestha and Lal 2011). Significant alterations to ecosystem functions and services occur from mining activities such as removal of vegetation, changes to topography and drainage, and degradation of soil structural properties. Reclamation of mined land to an approved end land use is required by law in Alberta. Reclamation improves land value and quality for human use, and facilitates recovery of ecosystem services and functions.

1.2. NATURAL REGIONS OF ALBERTA

Six natural eco-regions exist in Alberta on the basis of climate, topography, vegetation, soil and physiographic features. It is useful to delineate these regions as they strongly influence land use and management practices. Coal mining exploration occurs in almost all regions within Alberta. The Genesee Prairie Mine, located 70 km west of Edmonton, is located within the Dry Mixedwood Natural Subregion of the Boreal Forest Natural Region and the Central Parkland Natural Subregion of the Parkland Natural Region (Navus Environmental 2010).

1.2.1. Dry Mixedwood Natural Subregion

The Boreal Forest Natural Region is the largest natural region in Alberta with a land base of approximately 381,046 km^2 (58 % of Alberta) (Natural Regions Committee 2006). The Dry Mixedwood Natural Subregion is the largest subregion occupying 22% of the Boreal Forest Natural Region. Climate is characterized as warmer summers and milder winters than other natural subregions within the Boreal Forest. Temperatures range from an average 15.9° Celsius in the summer to -16.9° Celsius in the winter, with a mean annual temperature of 1.1° Celsius. About 70 % of the annual precipitation (461 mm) occurs in the growing season.

Dominant vegetation types are aspen forests and cultivated landscapes, with interspersed fens. The modal communities are aspen dominated. In the southern area, aspen with understory species of beaked hazelnut, prickly rose, wild sarsaparilla, cream-colored vetchling, purple peavine, and bluejoint develop on well drained Gray and Dark Gray Luvisols. Northern aspen areas have understories of low bush cranberry (*Vaccinium vitis-idaea L.*), rose (*Rosa woodsia*), Canada buffaloberry (*Shepherdia canadensis*), hairy wild rye (*Elymus villosus*) and bunchberry (*Cornus canadensis*). These occur in areas with imperfect or restricted drainage associated with Gray Luvisols and gleyed subgroups. In addition, balsam poplar, aspen and white spruce grow in moist, rich sites; fens and bogs in poorly drained areas; and jack pine stands grow in the driest sites. Parent material is dominantly lacustrine with undulating plains and hummocky uplands.

Approximately 50% of this sub-region has been cultivated. Forestry operations utilize aspen for paper and pulp production. Oil and Gas operations are heavily focused in the Cold Lake, Peace River, and coal mining around the Wabamun Lake area. Recreational activities such as hunting and fishing are popular in this sub-region.

1.2.2. Central Parkland Natural Subregion

The Parkland Natural Region comprises approximately 9 % of Alberta. Within this region, the Central Parkland Natural Subregion is dominate and occupies 53,706 km^2 (88 %). Bordering the Central Parkland to the west and north is the Dry Mixedwood Natural Subregion, and the Foothills Fescue, Foothills Parkland and Northern Fescue Natural Subregions to the south (Natural Regions Committee 2006). Summer precipitation is adequate with warm, long growing seasons capable of supporting forest ecosystems. Mean annual temperature is 2.3° Celsius, with the average summer temperature of 16.5° Celsius and average winter temperatures of -14.7° Celsius. Mean annual precipitation is 441 mm, with approximately 75 % occurring during the growing season.

Approximately only 5 % of the sub-region is occupied by native vegetation due to the conversion of forests to croplands. Native vegetation is dominated by a northern aspen portion and a southern grassland portion. Aspen understories are variable but include Saskatoon, prickly rose, beaked hazelnut and forbs and grasses (Natural Regions Committee 2006). Soils in these areas are imperfectly drained Dark Gray Luvisols and Dark Gray Chernozems. Grasslands are dominated by plains rough fescue and are found on well-drained Black Chernozems. Wetlands develop in depressions with poorly drained Gleysolic soils. Landscapes are mainly glacial till interspersed with lacustrine, fluvial and eolian deposits that form the undulating plains and hummocky uplands.

Land use comprises urban development, agriculture and industrial exploration and development. Urban centers such as Edmonton, Red Deer and Calgary are located within this sub-region. Agriculture is predominant and occupies 80 % of the plains and 65 % of the upland areas (Natural Regions Committee 2006). Agricultural and industrial activities are the primary causes of loss to native vegetation. Invasive species and removal of existing vegetation threaten the natural plant communities found in this sub-region.

1.3. RECLAMATION PRACTICES

Land is a limited resource, and reclamation is therefore essential to restore an exhausted coal mine back to a beneficial use. Reclamation aims to enhance site productivity, stability and sustainability by creating a system that reflects the adjacent landscape (Mukhopadhyay et al. 2013). Mining activities and reclamation are regulated by Alberta Environment and Sustainable Resource Development and the Alberta Energy Regulator (Alberta Energy 2015).

The Environmental Protection and Enhancement Act (EPEA), mandates land must be reclaimed to a specified land use - a reclamation certificate is issued if the reclamation efforts are successful (Province of Alberta 2000). Land capability should be equivalent to or better than to pre-disturbance conditions even if the intended land use is different. As of 2012, approximately 15,500 hectares had been temporarily or permanently reclaimed and 2,200 hectares had been certified (AESRD 2012).

Soil reconstruction on mined sites is site-specific and varies based on the pre-disturbance soil conditions. Topsoil and subsoil should be salvaged and stockpiled separately prior to mining activities (Alberta Environment 2005). Implementation of erosion control is required until vegetative cover is established by slope stabilization, reseeding, cross ditching and soil replacement (Alberta Environment 2005). Following the decommissioning of the mine, overburden materials are used to backfill the pit or strip. Salvaged materials are replaced on the landscape and graded and contoured to match the surrounding areas (Alberta Environment 2005). Woody debris, rocks, stones and other debris are removed as they may interfere with the intended land uses.

A major component of the reclamation process is a completed Environmental Site Assessment which identifies and remediates any contamination. Three Environmental Site Assessment Phases complete the due diligence obligations of the mining companies to fully reclaim and manage for future risks associated with the disturbed landscape. Phase I gathers information on the likelihood of a present contamination and determines if a Phase II is required. Phase I assessments are non-invasive assessments that identify potential spill locations, associated with previous infrastructure used for waste storage or disposal such as storage tanks, and flare pits (Alberta Environment 2011). A Phase II assessment is conducted if contamination has been identified or there is a high probability of contamination. Intrusive soil assessments are completed to delineate the horizontal and vertical dimensions of any contamination. The third phase is the remediation or removal and replacement of contaminated materials. Remediated soil must comply with Alberta Tier 1 or Tier 2 guidelines to maintain contaminate concentrations below guidelines to protect human and ecological receptors. Guidelines used must ensure that the most sensitive receptors are protected.

Following remediation, vegetation cover is established to quickly stabilize landscapes and prevent excessive erosion. Industry may be encouraged to promote rapid vegetation cover in the short term to meet reclamation objectives (Holl 2002). Agronomic species are utilized in reclamation for their ability to quickly establish, provide protective cover for the soil and increase the landscape productivity (Pollster 2000; Powter et al. 2012). However, these species are not representative of the natural plant communities found in the natural regions of the coal mining areas. More recently, reclamation practices have shifted to strive to re-establish forest ecosystems on disturbed landscapes. Practices to enhance forest reclamation will increase ecosystem functioning and services provided.

1.4. SOIL GENESIS

In terms of pedogenesis, human activity is a potentially catastrophic event that returns the soil to a new “time zero” (Richter 2007). Reconstruction of a functioning soil with suitable aeration and water holding capacity is required for establishment of vegetation and successful land reclamation. Soil forming factors that influence soil genesis and horizon development after mining are: climate, organisms, relief, parent material, time and beneficial human interference (i.e., soil reconstruction or anthropogenic parent material selection and deposition). Reconstruction changes the original placement and organization of soil and overburden materials which can be viewed as a parent material deposition. Alterations to this new material occur through deposition of organic residues, weathering of primary or secondary minerals, and solute and ion leaching.

Climate and vegetation determine the water supply (precipitation) and demand (potential evapotranspiration) and are key drivers of the soil water balance. Topography and soil parent material are secondary influences on the soil water balance and affect the partitioning of precipitation into overland flow or infiltration/internal drainage and soil water holding capacity. The soil-plant-atmosphere continuum links the soil hydrologic properties (hydraulic conductivity) and hydrological processes (infiltration, percolation, evapotranspiration and drainage) within and between soils, plants and the atmosphere (Hillel 1998). As genesis continues, soil physical properties (soil structure and the pore system) that govern soil hydraulic function and the soil water balance develop by the simultaneous and interdependent interaction of the soil forming factors.

1.5. SOIL PHYSICAL PROPERTIES

Soil texture is based on the relative proportions and sizes of particles in the soil. Texture can be divided into three primary subgroups: sand, silt and clay. Sand is classified as particles with diameters between 0.05 mm and 2.0 mm, silt particle diameters are 0.002 and 0.05 mm and clay particle have diameters < 0.002 mm. The strength and amount of particle aggregation is influenced by soil texture and organic matter type and accumulation. Coarse textured soils are characterized by low aggregation, with large void spaces that cause water contents to be relatively low (Naeth et al. 1991). Fine textured soils have a higher degree of aggregation, small but many void spaces with high water holding capacities (Naeth et al. 1991).

The arrangement and geometry of soil particles defines the soil structure. Soil structure can be grouped into three categories: single grained, aggregated, and massive. Single grained structures arise when soil particles are completely unattached in a loose unconsolidated manner. A consolidated structure is defined as massive when individual particles are tightly packed and do not break apart easily. Aggregated structures are the most favourable for plant growth and develop through the clustering of individual particles into defined aggregates or peds (Hillel 1998). Coarse-textured soils have a granular structure and soils dominated by clay develop aggregated or massive structures (Naeth et al. 1991).

Aggregates or peds are the primary building blocks of aggregated soil structures and develop as particles become more clustered or grouped together. Soil organic carbon, soil biota, and ionic bonding promote aggregation by the cementation and flocculation of soil particles (Bronick and Lal 2005). Organic exudates act as biological glues that form soil aggregates. Climate and vegetation have an indirect role on aggregation. Shrinking and swelling by freeze-thaw, wet-dry cycles can have pronounced stress-strain effects on the formation and destruction of aggregates.

Aggregates can be classified as macro-aggregates (> 250 um) and micro-aggregates (< 250 um) (Bronick and Lal 2005). The hierarchal organization of soil aggregates indicates that macro-aggregates develop from arrays of more densely packed micro-aggregates. Smaller aggregates have a greater mass to volume ratio and bulk density in comparison to larger aggregates. The volume of spaces between and within the macro- and micro- aggregates determine the soil porosity and pore size distribution. Pores can be divided into inter- and intra- aggregate pores, where they occur between or within aggregates, respectively. Inter-aggregate pores are commonly macro-pore size and intra-aggregate pores are commonly micro-pore size.

Pore spaces and connectivity are important for gas and water transport, and water retention properties in the soil. Pore diameters can be simplified to cracks and/or bio-pores (> 500 μm), macro-pores (100 to 500 μm), meso-pores (50 to 100 μm) and micro-pores/storage pores (0.2 to 50 μm) and residual pores (< 0.2 μm) (Kutilek 2004; Pagliai et al. 2004). Meso- and macro-pores can be further classified as structural and/or transmission pores (50 to 500 μm). Macro-pores are usually inter-aggregate, inter-connected pores and are responsible for the rapid flow of water and air. At saturation, the adsorptive forces between water molecules and soil particles are low (Hillel 1980). When the soil dries, these pores restrict water movement to capillary flow in thin films along the pore edges (Hillel 1998). Meso- and micro- pores have a small contribution to water flow at saturation. Water is held more tightly to the pore surfaces by adsorptive and capillary forces. Unsaturated flow in soils is dictated by the proportion of meso- and micro-pores. These pores will retain water at low potentials and contribute to the movement of water as thin films that flow by capillary forces. Soils with a higher proportion of meso- and micro- pores will participate in water flow in unsaturated conditions.

Soil water holding capacity at saturation is governed by the proportion of macro-pores. Soils with a greater proportion of macro-pores typically have greater volumetric water content at saturation. Macro-pores have low adsorption with water causing the force of gravity to drain these pores first as the soil dries. Finer pores retain water in thin films on the pore edges. These pores typically remain saturated even when the soil dries. Residual water develops within micro-pores that do not form interconnected networks (Hillel 1998) and the occluded micro-pores obstruct the passage of water between pores. When the soil reaches potentials less than -15,000 hPa, the remaining residual water creates a hydration envelope around the soil particles (Startsev and McNabb 2001). Water within these pores is typically unavailable for plant extraction.

1.6. THE HYDROLOGIC CYCLE AND THE SOIL-PLANT-ATMOSPHERE CONTINUUM

Water is continually cycling from the atmosphere to the earth's surface through what is known as the hydrologic cycle. The fate of precipitation depends on whether it is intercepted or reaches the soil surface where it can infiltrate into the soil, redistribute and eventually drain into the groundwater. If the infiltration capacity of the soil is exceeded by the precipitation rate, water can flow over the soil surface as runoff or be evaporated from the bare soil.

Water that reaches the soil surface can infiltrate into the soil or become runoff. The maximum rate of soil absorption of precipitation is termed the soil infiltration capacity. Infiltration is controlled by the storage capacity and transmission of water in the profile. Soils with a higher saturated hydraulic conductivity generally have greater infiltration capacity. Sandy and well-granulated soils have larger pores and allow for a high proportion of water to infiltrate compared to dense clayey soils (Brady and Weil 2002). Macro-pores can have a significant effect on the rate of infiltration depending on their size, distribution and connectivity. Guebert and Gardner (2001), stated that macro-pores increase the volume of water stored at the soil surface and increase the connectivity of pores to greater depths, thereby reducing surface runoff. Infiltration into the profile is maintained as water redistributes from macro-pores at the surface to finer pores deeper in the soil matrix. Further, cracks or fissures that develop can act as preferential pathways in which water rapidly infiltrates and extends to some depths in the profile (Hillel 1998). These typically develop in clayey soils that are prone to shrink-swell cycles from drying and wetting.

When the precipitation rate exceeds the infiltration capacity of the soil, water ponding occurs and the infiltration becomes "surface limiting". Areas with topographic relief can experience surface runoff when the ponded water flows over the soil surface. Timing of precipitation can greatly influence the level of runoff. Periods of intense rainfall will have higher runoff than periods of low intensity spread over a larger time period.

Water that remains ponded at the soil surface is exposed to evaporation. Climatic factors (temperature, wind, relative humidity) control the potential evaporation that can occur from the bare soil. High rates of evaporation can cause the soil to dry, contract and develop vertical, desiccation fissures or cracks. Large cracks in the soil can further promote evaporation from deeper in the profile and increase the infiltrability of the soil during precipitation events (Hillel 1998).

Water that infiltrates into the soil redistributes horizontally and vertically within the profile. Redistribution of soil water is governed by soil water potential gradients. Water percolates within the profile according to hydraulic potential gradients - from areas of high potential to low potential. Hydraulic potential at any given location in the soil is the sum of gravitational, matric and osmotic potentials which are defined with respect to a reference state. As water redistributes, the gradient decreases and the water begins to reach a state of equilibrium at which water flow ceases (Hillel 1998). Rates of redistribution are influenced by the magnitude of the gradient and the hydraulic conductivity of the soil. Macro-pores at the surface can cause water to redistribute rapidly to greater depths in the profile, bypassing the majority of the soil matrix through a process called channeling (Guebert and Gardner 2001). Channeling can cause low storage in the root zone of the profile if the macropores extend below the rootzone. Poor water storage is detrimental to plant growth.

Field capacity is used to describe the state of soil water when all “free” water has drained from the macropores by gravitational forces. Soil texture influences the rate of redistribution. In coarse textured soils field capacity is reached quickly because once large pores have drained, the hydraulic conductivity decreases by many orders of magnitude. Fine to medium textured soils require more time to reach field capacity because they generally have wider pore size distributions with percolating water moving slowly from the rooting zone, taking an appreciable amount of time to reach field capacity (many days).

At field capacity, water is held in “matrix pores” by capillary forces and is available for plant uptake within the root zone. Whereas infiltration is dominated by gravitational gradients, redistribution at field capacity is a “competition” between downward gravitational gradients and upward gradients created between soil-plant root interface and the atmosphere. Water taken up by plants, returns to the atmosphere as water vapor through the process of transpiration. The soil-plant-atmosphere continuum is the pathway for root zone soil water to return to the atmosphere through the processes of evaporation and transpiration. Plant roots exhibit a strong influence on the development of pores with emphasis on macro-pores. Bio-pores are macro-pores that form by biological activity such as root expansion. Processes of water transport and storage can be significantly influenced by the presence of bio-pores that permit rapid flow of water through the profile (Hillel 1998).

Water flows from the soil to the plant roots, up the stems and to leaves along a gradient of higher potentials to lower potentials. Transpiration then occurs where water is lost through the stomates to the atmosphere. The entire loss of water from the earth's surface includes (because it is often difficult to separate) water loss from the soil and vegetation termed evapotranspiration. Evapotranspiration is controlled by the vapour pressure or relative humidity of the atmosphere and the diffusion water deficit of the plant (Hillel 1998). Plant roots draw water from the soil ultimately affects water redistribution and reducing water storage.

Vegetation intercepts precipitation, reduces surface runoff and can increase infiltration. Plant roots can open up the soil and increase the permeability of water to the root zone. Internal drainage is considered in the root zone but as water flows further in depth it is termed deep percolation. Plant roots are not able to uptake water from this zone and it drains to the groundwater. However, evapotranspiration can occur within the soil profile and not just at the surface. Plant roots increase the depth to which water can be withdrawn from the profile via the atmosphere (Brady and Weil 2002).

Some water may percolate below the root zone to become deep drainage, after which it can reach the groundwater table or an aquifer. Drainage is important for increasing aeration in the rooting zone and removing salts and toxins. Poor drainage can increase the susceptibility of the soil to salinization through capillary rise of water into the rooting zone. High rates of drainage can reduce plant available water.

1.7. SOIL COMPACTION

1.7.1. Soil structure and the pore system

Soil reconstruction aims to create a soil at the foundation of a suitable ecosystem capable of supporting the intended end land uses. However, heavy equipment and machinery used during surface mining and the subsequent handling of materials for reclamation causes soil compaction. Reclamation objectives should aim to re-establish a plant community representative of or with the equivalent capability to the natural ecosystems. Assessment of reconstructed mine soils indicates compaction prevents healthy establishment of native vegetation (Shrestha and Lal 2011). Therefore, compaction hinders the success of reclamation by impeding the re-vegetation process following soil reconstruction.

Compaction causes changes to soil physical properties such as increased bulk density and reduced macroporosity which reduce soil permeability, infiltration capacity, drainage and water holding capacity. The disruption of the soil water balance as a result of these compaction-induced changes to the soil further negatively effects the chemical and biological functioning in the soil because of poor aeration.

Compaction causes the aggregates to break apart into smaller peds or become compressed. Many reclaimed mined soils develop a weak platy granular structure following reclamation (Horn et al. 2003; Naeth et al. 1991). Eventually under repeated stress, complete homogenization of the soil can occur resulting in a massive soil structure (Horn et al. 2003). The arrangement of voids and solids following compaction will determine the total porosity, pore size distribution and continuity. (Figure 1-1)

Bulk density is the ratio of the mass of soil solids to its bulk volume (i.e., including solids and voids). Compaction causes void spaces to be compressed decreasing the distance between soil particles, increasing the bulk density. Air is the entity in soil that becomes most displaced during compaction as its expulsion from the soil matrix is rapid compared to liquids (Hillel 1980). In a moist soil, void spaces will be filled with water under a repeated stress restricting aeration and gas diffusion.

In a study by Page-Dumroese et al. (2006), compaction increased bulk density in fine textured soils by up to 31 % and 58 % at 10-20 cm and 20-30 cm depths, respectively. Porosity is inversely proportional to the bulk density. Richard et al. (2001a) argued that a subsequent increase in bulk density corresponds to decreases in structural porosity (large pores developed via tillage, traffic, weather and biological activity). A reduction of 25 % in total porosity has been observed in silt loam volcanic ash soils at 10-20 cm depths with increasing bulk density (Page-Dumroese et al. 2006).

Void ratio is a useful measure of porosity as it directly relates the volume of pore spaces to the volume of solids. A reduction of the void ratio to 0.76 corresponded to a decrease in the bulk soil by 50 percent in a decomposed granitic soil (Li and Zhang 2009). Changes in total porosity and bulk density are indicators of the severity of compaction. Void ratio may be a better indicator of compaction as it relates to the packing density of a soil. Changes in macro-pore volume are often undetected by bulk density and porosity; however, macro-pores have an important role in water storage, and water and gas permeability. Guebert and Gardener (2001) indicate measures of bulk density and porosity may not be conclusive regarding the effects of compaction to soil functionality. Air and water permeability are better indicators of compaction as it can inform the changes in pore size distribution and pore continuity (Grevers and de Jong 1992).

Macro-pores are most sensitive to applied stresses. In a study by Schaffer et al. (2007a), increasing the level of induced trafficking on a loamy soil caused a reduction in macro-porosity from 0.130 $m^3 m^{-3}$ to 0.022 $m^3 m^{-3}$. Macro-pores decreased and shifted to smaller, regular shaped micro- and meso- pores with changes in mean pore diameters from 1.07 mm to 0.85 mm (Schaffer et al. 2007a). Micro-pores may be more stable and unaffected by compaction as the particles surrounding small pores are already tightly packed and exhibit greater resistance to deformation (Page-Dumroese et al. 2006). Distances between aggregates are reduced leading to increased macro-pore isolation (Lehnard 1986; Schaffer et al. 2007a) and increases in micro- and meso-pore connectivity (Richard et al. 2001a). Relict macro-pores become accessible only through the narrow micro-pore necks.

Compaction induced changes to a soil's pore size distribution, decreases the saturated hydraulic conductivity of the soil (Hillel 1980; Barnes et al. 1971). Soils composed primarily of smaller pores have lower conductivity than those with a greater proportion of large pores (Hillel 1980). Startsev and McNabb (2001) found saturated hydraulic conductivity to decrease by a factor of two or more in a medium textured Gray Luvisol following a series of skidding traffic on experimental soil plots. A homogenized soil structure lacks a continuous network of macro-pores with an increased distribution in micro-pores (Shukla et al. 2004). As a result the tortuosity of the soil is increased, the path length and frictional forces of water moving through the pore spaces increases causing reductions in saturated hydraulic conductivity.

As the matric potential decreases as a result of soil drying, hydraulic conductivity is decreased. The unsaturated hydraulic conductivity will be lower in a compacted soil at low negative potentials (i.e., wet) compared to a natural soil. As the soil continues to dry, the hydraulic conductivity may become greater in the compacted soil because compaction causes increased contact between aggregates and, therefore, connectivity between the micro-pores within those aggregates. (Horn et al. 1995; Horn et al. 2003). With a greater proportion of connected micro-pores between aggregates, hydraulic conductivity declines more slowly (Richard et al. 2001a). Larger pores become hydraulically inactive as a result of pore isolation (Startsev and McNabb 2000) - these pores would not contribute to the water movement in soils (Richard et al. 2001a).

Moisture retention in soils is controlled by the capillary and adsorptive forces between water molecules and soil particles – i.e., the pore size distribution. The relationship between the water content, on a volume basis, and matric potential is described by the soil moisture retention curve. In a compacted soil the saturated moisture content and initial decline in moisture content resulting from decreased matric potentials are reduced (Hillel 1980). The air entry point or point on the curve where the largest pores are drained occurs at more negative potentials in a compacted soil compared to a non-compacted soil (Richard et al. 2001a). This occurs because compaction reduces the proportion of large pores which retain water by capillary forces and are easily drained under low matric potentials and gravitational forces. Smaller pores retain water by adsorptive forces, with a greater proportion of small pores in compacted soils results in higher water contents at lower matric potentials (Hill and Sumner 1967; Hillel 1980). At low matric potentials, residual water forms a thin film around soil particles and is tightly adsorbed to the soil surface (Startsev and McNabb 2001). The overall slope of the compacted moisture retention curve is flatter in comparison to a non-compacted soil, as a result of a change in pore structure dominated by smaller pore sizes.

1.7.2. Soil water dynamics

Water transmission in soils is affected by a decrease in macro-pore space. A compacted soil is characterized by having an anisotropic pore system, in which infiltration, redistribution and drainage are altered (Horn et al. 1995). At the soil surface, the loss of macro-pores or formation of an anisotropic pore system can have a significant effect on infiltration capacity. Startsev and McNabb (2000) found skidding caused significant decline in the infiltration capacity in a medium textured Gray Luvisol. If compaction causes a hardpan below the surface of the soil, infiltration capacity may be affected by the water content above this layer. Wet conditions created by capillary rise above the subsurface hard pan reduce the potential gradient, and can reduce infiltration at the soil surface (Alaoui et al. 2011).

Soil infiltration is further reduced by the formation of structural crusts that develop via aggregate breakdown. Crusting causes an increase in bulk density, narrower pore openings, and a lower saturated hydraulic conductivity than the layer beneath (Hillel 1998).

Platy structures dominate compacted soils; as a result the horizontal axes of aggregates are substantially longer than the vertical axes (Ayres et al. 1973). Planar pores shift from vertical orientation to predominantly parallel to the soil surface (Grevers and de Jong 1992). Pore connectivity is reduced, resulting in smaller confined pore spaces (Shukla et al. 2004). Water is impeded or must move along a tortuous path to migrate through the soil profile, in which the hydraulic conductivity is greater in the horizontal direction than in the vertical direction by several orders of magnitude (Horn et al. 1995). Etana et al. (2013) measured the distribution of dye tracing to determine water distribution in compacted and non-compacted soils. Dye coverage decreased significantly below the topsoil and 20-25cm depth due to the presence of a compacted layer. Overall dye coverage was less in the compacted soils because movement was confined to smaller pores (Etana et al. 2013).

Compacted soils are rarely uniform causing drainage patterns to be affected. Drainage is impeded as a result of reduced saturated hydraulic conductivities (Ayres et al. 1973) and formation of hardpan layers below the soil surface. Water redistributes horizontally before permeating to a more permeable layer below the compacted layer. A hydraulic barrier is formed since the large proportion of micro-pores in the compacted layer has a higher suction than the layer below. Water will not enter the more permeable layer until lower potentials are reached to draw water from the micro-pores (Barnes et al. 1971). Moderately well drained and imperfectly drained soils may be at risk of a shift in drainage class (Startsev and McNabb 2009). If the positive pressures cannot exceed the negative pressures in the soil, drainage may be restricted.

1.7.3. Compaction and its effects on the environment

Effects of compaction are widely known to cause negative impacts to the soil environment including changes to the chemical and biological properties. Reductions in macroporosity cause a decline in gas diffusion. Biota and plant roots rapidly exhaust the available oxygen; the soil becomes more anaerobic shifting the species present from aerobic to anaerobic (Startsev and McNabb 2009). Furthermore, the redox potential of the soil is affected by aeration. Compaction can result in prominent mottling and gleying, indicative of pronounced fluctuations between oxic and anoxic conditions in an Orthic Gray Luvisol (Startsev and McNabb 2009).

Plant productivity and viability are reduced with increasing compaction. Root growth is impeded because roots are unable to penetrate the dense soil and because of poor aeration in the subsoil. As a result root growth is stunted and confined to below the base of the plant. Plants that are able to survive in compacted soils may develop root diseases as a result of reduced drainage and poor aeration (Hamza and Anderson 2005). Available plant water is reduced since water is confined to micro-pores with high adsorptive forces. Roots expend more energy in root development in harsh conditions, causing reduced above ground biomass.

Ponded water as a result of impeded infiltration can cause erosion in slightly sloped landscapes of > 100 Mg ha^{-1} $year^{-1}$ (Horn et al. 1995). As a result nutrient loading may occur in the lower landscapes from migration of soil particles caused by surface water ponding and runoff.

1.7.4. Compaction amelioration

Soil compaction has become a global environmental issue because of its negative effects on the productivity and functioning of healthy ecosystems. Reclamation success is dependent on the alleviation of soil compaction and subsequent re-vegetation. Soil texture, organic matter content, level of biological activity and degree of natural soil expansion can greatly reduce the degree of compaction. For example, rapid recovery of soil properties following compaction was observed on coarse textured soils in Minnesota 1-year after heavy traffic by skidders during forest operations (Rab 2004). Regions that have pronounced freeze-thaw and wet-dry cycles can produce greater upheaval and soil loosening. As well, coarser textured soils have a greater proportion of macro-pores that promotes water infiltration and root elongation leading to soil expansion. However, natural attenuation may be too slow to meet reclamation objectives and re-establish native vegetation (Page-Dumroese et al. 2006; Spoor et al. 2003). Methods aim to improve total porosity which correlates to reduced bulk density and soil compaction (Hamza and Anderson 2005). Mechanical and biological methods create suitable soil conditions for re-vegetation purposes.

Mechanical loosening involves the use of machinery to alleviate compacted soils. Subsoiling is an effective method to ameliorate compaction (Naeth et al. 1991). Equipment selection must be chosen to induce positive changes to the soil physical structure and be economically feasible. Increasing the volume of soil per area deep ripped can be done by using winged subsoilers, paraplows, subsurface sweeps (Spoor et al. 2003) or leading tine ripper which allows multiple depth ripping at once (Hamza and Anderson 2005). Soil has shown to benefit when subsoilers are able to create fissures and cracks in the soil mass (Spoor et al. 2003) or shatter dense subsoil layers (Hamza and Anderson 2005). "Generating tensile soil failure zones by lifting the soil with a subsoiler and allowing the soil to flow and bend over the blade to create fissures while leaving the bulk of the soil intact with increased strength" is desirable (Spoor et al. 2003). Subsoil ripping depth and tine spacing has a significant effect on the degree of soil loosening. Depth of ripping is dependent on site specific properties and the thickness of the compacted layer. In general, increasing the depth of tillage and tine spacing has been most effective at alleviating compaction (Hamza and Anderson 2008).

Subsoil ripping can reduce soil bulk densities and increase macroporosity which increases soil infiltration capacity, drainage and water holding capacity (Sojka et al. 1997; Hamza and Anderson 2005). Improving these parameters can increase plant growth and productivity on reclaimed landscapes. Ripping of reclaimed mined lands in Ohio increased mixed hardwood survival from 48 to 71 % with hybrid poplar biomass increases from 1.51 to 8.97 Mg ha^{-1} (Field-Johnson et al. 2014).

Abu-Hamdeh (2003a) found air porosity to increase by 10.6 % on a soil compacted with a 17 ton axle load when subsoiled to a 45 cm depth. Air-filled pores are important for reducing anoxic water logging conditions that can occur in compacted soils. Increased porosity could indicate increased soil water storage, pore connectivity, permeability and soil water dynamics. Pagliai et al. (2004) examined the effectiveness of ripping to a 50 cm depth on pore characteristics. Storage pores (0.5-50 μm) and elongated transmission pores (50-500 μm) increased with ripping. Soil water storage is expected to increase with a greater proportion of macro-pores. Transmission pores would indicate reduced pore isolation and greater permeability.

Drainage and redistribution can be enhanced with sub soiling. Creation of large pores by rupturing compacted layers enhances the permeability of water. Vertical transmission of water increases with sub soiling. Travis et al. (1991) found water depletion and infiltration to be more uniform and an increase in the water table in a Solonetzic soil subjected to deep plowing. Indicating hydraulic barriers were mended and water movement was no longer restricted by micro-pores.

Ripping compacted soils has been found to have short-term positive effects on reclaimed landscapes. Hamza and Anderson (2005) state re-compaction will occur because of reconsolidation of the soil profile. Reconsolidation can occur from rain-fall impact energy; shrink-swell cycles and by the deposition of fine particles by wind and water into the newly constructed void spaces. Subsoiled soils are subject to re-compaction due to the weaker soil structure and internal strength (Chamen et al. 2003). Re-vegetation activities following compaction alleviation cause soil destabilization and re-compaction - timing of seeding must be planned in order to moderate these effects. Subsoiling has been found to cause initial improvements in soil bulk density and volumetric water content but these effects diminish over time (Evans et al. 1996). Deep ripping of Solonetzic soils showed no significant differences in bulk density between ripped and unripped plots 15 to 20 yrs after the ripping treatment (Mathison et al. 2002). Chamen et al. (2003), suggests that re-compaction is most likely to occur within two years following the treatment.

1.8. RESEARCH OBJECTIVES

The overall objective of this research is to assess the longevity of deep rip ploughing as a reclamation technique to alleviate compaction in order to improve soil physical and hydraulic properties to provide the necessary conditions for aspen regeneration on a reconstructed soil at the Genesee Prairie Mine, Alberta, Canada. To achieve this objective research will be focused on characterizing the static and dynamic soil water characteristics in non-ripped soils and soils ripped to a 60 cm depth.

Specific research objectives include quantification of the:

- Pore size distribution in ripped and non-ripped soils
- Differences in hydraulic measurement techniques
- Structural development of ripped and non-ripped soils
- Saturated and un-saturated hydraulic properties

1.9. HYPOTHESIS

- Pore distribution will shift in ripped soils to increase in the proportion of macropores (> 100 μm)
- Ripped soils will exhibit fractal behavior indicating the formation of increased inter- and intra-aggregate porosity and the development of aggregates
- Saturated hydraulic conductivity of ripped soils will be higher than non-ripped soils resulting from an increase in the proportion of macropores and macropore connectivity
- Trends in the saturated hydraulic conductivity for different measurement techniques will be similar in the ripped and non-ripped soils

2.0. FIGURES

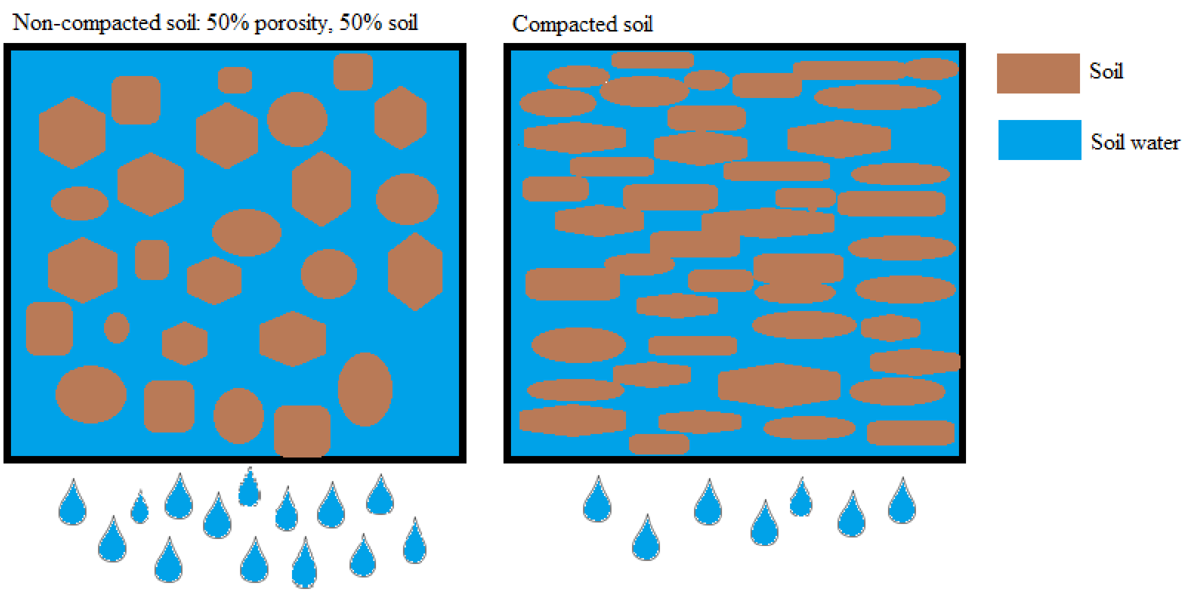

Figure 1-1. Depiction of a non-compacted and compacted soil. Non-compacted soils with well aggregated soil structure, high saturated water content and saturated hydraulic conductivity. Compacted soils indicate compressed aggregates, reduced pore space and saturated water content and hydraulic conductivity.

2. CHAPTER 2 – LONGEVITY OF SUBSOILING EFFECTS ON STATIC SOIL WATER CHARACTERISTICS OF RECONSTRUCTED SOILS

2.1. ABSTRACT

Anthropogenic activities, such as industrial development and subsequent landscape reconstruction, cause degradation to soil quality by inducing a high degree of soil compaction. Compaction negatively affects soil aggregates and the pore size distribution which increases the bulk density and alters water storage. Poor vegetative growth has been reported on many reconstructed landscapes as a result of compaction. To alleviate compaction, subsoil ripping has been employed to increase the pore volume and shift pore size distribution to intermediate and large pore classes. The main objective of this study was to assess the medium term (~ 4 yrs) effects of subsoil ripping to a depth of 60 cm at the Genesee Prairie Mine, Alberta, Canada. Soil structure and the pore size distribution, as well as the moisture retention curve (MRC) for ripped and non-ripped soils were quantified to determine residual effects of the ripping treatment four years after its application. Results showed small differences between treatments. Observable differences were noted in the 15-20 cm depth on the shape of the MRC indicating the development of some discrete medium pore classes (2-100 μm diameter) in ripped soils. Saturated volumetric water content was increased by ripping in the 5-10 and 15-20 cm depths by 1 % and 4 %, respectively. Evidence of natural attenuation by biological and physical processes was found in non-ripped soils in the surface layers (5-10 cm) as shown by a lower bulk density. Bulk density was 1.14 g cm^{-3} in non-ripped soils and 1.18 g cm^{-3} in ripped soils. Ripping appears to have a positive influence on soil structural development indicating early stages of pedogenesis and aggregate formation in reconstructed soils.

2.2. INTRODUCTION

Surface mining causes severe degradation to landscapes in terrestrial biomes worldwide. Industrial development removes vegetation, soil and overburden materials, altering the topography, hydrology and energy transfer processes (Shrestha and Lal 2011). Natural recovery of processes is often too slow to meet land reclamation objectives and to overcome many of the ecological barriers to plant establishment on these disturbed landscapes. Amelioration of the soil properties is often required to improve initial conditions through seeding of non-native species to increase soil organic matter, reduce erosion and increase soil tilth (Holl 2002; Pollster 2000). However, non-native species are not representative of the natural plant communities and reclamation objectives should be aimed at establishing native ecosystems (Macdonald et al. 2012). Reclamation increases the ability of the land to provide a safe, stable and productive ecosystem with a beneficial use to society (Maryati 2013; Mukhopadhyay et al. 2013). Without reclamation many post-mining landscapes would be unusable with poor vegetative growth, and pose health and safety issues (Alberta Environment and Parks 2015). Non-native species may be useful in achieving short term reclamation objectives but native forest ecosystems are required to improve ecosystem functioning on disturbed landscapes. Successful reclamation is based on the appropriate plant species selection that will provide the greatest and most beneficial use (Mukhopadhyay et al. 2013). Beneficial use is dependent on the objective of the end land use and is anthropocentric to who is using the land following the reclamation.

Soil that has been removed and stored is sensitive to compaction during landscape reconstruction and reclamation activities (Schaffer et al. 2007). Compaction negatively affects the establishment of forest ecosystems; poor survival and growth often occur, causing a state of arrested ecological succession (Evans et al. 2013). Compaction degrades the soil structure, weakening aggregates and reducing soil porosity (Horn et al. 1995). Macro-pores (> 100 μm diameter) are most sensitive to applied stresses (Pagliai 2004; Startsev and McNabb 2001; Startsev and McNabb 2009) and compaction generally causes a reduction of vertical, planar pores important for aeration and drainage (Grevers and de Jong 1992). For example, Schaffer et al. (2007b) found macroporosity to decrease by 20 % and 74 % after two and ten passes of heavy machinery in loamy topsoil, respectively. Poor aggregation can negatively affect soil moisture content, aeration, and bulk density which can result in reduced biological activity, yields and growth, increased anaerobic conditions, runoff and erosion (Bottinelli et al. 2014; Horn 2003; Lipiec and Hatano 2003).

Recovery of soil structure and macroporosity following compaction has been shown to be slow in forest ecosystems (Page-Dumerose et al. 2006; Rab 2004). However, some have reported that freeze-thaw cycles and biological activity can alleviate surface compaction and increase the volume of macro-pores (Bottinelli et al. 2014; Goutal et al. 2012; Lipiec and Hatano 2003). In a study by Bottinelli et al. (2014) silt loam forest soils were subjected to heavy traffic with a forwarder to initially compact the soil causing a large decrease in macro-pores. Over 2-3 years macroporosity developed in the 0-7 cm depth as a result of root channels and freeze-thaw cycles. Subsoil compaction is often persistent and mechanical management is required to improve soil conditions below 10 cm (Bottinelli et al. 2014; Etana et al. 2013; McNabb 1994). As indicated by Bottinelli et al. (2014), no significant increases in macroporosity were found at 15-30 and 30-45 cm depths due to natural processes.

Deep ripping can be an effective method to alleviate compaction by reducing the soil bulk density and increasing the proportion of macro-pores (Grevers and de Jong 1992; Hamza and Anderson 2005). Subsoiling involves lifting the soil along winged shanks of the subsoiling implement which causes the soil to break along natural planes of weakness (McNabb 1994). Aggregation can be improved by increasing the volume of macro-pores that allows for increased infiltration into the profile. Soil is subject to freeze/thaw or wet/dry periods that are increased when water infiltrates to depth (Grevers and de Jong 1992). Macro-pores provide easier entry for roots to penetrate deeper into the soil that exert pressure on soil particles pushing them into closer contact. Decomposition and root exudates can further promote aggregation through particle bonding and act as cement that coats the pore walls (Bodner 2012; Hillel 1998; Horn et al. 1994).

Fast-growth and recovery of forests on reclaimed coal mines is attributed to high soil quality (i.e., adequate nutrients, water holding capacity, tilth) (Burger and Evans 2010). Deep ripping can improve the soil conditions to allow for increased forest growth (Bulmer 2000; Pollster 2000). However, beneficial effects of subsoiling have not been long-lasting in many cases (Evans et al. 1996; Hamza and Anderson 2005). Re-compaction of subsoiled soils can occur through deposition of particles into cracks or soil consolidation (Busscher et al. 2002; Hamza and Anderson 2005). Deep-ripping has been shown to be effective from one year (Pagliai et al. 1998) to up to 2.5 to 5 years (Busscher 2002; Drewry and Paton 2000; Evans et al. 1996; McNabb 1994) and in some cases the effects of ripping have lasted decades (Baumhardt 2008; Burger and Evans 2010). Evans et al. (1996) showed that bulk density decreased with subsoiling to a 41 cm depth in the first two growing seasons but in the third growing season the bulk density was not significantly lower or was higher than prior to the treatment. Similar results have been reported by Baumhardt et al. (2008), Grevers and de Jong (1992), Mathison et al. (2002), and Reeders et al. (1993). Drewry et al. (2000) found macroporosity increased by up to 27 % in a silt loam soil. After 2.5 years, the improvements on macroporosity were evident at the 18-24 cm depth and re-compaction had occurred above the 18 cm depth.

2.3. RESEARCH OBJECTIVES

Literature available on the long-term effects of subsoiling on physical properties of reconstructed soils is limited. To assess whether subsoiling is an effective tool to improve soil properties, studies are required to address the longevity of ripping on soil physical and hydraulic properties. The overall objective of this chapter was to assess the longevity of a subsoil ripping treatment on a reconstructed compacted and un-compacted soil following surface mining. To achieve these objectives soil structural and static hydraulic properties (i.e., water retention) were assessed to determine whether ripping is an effective reclamation strategy to ameliorate heavy compaction and to improve soil conditions for aspen forest regeneration. Specific research objectives for chapter 2 are:

- To compare soil physical properties of ripped and non-ripped soils including soil bulk density, void ratios, and fractal dimension of soil aggregation.
- To compare the soil-water characteristics between the ripped and un-ripped plots at different depths by deriving the van Genuchten parameters for soil moisture retention curves and associated S-index which is the slope of the retention curve at the inflection point.
- To determine pore size distribution variability between ripped and un-ripped plots at varying depths.

- To assess the effect of macroporosity on the soil air entry potential, the potential at which the largest pores in the soil begin to drain, for ripped and non-ripped soils.
- To assess if soil compaction influences calibration curves for soil moisture probes in ripped and non-ripped reconstructed soils.

2.3.1 Hypothesis

- Ripped plots will have a fractal dimension significantly different from 3 and non-ripped plots will not have a fractal dimension significantly different from 3
- Bulk density and void ratio for ripped soils will be lower than non-ripped soils
- Ripped plots will have a greater proportion of macropores (> 100 µm) and macropore continuity
- Saturated volumetric water content will be greater in ripped plots, indicated by the van Genuchten model (1980) for moisture retention
- Air entry point on the moisture retention curve will occur at a higher matric potential in the ripped plots
- Three-dimensional laser scanning will provide valid quantification of the soil structure

2.4. MATERIALS AND METHODS

2.4.1. Site description

The experimental site is located NW¼-20-050-02 W5M (known hereafter as the site) at the Genesee Prairie Coal Mine (53°20'14.6" N 114°16'09.2" W), approximately 70 km west of Edmonton, Alberta, Canada. The site is located within the Dry Mixedwood Natural Subregion of the Boreal Forest Natural Region and Central Parkland Subregions of the Parkland Natural Region (Navus Enviromental 2010). Patches of aspen groves and aspen-white spruce forests are intermingled with crop lands in these regions. Annual average temperatures range from monthly means of 16°C to -9.9°C with a yearly average of 3.9°C, and a total of 487.8 mm of precipitation, mostly consisting of rain (Environment Canada 2014).

Land use at the site was previously agricultural with a capability rating of three which corresponds to moderate to severe limitations arising from lack of organic matter in the topsoil (fair to poor) and textural restraints in the subsoil (fair). Pre-disturbance soil was classified as a Cooking Lake Series Orthic Gray Luvisol. Slope classes of two to four (0-9 % grade) with predominate slope class of four (5-9 %) were present on the site. Prior to the disturbance ungulate browsing was mainly white tailed deer, mule deer and moose from the surrounding areas.

Mining began in the northern portion of the site in 1990 and was completed in 1992. Salvaging of selected subsoil and topsoil occurred prior to mining and materials were placed on adjacent mining areas. Soil reconstruction began in 1991 with the application of fly ash to the spoil piles that were leveled in the disturbed area. Salvaged subsoil was placed in approximately 1 m single lifts from 1991 to 1997. Deep ripping of subsoil prepared the soil for topsoil placement and brought rocks to the surface that were raked and removed. Topsoil placement began in 1993 and was completed 1999.

Following topsoil placement, a cover crop of cascade oats and boreal creeping red fescue were seeded. A grass-legume mix consisting of Algonquin alfalfa cv. (*Medicago sativa*), climax timothy (*Phleum pretense*), alsike clover (*Trifolium hybridum*) and red clover (*Trifolium pretense*) were underseeded to improve initial soil conditions. Developing soil structure and controlling erosion were the main objectives of seeding these species from 1993 to 2004. An oats crop was seeded in 2004 but environmental conditions resulted in crop failure. In 2005 an alfalfa crop was sown and annual cutting and baling was completed for five years.

The site is bordered by country access roads and reclaimed land to the north, east and west. A reclaimed haul road borders the south of the site. (Navus Environmental 2010)

2.4.2. Experimental design

An area of 575 m X 25 m was delineated for this research in the reclaimed portion of NW¼-20-050-02. Two treatments were prescribed 1) ripped (non-compacted); and 2) unripped (compacted). Six blocks each containing 2 - 36 m X 25 m plots were separated by 20 m buffers with one plot per block on the east and west sides (Figure 2-1). In the fall of 2010, a McNabb Winged Subsoiler attached to a D7R XR Caterpillar completed ripping to a 60 cm soil depth on 6 randomly selected plots. Six replicates of the ripped and unripped treatments resulted. The east and west plots were unripped and treatments alternated between these plots every 72 m. Within each plot, 4 subplots were established and randomly assigned vegetation treatments of no vegetation, a grass (*Bromus inermis*), aspen (*Populus tremuloides*), or both grass and aspen.

2.4.3. Field sampling

Soil samples were collected in August 2014. Cores were collected at 5-10, 15-20 and 30-35 cm depths at three randomly chosen locations per plot, for analysis of hydraulic conductivity and moisture retention. An additional 24 samples were collected for moisture sensor calibrations at 1 location per plot at 10 cm and 30 cm depths. Following removal of soil above the sampling depth with a shovel, sampling was completed using stainless steel cylindrical cores, 80 mm in diameter and 50 mm in height which were driven into the soil with a slide hammer and then excavated with a shovel. Following excavation, the soil in the core was leveled off with a ruler and plastic covers were placed on the top and bottom of the cores. Cores were stored in bins lined with bubble wrap for transport and stored in the lab at 0-4°C until analysis.

Clod/aggregate sample collection was based on methods outlined in Hirmas et al. (2013). Thirty-six clod samples were excavated from between the rows of planted aspen and within control subplots, with three point locations being chosen from each plot. Clod samples were collected at 15-20 cm depths using a shovel. Clods were carefully wrapped in two layers of aluminum foil and placed in zip lock bags. Samples were placed in bubble wrap lined bins for transport to minimize disturbance. In the lab, cores and clods were stored and maintained at 0-4° C until they were used for analysis.

2.4.4. Moisture retention curve

Three undisturbed cores from each plot were used to determine the moisture retention curve and pore size distribution by the evaporation and pressure plate extractor method. Three dimensional (3D) laser scanning was used to supplement the moisture retention curve data obtained, with focus on the mass fractal dimension. Clod samples were used for measurement of the fractal dimension.

Moisture retention for the wet end of the curve (> -1000 hPa) was determined by the evaporation method using a HYPROP apparatus (UMS, http://www.ums-muc.de/en/products/soil_laboratory.html). A HYPROP measures the evaporation rate and tension to quantify soil water content and water potential under natural drying conditions (Schindler et al. 2010). Three cores from each plot were measured at 5-10, 15-20, and 30-35 cm depths; 106 samples were analyzed. Methods were based off procedures by Schindler et al. (2010) and UMS (2012).

Tensiometers and the sensor unit (i.e., pressure transducer) were prepared by filling each with de-ionized and de-gassed water. Re-filling was completed by attaching the tensiometers and sensor unit to a vaccum pump for 24 hrs and maintaining a pressure close to -900 hPa. Entrapped air was removed by gently tapping the sensor units.

Sensor units were assembled by screwing the tensiometers into the thread of the sensor head. Tensions at the pressure transducers within the threads were monitored through the TensioView software on a computer to maintain tensions below 1000 hPa. Once threaded, the tensiometers tips were covered with water filled droplet syringes to equilibrate to zero for use.

Prior to the experiment, soil cores were saturated. Cheesecloth was placed on the top of the sample and secured with a perforated saturation plate. Cores were placed in a wetting bin with 2 cm of de-gassed water for 4-6 hours. Subsequently, the water level was raised to 1 cm below the core surface to obtain complete saturation. Once saturated, holes were augured into the cores to 3.75 and 1.25 cm depths. A gasket was secured around the tensiometers and the tensiometers were inserted into the soil cores.

Pressure transducers at the base of the tensiometers were connected to a desktop computer via a bus port from the sensor unit. Tensions were automatically measured in ten minute intervals. Weight changes from loss of water were manually measured by removing the bus cable and placing the soil core, holder and tensiometer on a scale. The core samples were weighed 3-4 times daily to determine the water flux (cm^3) and soil matric potentials (hPa). Experiments were completed in 7-9 days under standard laboratory conditions of 20-22° C and 98-103 kPa atmospheric pressures.

A pressure plate extractor uses applied gas pressure to push water from the soil cores and is efficient at lower matric potentials (dry soil conditions). Soil water characteristic determination using the pressure plate extractor is outlined in Reynolds and Topps (2006). Following completion of the HYPROP experiments, soil cores were sub-sampled to obtain a soil sample 5.5 cm diameter and 1-3 cm in thickness. To subsample, 5.5 cm diameter plastic cores were hammered inside the hyprop cores; excess soil was placed into tins for drying. Soil cores were covered with cheesecloth and secured with an elastic band. Cores were saturated in degassed water for 1-2 days. Initial weight was measured and recorded. A ceramic plate was saturated in de-gassed water for 24 hrs prior to use. Silica sand 1 mm in thickness was placed between the soil core and the ceramic plate to maintain hydraulic contact during equilibration. Ceramic plates were placed in pressure vessels and gas pressure was slowly applied to the vessels to equilibrate to potentials of 1000, 5000 and 15, 000 hPa for 3-5 days dependent on sample thickness. At each equilibration point the core weights were measured. Final weights were determined after oven drying the cores and tins at 105°C for 24 hrs.

Single points on the moisture retention curve determined by the HYPROP and pressure plate extractor are calculated from the change in mass of water loss (g) and the mean tension at time *t*.

Data collected at each potential was fitted to the van Genuchten (1980) model. Model fitting requires up to four parameter estimates which includes an inflection point. Measured saturated water contents were used to constrain the θ_s parameter of the VG model. Deviations from the true moisture content and those estimated by the Hyprop can occur (Schelle et al. 2013). Parameters for the moisture retention curve were determined using the HYPROP Data Evaluation Software. The van Genuchten moisture retention curve is:

$$\theta = \theta_r + (\theta_s' - \theta_r)[1 + (-\alpha h_m)^n]^{1-\frac{1}{n}} \qquad [1]$$

where S_e is the effective saturation, α (hPa^{-1}), n and m (m=1-1/ n) are empirical parameters, h_m is the matric potential (-hPa), θ is the volumetric water content (cm^3 cm^{-3}), θ_r is the residual water content (cm^3 cm^{-3}); θ_s is the saturated water content (cm^3 cm^{-3}).

A physical parameter used to assess the soil quality is the S-index which measures the slope at the inflection point on the moisture retention curve. The suggested boundary for adequate soil structure is a value of > 0.035; poor soil quality is defined as S-indexes < 0.020. S-index (S) was calculated from gravimetric water contents using the equation outlined in Dexter (2004):

$$S = -n(\theta_s - \theta_r)\left(\frac{2n-1}{n-1}\right)^{(1/n-2)} \qquad [2]$$

2.4.5. Void ratio and the pore size distribution

The void ratio is a measure of the volume of void spaces to the volume of soil solids (Hillel 1998). Void ratio is a useful measure to determine the change in porosity and pore size distribution and can be calculated from the equation:

$$e = f\frac{D_p}{D_b} \qquad [3]$$

Porosity was calculated for each core at three depths (5-10, 15-20, and 30-35 cm), given by:

$$f = 1 - \frac{D_b}{D_p} \qquad [4]$$

where f is the total porosity, D_b is the soil bulk density (g cm^{-3}) defined as the ratio of dry mass to total volume, and D_p is the particle density typically assumed to be 2.65 g cm^{-3} for mineral soils.

2.4.6. Mass fractal dimension of soil structure

Three dimensional laser scanning was used to classify soil aggregates and their mass fractal dimension. The fractal dimension is a measure of how soil aggregates hierarchically organize in the soil profile, where larger aggregates develop from smaller, more densely packed aggregates and describes the ratio of soil mass to the soil volume as a function of the scale of the structural units (Hirmas et al. 2013). Aggregate arrangement determines the soil structure and governs transport and storage processes through the effective porosity. Mass fractal dimensions were used to supplement the moisture retention curve for ripped and unripped soils. Two clods from each plot were randomly sampled at 15-20 cm for fractal analysis. The procedure used for analyzing clods is based on methods described by Hirmas et al. (2013) and Rossi et al. (2008).

A desktop computer with 3D laser scanner was used to scan clods. A parent clod was placed on the sample holder and scanned in 60° sections. The first scan family obtained is the partial 3D image of the clod. Clods were rotated to scan the remaining portions and a second scan family was produced. Scan families were aligned to produce a 3D image of the clod. Clod volume and surface area were calculated and the clod was weighed to determine the clod properties (bulk density, surface area to volume ratio, porosity, and diameter). Clods were manually broken down to five smaller size classes (0.25 – 0.5, 0.5 – 1, 1 – 2, 2 - 4, 4 - 8 cm diameter). Two aggregates from each size class were randomly selected and scanned in 72° sections. The scanner software created a 3D image of each aggregate to calculate volume and surface area. To determine the normalized clod diameter it was assumed that aggregates were spheres having the estimated aggregate volume. Aggregates were oven dried at 105°C for 24 hours and weighed to determine dry bulk density. Calculations of mass fractal dimensions were completed to characterize the mass scaling of soil aggregates from data obtained from the 3D scanning software.

Fractal dimension was determined by applying an equation outlined by Gimenez et al. (2002), it can be expressed as:

$$M(d) = k_m d^{D_m} \quad [5]$$

where $M(r)$ is the mass of aggregates (g) with diameter d (cm), k_m is a constant representing the mass of aggregates unit diameter, d is the aggregate diameter (cm), and D_m is the fractal dimension. The k_m constant was calculated by rearranging the equation. Lacunarity (L) measures the void spaces of a soil, where soils with a higher proportion of voids have a greater lacunarity and can be determined by determined by applying the equation:

$$L = var\left(\frac{M_{(di)}}{{d_i}^{Dm}}\right) \quad [6]$$

2.4.7. Soil bulk density

Dry bulk density (P_b) is a ratio of the weight of soil to the total soil volume. Soil quality is often assessed using this measure and can be used to determine the soil porosity (Rossi et al. 2008). A traditional method to calculate soil bulk density employs collecting a known volume of soil and determining its dry weight. In this study, we follow two independent methods for determination of bulk density: the core method (as described above) and the clod method. Three dimensional laser scanning can be a useful tool to calculate the bulk density of individual aggregates. P_b was determined for 34 undisturbed samples in the 15-20 cm depths using wet (saturated) core weights of the soil and soil weights of samples oven-dried at 105° C for 24 hrs.

P_b of aggregates was calculated for 24 samples at one depth (15 cm) using the three dimensional scanning method. The procedure for determining aggregate volume and mass was described in section 2.4.6 and was outlined by Rossi et al. (2008). Briefly, soil clod/aggregate sampling was discussed in section 2.4.3. Parent clods were placed on a sample holder and scanned in 60° sections to obtain the first scan family. A second scan family was obtained by repositioning the sample to expose the unscanned portions and rescanned with a 60° rotation. Scan families were matched and aligned to obtain a 3-dimensional model of the clod. Clod volume was calculated using the ScanStudio Pro Software and mass was measured and recorded on a scale. Clods were oven dried at 105°C for 24 hrs to obtain final weights.

2.4.8. Calibration of soil moisture sensors

Accurately quantifying soil moisture content through non-destructive in-situ methods is important in environmental monitoring applications. Di-electric probes are typically calibrated by manufacturers with specified equations (Mortl et al. 2011). This can result in inaccurate measurements of moisture content and potentials in compacted soils. To improve the accuracy of in-situ measurements, a soil-specific calibration curve was determined for ripped and unripped plots. Volumetric water content and dielectric constant were determined using a 5TM soil moisture probe. Procedures for the 5TM probes are outlined in Decagon Devices Inc. (2014). These measurements are calibrated with the results of moisture content determination using the pressure plate extractor. Potentials were set to 0, 50, 100, 500, 1000, 5000, and 15,000 hPa.

Undisturbed cores sampled at 10-15 and 30-35 cm were used. The sampling procedure was described above in section 2.4.3. A pressure plate extractor was used to equilibrate the cores following the procedure described in section 2.4.4 by Reynolds and Topps (2006). Briefly, cheesecloth was placed on the bottom of the cores and secured with an elastic band. Cores were placed in a wetting tank with de-gassed water and allowed to saturate for 3 days or until ponding occurs. Shrink wrap was used to cover the surface of the wetting tank to prevent evaporation. A ceramic plate was saturated in degassed water for 24 hrs prior to use. Following saturation, a 5TM moisture probe was inserted into each core and connected to an Em50 data logger for measurements. At saturation cores were weighed, and the di-electric constant and volumetric water content measured. Plates were placed in the pressure vessel, pressurized to preselected tensions, and equilibrated for up to two weeks. At each equilibration point soil weight was measured using a scale, dielectric constant and volumetric water content was recorded with the probe and data logger.

Statistical analysis was performed on the experimental data using statistical analysis R software. Linear and piecewise regressions were used to determine regression parameters, which were further analyzed using a two-tailed t-test. Statistically significant differences were reported with a probability of 0.05 for the mass fractal dimension. Model selection criterion for MRC were based on the root mean square error for the measured and fitted data.

2.5. RESULTS

The van Genuchten (VG) model reasonably depicted the soil water characteristics of ripped and non-ripped treatments. A typical limitation to using the VG model is that it does not fit well on soils with a bimodal pore size distribution and often the residual water content when fitted is negative and has to be set to zero. However, results from our study indicate that soils did not have a bimodal pore size distribution (Figure 2-2). The root mean square error for each treatment was low over the measured depths (≤ 0.058 cm^3 cm^{-3}; Table 2-1) indicating effective fitting of the VG model in our datasets. In general, no major differences were observed between ripped and non-ripped soils based on the modelled results.

Parameters for the VG model show no observed differences in saturated volumetric water content in the 5-10 and 30-35 cm depth (Table 2-1). The effect of ripping on the saturated volumetric water content was evident at the 15-20 cm depth; saturated water content was 4 % higher in ripped soils. Results showed that ripping also increased the residual water content (water content at potentials of -15,000 hPa) in the 5-10 and 30-35 cm depth by 4 % and 3 %. The slope of the curve was steepest at the 5-10 and 15-20 cm depth in ripped soils. The slope was relatively flat for non-ripped soils indicating a gradual change in water content with decreasing potentials.

Differences were observed in the shapes of the moisture retention curves for ripped and non-ripped soils at all 3 depths (Figure 2-2). Greater volumetric water contents in ripped soils were observed at tensions less than approximately 10^4 hPa in the 5-10 and 30-35 cm depths. The S index at the inflection point of the MRC showed certain numerical differences between treatments and depths (Figure 2-3). The greatest S-index median was observed in non-ripped soils in the 5-10 cm (0.020), while the more pronounced difference between treatments was observed in the 15-20 cm depth; ripped soils had an S-index of 0.018, non-ripped soils had an S-index of 0.014.

Analysis of the pore size distribution indicates little variation between ripped and non-ripped soil treatments (Figure 2-4). Pore sizes can be classified as residual ($\leq 0.2\ \mu$m diameter), storage/matrix (0.2-50 μm), structural/transmission (50-500 μm) and cracks and/or bio-pores ($\geq$ 500 μm). In our analysis all pore sizes will be expressed as median pore diameter. Matrix pores are important for water storage and structural pores for transmission and root expansion. Residual pores retain adsorbed water that remains relatively unavailable to roots. Bio-pores and cracks allow for the rapid transmission of water and typically do not retain water even at low water potentials.

Results showed no differences in median structural pore volumes (Figure 2-4). However, non-ripped soils had greater variability and were skewed higher for pore sizes in the range of 100-500 μm diameter, indicating some samples had a greater volume of transmission pores. Pores 50-100 μm in size contributed the least to the soil porosity. Storage pores were higher in the ripped soils at the 15-20 cm depth. Residual pore volume was slightly higher in the ripped soil at the 5-10 and 30-35 cm depth. No differences were observed at the 15-20 cm depth.

Average air entry potentials for ripped and non-ripped treatments were -10.84 ± 7.43 hPa and -11.66 ± 7.06 hPa, indicating no significant differences between ripped and non-ripped soils (Figure 2-5). Estimated pore diameters that drain at the air entry potential were 255 and 274 μm for non-ripped and ripped soils, respectively. In numerical terms, ripped soils had a higher air entry potential in the 15-20 and 30-35 cm depths. Slight differences were observed in the 5-10 cm depth; non-ripped soil had an air entry potential of -6.83 hPa and ripped soils had an air entry potential of -8.54 hPa. Ripped soils at the 5-10 cm depth were skewed to lower (more negative) tensions indicating some samples which have a lower air entry potential. Air entry potential was lowest (more negative) in both treatments in the 15-20 cm depth which is explained by the lower macroporosity (100-500 μm) at this depth.

Both void ratio and bulk density (g cm^{-3}) showed the same trends (Figure 2-6). A higher void ratio and lower bulk density was observed for ripped treatments in the 15-20 cm depth; ripped soil had a void ratio of 1.09 ± 0.206 and a bulk density of 1.28 ± 0.12, while non-ripped soil had a void ratio of 1.00 ± 0.26 and a bulk density of 1.34 ± 0.17. The average total void ratio and bulk density for non-ripped and ripped treatments, over all depths, was 1.14 ± 0.275 and 1.26 ± 0.17, and 1.08 ± 0.227 and 1.29 ± 0.14, respectively. The greatest difference was observed in the

30-35 cm depth where non-ripped soil had a void ratio of 1.14 ± 0.275 and ripped soil had a void ratio of 0.92 ± 0.16. Bulk density was 1.18 ± 0.10 and 1.39 ± 0.11 in ripped soils, and 1.14 ± 0.09 and 1.31 ± 0.16 in non-ripped soils in the 5-10 and 30-35 cm depths, respectively. Results showed bulk density decreased successively over four years in non-ripped soils (Figure 2-7). Bulk density of ripped soils increased in 2013 and decreased to the original measured density in 2014.

Analysis showed differences in soil bulk density and porosity between soil cores and clods (Figure 2-8). Soil clods had a higher bulk density and lower porosity for ripped and non-ripped soils in the 15 cm depth. Variability was greater for measurements made on the soil clods, with the highest variation in the non-ripped soils.

A linear regression analysis was used to assess differences between the mass fractal dimension (D_m) of ripped and non-ripped soils. Subsequently, a piece wise regression was used to assess breaks in the domain of aggregate mass-diameter relationship. A two-tailed t-test was conducted with a critical level $\alpha = 0.05$ to test whether significant differences were present between the slope of the regression and a fractal dimension of 3. Where a D_m not significantly different from 3 indicates a non-fractal pattern of soil aggregation. No significant differences were determined for the linear and piece wise regressions for both ripped and non-ripped soils (Figure 2-9). Breakpoints and parameters for the linear and piecewise regression are shown in Table 2-2.

Breakpoints for the ripped and non-ripped soils were 1.2 cm and 6 cm, respectively. Non-ripped soils above the breakpoint and ripped soils below the breakpoint show slight deviations from the non-fractal behavior. Lacunarity (L) measures the heterogeneity of the soil clods. Values for L ranged from 0.25 to 0.33, analysis shows that no significant differences exist between the ripped and non-ripped soils regarding these various fractal parameters.

Volumetric water content was plotted with the square root dielectric constant to obtain calibration curves for ripped and non-ripped soils at 10 cm and 30 cm depths (Figure 2-10) (Ferre et al. 1996). Parameter for the linear regression lines for ripped and non-ripped soils show some deviations from the standard calibration curves (Table 2-3). Some variability was observed as indicated by the R^2 values. The R^2 values for non-ripped soils at 10 and 30 cm depths were 0.73 and 0.37, and for ripped soils at 10 and 30 cm depths were 0.28 and 0.41, respectively.

2.6. DISCUSSION

2.6.1. Soil structure and pore size distribution

Pore size distribution measurements can reveal the important fluid storage and transport processes in the soil (Pagliai 2004). Assessment of structural porosity (50-500 μm diameter) shows a less than 1 % difference in pore volume between ripped and non-ripped soils at all depths (data not shown). Structural porosity is sensitive to re-compaction as the strength of the soil aggregates and the resiliency of these pores to consolidation is low, especially when organic matter is low (Pagliai 2004; Zhang 1994). However, as these pores consolidate there may be an increase in meso- and micro- structure which increases the volume of storage and residual pores. This is apparent when considering how coarse porosity shifts to fine to intermediate porosity by the packing of soil aggregates during compaction (Schaffer et al. 2007b). Our experiment shows that residual porosity is higher in the ripped 5-10 cm and 30-35 cm depth and not different between treatments in the 15-20 cm depth (Figure 2-4). This may indicate some re-compaction of the ripped soil at these depths. Although some degree of re-compaction may be occurring in our study, the small differences may be the result of shifts in pore volume to narrowing of pore classes in ripped soils. As discussed below and to further examine this hypothesis, we calculated the volume fractions for two specific pore classes: macro-pores and meso-pores.

Soils with macroporosity (pores with diameters > 100 μm) less than 10 % of total porosity are considered dense; moderately porous soils have a macroporosity of between 10-25 % (Pagliai et al. 2004). In our study, soils can be classified as moderately porous in the 5-10 cm depth and dense in the 15-20 and 30-35 cm depths (Figure 2-4 and data not shown). Surface soils are likely more porous as a result of higher organic matter content from aspen root decomposition and residual plant matter from previous forage crops.

Compaction likely caused the loss of meso-pores (50-100 μm diameter pores) when aggregates were compressed/fractured and soil particles deposited into the initially existing pore spaces (Figure 2-4). This increased the proportion of storage and residual pores. Four years after the ripping treatment, any initial effects that may have been present are no longer evident with respect to this pore class - there was no observable difference between ripped and non-ripped treatments. The low volume of these meso-pores may be attributed to pressure of growing ice crystals in the voids formed by ripping. Ice lenses may promote storage (0.2-50 μm) and macro-pore (> 100 μm) development but reduce meso- porosity (Rasa et al. 2012).

Bulk density can be used to assess compaction and a high soil bulk density negatively affects root growth, aeration and hydraulic properties (Schaffer et al. 2007a). Soils with a high bulk density and low macroporosity have low functionality (Krummelbein and Raab 2012). Ripping likely improved the soil conditions at the 15-20 cm depth as shown by increased void ratio and reduced bulk density (Figure 2-6). This is further supported by assessing differences in bulk density results between the clod and core methods for both ripped and non-ripped soils. Clods had a higher bulk density than cores (Figure 2-8) as a result of a loss of inter-aggregate pore spaces inherent during clod field sampling. Non-ripped soils had a greater difference in bulk density between clods and cores which may reflect the higher volume of pores ≥ 500 μm in the 15 cm depth (Figure 2-4). The lower bulk density of clods in ripped soils may be occurring because there is a greater volume of intra-aggregate pores (0.2-50 μm). Gimenez et al. (2002) stated that micro-cracks may develop within aggregates due to tillage causing increased intra-aggregate space, reducing the bulk density of the aggregates. In our study, ripping likely increased vertical cracks and horizontal fissures when the soil was lifted along the subsoiler shanks and was broken along natural planes of weakness. McNabb (1994) stated that "tillage may induced fracturing of soil into platy clods connected by ribbons of soil", which may promote vertical and horizontal porosity in the soil. Ripping initially increased the volume of soil, but over time settling and re-compaction may have occurred causing an increase in the bulk density of ripped soils. There was a simultaneous decrease in bulk density over the four years in non-ripped soils which was likely the result of biological and physical processes loosening the surface layers, discussed in greater detail below. Surface re-compaction may be the result of site activities such as weeding and planting, and subsoil compaction can also occur due to consolidation. Overall, lower bulk densities translated into higher saturated water content and storage porosity (0.2-50 μm). Under certain instances, reductions in bulk density may have a positive effect on available water for plants. According to Azooz et al. (1996) this would occur as pores 0.1 to 15 μm in diameter are important in increasing water availability for plants.

Results show non-fractal behavior for ripped and non-ripped soils indicating poor aggregation development and the inexistence of hierarchical aggregation (Figure 2-9). Breakpoints near the boundaries of the data set could indicate that no distinct fractal scaling domains exist (Table 2-2). Further examination beyond our measurement boundaries may be required to assess whether domains are present. By contrast, in studies conducted in grassland, cropland and forest ecosystems, aggregated soils have been found to have fractal breakpoints around 1 cm diameter (Hirmas et al. 2013) and 1 cm radii (Gimenez et al. 2002). And larger aggregates developed a pronounced fractal behavior above this threshold point.

Heavily disturbed soil resulting from compaction and tillage in a reconstructed soil profile may require extended periods of time to develop aggregate hierarchy. Deep ripping randomizes the soil structure and causes a non-fractal pattern; however, at the time the experiment was setup the initial fractal dimension was not assessed. Based on the lack of fractal structure in the non-ripped treatments, it is assumed that the soil did not exhibit a fractal structure prior to ripping and ripping could not destroy structural hierarchy when it was not present initially. Over time, increased organic matter input can actuate the aggregation process and structural development. Microaggregates, that structurally form macroaggregates, develop under the influences from root exudates and organic matter input (Tripathi et al. 2012). Soil profile reconstruction at our study site was completed recently (~ 6 yrs prior to soil sample collection) and organic inputs may be too low to promote aggregation at this time. Non-fractal patterns are more prominent at greater depths, as well. The 15 cm sampling depth may be too deep to assess if aggregation is occurring in the profile. Surface aggregation may be greater due to freeze-thaw and increased root development; further analysis is recommended under shallow soil layers in subsequent years.

2.6.2. Moisture retention curve

Results show small differences in the S index and van Genuchten parameters of the moisture retention curves of ripped and non-ripped soils (Figure 2-3 and Table 2-1). Saturated water content between treatments showed some differences, but these differences were small (≤ 1 %) in the 5-10 and 30-35 cm depth. It would be expected that the saturated water content in our experiment would be higher in the ripped soil as ripping increases the proportion of large pores (≥ 50 μm) which retain more water at low potentials, and that are typically reduced during compaction (Drewry et al. 2000; Pagliai 2004). The higher porosity in the non-ripped soils which is not coupled with an increase in saturated water content may reflect the gradual increase in porosity over time. As a result, small and medium size pores may increase but remain non-interconnected (Bottinelli et al. 2014). It is likely that as these pores develop there may be the presence of air entrapped that would decrease the saturated water content. In the 15-20 cm depth increment, the higher porosity and saturated water content indicates that there is a greater volume of structural pores (50-500 μm) in the ripped soil. Richard et al. (2001a) found that the water ratio (the volume of water per volume of solid) was 5 % higher in an Orthic Luvisol that was uncompacted compared to a compacted soil. However, structural void ratio ($\geq$ 40 μm) was significantly higher in the uncompacted soils (0.61) than compacted soils (0.02). In our study saturated water content was 4% higher in ripped soils but no differences were detected in the structural void ratio between treatments (data not shown). Our results are similar to those of Evans et al. (1996) who found the saturated volumetric water content of ripped and non-ripped soils to be 0.4 % different between treatments. Results were measured 3 years after ripping to a 41 cm depth on a clay loam soil.

The higher residual water content in the 5-10 and 30-35 cm depth indicates increased formation of very small pores $\leq$ 0.2 μm in ripped soils, although these differences were small. Compaction can increase residual water content or cause no change; in our study it appears compaction that occurred during soil reconstruction (evidenced from non-ripped soils) caused a relatively more continuous pore size distribution rather than increasing residual water content. In our study, the higher residual water content in the ripped soil appears to be the result of re-compaction; residual water content increased through loosening and consolidation of the loosened soils that can increase the volume of pores that retain water at low potentials (Richard et al. 2001a). Ripping improves porosity and pore distribution when the soil breaks along natural planes of weakness (McNabb 1994). In the deeper profile significant remolding under pressures from the subsoiling implement may have occurred, and the natural planes of weakness were destroyed causing the soil to have low resiliency to re-compaction. Low soil strength may have contributed to consolidation at in the 30-35 cm depth. Ripped soils in our study may also have been susceptible to increased residual porosity through the development of relict structural pores; slaking of poorly aggregated soils caused particles to become deposited into the recently developed fissures (Richard et al. 2001a). As well, during the ripping process the ripper shanks exert pressure on the surrounding soil matrix as the wings are pulled through the soil. As a result, meso and macro-aggregates may have been compressed causing increased micro-porosity.

Our results from the van Genuchten (1980) parameterization can be compared to those of compacted and non-compacted soils in a study by Startsev and McNabb (2001) in conifer stands of west-central Alberta. Their most severe compaction treatment (12 skidding cycles) in the first year of the study was compared to our results with the aim of conducting a pronounced contrast between our studies. Saturated water content was significantly higher in non-compacted soils; compacted soils had a water content of 0.47 $cm^3\ cm^{-3}$ and non-compacted soils 0.56 $cm^3\ cm^{-3}$ in their study. Startsev and McNabb (2001) found residual water content was not significantly different between treatments but was higher in compacted soils. This indicates non-compacted soils had a higher macroporosity. Our study showed saturated water content was not different between treatments across all depths. Results from Startsev and McNabb (2001) showed similar trends as ours only in the 15-20 cm depth (Table 2-1).

Ripping may not have caused a significant effect on saturated water content at all depths but differences in the MRC indicate some differences between treatments with respect to the pore size distribution (Figure 2-2). Ripped soils have a steeper slope in the MRC between 30/60 hPa and 1000/1500 hPa. Discrete pore classes of 2-100 μm diameter may be developing as a result of ripping. This effect is most evident in the 15-20 cm depth of ripped soils as supported by a higher S-index (Figure 2-3). A larger S-index suggests greater organization of pores into a narrower pore size distribution, where a higher S-index indicates better structural and pore development (Dexter 2004). In our study, although S-index was lower in the 5-10 and 30-35 cm depth in ripped soils, MRCs reveal that discrete pore classes (2-100 μm) may be developing. In the 5-10 and 30-35 cm depths, a higher S-index may reflect the wider distribution of pores and the lack of structural development in non-ripped soils. Dexter (2004) stated that compaction can cause aggregates to crumble (when drier) and fill void spaces, thereby reducing the volume of large pores while increasing the proportion of intermediate pore spaces. This increases the volumetric water content causing the S-index to be higher. This is supported in our study by the gradual slope of the MRC for non-ripped soils which suggests greater variability in the pore size distribution. An S-index of 0.035 is the critical value for soil quality and an index of $\leq$ 0.020 is considered a soil with poor soil quality. Accordingly, both non-ripped and ripped soils fall into poor soil quality classification with the marginal exception of non-ripped soils in the 5-10 cm.

Soils with higher macroporosity have a higher air entry point as less suction is required to draw water from pores with low adsorption and capillary forces. The air entry point for each depth should therefore correlate to the volume of macroporosity. Air entry potential follows a similar trend as the volume of macro-pores with diameters of 100-500 μm for each depth and treatment. Non-ripped soils in the 5-10 cm had a greater volume of macro-pores and a lower (less negative) air entry potential than the ripped treatment; non-ripped soils in the 15-20 cm and 30-35 cm depths had a higher (more negative) air entry point as they had a lower macroporosity than the ripped treatments. Ripping four years after its application is affecting the 15-20 and 30-35 cm depths by increasing the volume of pores greater than 100 μm. These effects are not evident at the 5-10 cm depth. Startsev and McNabb (2001) found air entry points of -140 hPa for compacted and -83 hPa for non-compacted soils. Our results were substantially different which may be the result of lower macroporosity volume in their soils. Further, differences in methods were apparent and can cause variability in the results between studies. Oxygen diffusion may be increased in ripped soils (15-20 and 30-35 cm depths) and this would induce a positive effect on initial aspen growth in our study site. According to Hernandez-Ramirez et al. (2014), root elongation and metabolism may be slowed by rhizosphere hypoxia when O_2 diffusion is low. In our study it would be expected that the lower air entry potential would allow for greater gas diffusion at higher (less negative) water potentials, therefore, improving rhizosphere conditions.

2.6.3. Longevity of ripping effects

The longevity of ripping was most evident in the 15-20 cm depth as seen by an increase in the volume of storage pores. Drying-shrinkage is important when the soil is dried to water potential lower than it has been previously (Dexter 2004). This effect may be less evident in the surface layers as they are often more porous than subsurface layers and subject to greater fluctuations in water content. However, in the 15-20 cm depth ripping likely opened the soil structure by increasing the connectivity and volume of pores that would develop around the shanks of the deep plough as it was pulled through the soil. This would allow greater fluctuations in water content and water potential. Therefore, the longevity of ripping was aided by increased structural development through aggregate formation by shrinking and swelling in this depth. Bodner et al. (2013) stated that "short-period fluctuations resulted in higher pore heterogeneity" this may occur in the ripped soil where infiltration and evapotranspiration would be expected to be higher due to increased fracturing. According to our fractal dimension results (Figure 2-9 and Table 2-2) hierarchical aggregation is absent or still low, but there may be primordial development of increased intra-aggregate pore caused by ripping as mentioned above and as indicated by higher storage porosity. Ripping also decreased the volume of residual pores which do not have a significant role in the plant available water and transport processes at tensions less than -15000 hPa.

Bulk density has been decreasing over time (Figure 2-7). Trends show no further improvement or even a slightly increasing trend in bulk density of ripped plots since 2011 over all measured depths (0-35 cm). Non-ripped soils show a decreasing trend in bulk density over 4 years. Surface compaction has reduced significantly in the non-ripped soils and may be the result of natural attenuation by aspen roots and freezing cycles/drying-shrinkage. The lower bulk density of the non-ripped 5-10 cm soils corresponds to a higher volume of pores ≥ 500 μm and 0.2-50 μm (Figure 2-3). A study by Bottinelli et al. (2014) showed that natural regeneration of small (240-500 μm diameter) and medium macro-pores (500-1000 μm diameter) was the result of plant-root penetration and physical processes. Higher amplitudes in soil temperature and precipitation, coupled with increased understory vegetation in the surface (0-7 cm) depth fully regenerated small and medium macro-pores in 2-3 years on a silty temperate forest soil. Our results are similar to these as pores ≥ 500 μm were found in non-ripped soils to be higher than ripped soils. Although not always evident, increased cracks and storage porosity appear to be having a significant effect on the bulk density at our site.

Short-term effects of ripping are evident in our study. Over time, the differences between the ripped and non-ripped soils at the 5-10 cm depth may become negligible. In our study, low organic matter may cause the soil to naturally consolidate (Bodner et al. 2013). Previous research has shown the positive effects of ripping to decrease with time (Kargas and Londra 2014; Pagliai et al. 1998). A study by Evans et al. (1996) showed that sub-soiling improved soil bulk density in the 0-15, 15-30, and 30-45 cm depths in the first year of the experiment. In the third year residual effects of ripping on bulk density were only apparent in the 15-30 cm, although this effect was not significant. Likewise, Duval et al. (1989) found surface properties of a compacted and ripped soil non-distinguishable ten years after the experimental set up. Climatic and biological factors were found to play a significant role in surface soil development causing differences due to ripping and compaction to be only secondary, masked or even negligible. This study was conducted on a Gray-Brown Luvisol with a clay texture which was seeded to silage corn throughout the experiment. Mottling in the surface (Ae horizon) was persistent and revealed periods of significant oxic and anoxic conditions caused by poor or imperfect drainage indicating similar surface conditions between compacted and ripped soils.

2.6.4. Calibration of soil moisture sensors of compacted and deep ripped soils

Calibration curves supplied by the manufacturer for determination of soil moisture content by the 5TM moisture probe are not soil specific. Soil structure can significantly affect the calibration curves. Compaction and deep ripping alter the soil structure by changing the bulk density and pore size distribution. Gong et al. (2003) stated that with increasing bulk density the intercept of the equation increases as the volume of dry soil to air increases. However, this effect may only be noticed in the area surrounding the probe and not the entire sample (Ghezzehei 2008). This relationship between increasing bulk density and increased intercepts can explain the variation for the soils and depths in our study, with the sole exception of non-ripped at 10 cm depth (Table 2-3).

Compacted soils with a higher proportion of matrix and residual pores may cause a bound water effect. Water held tightly to the soil surface has less rotation when an electrical field is applied resulting in less polarization compared to free water and a lower dielectric permittivity results (Gong et al. 2003). In our study, a higher residual porosity in the ripped soils translated into a greater intercept which may indicate a bound water effect in these soils.

Soils with higher clay contents have a higher electrical conductivity, which causes signal attenuation and overestimates the water content. The soil at our study site was classified as a loamy texture consisting of approximately 20 % clay; there were no differences in the texture between the ripped and non-ripped soils.

Compaction has been found to influence measurement errors of soil moisture probes. Water contents are generally underestimated when the soil is very wet and overestimated in the mid- to low- range of soil wetness (Ghezzehei 2008). Soil water content along the probe may be unevenly distributed due to the compaction and subsoiling treatment. As a result, the averaging of the water content along the probe may be inaccurate and cause an averaging error (Gong et al. 2003). During drying the distribution and movement of water does not occur evenly throughout the sample because of discontinuity and heterogeneity in the pore network, which may cause an underestimation of the water content and erroneous deviations from the linear regression. This is reflected in the slope of the calibration curve.

2.7. CONCLUSIONS

The objective of this chapter was to assess the impact of compaction and the longevity of subsoil ripping on pore size distribution and aggregate development in a reconstructed soil profile following surface coal mining. Pore shape and distribution can have significant influence on soil water storage, availability and air entry, as well as pore continuity and connectivity which affects the soil quality as a growing medium. Results from our study show some small differences in the pore size distribution as reflected by the van Genuchten parameters and moisture retention curve for ripped and non-ripped soils. Overall, it appears that some distinct pore classes may be developing in ripped soils as a result of the ripping treatment although the difference in pore volumes was low.

Longevity of the ripping treatment in improving soil physical properties is evident only in the subsurface layer - 15-20 cm depth. As a land reclamation practice, ripping has shown to increase fracturing at this subsurface layer which can likely allow for increased root expansion and greater fluctuations in water potentials that collectively seems to underpin the longevity of ripping effects. Further, these processes could increase the volume of medium size pores - 0.2 - 500 μm diameter. This would contribute to higher saturated water contents and also increase water availability for plants. More importantly, this higher water availability for plants entails increases in water storage at this subsurface depth over the growing season, whereas surface water is rapidly lost through evaporation. As well, ripping would contribute to root elongation through increased oxygen diffusion. Consolidation of the surface and lower subsurface depths may have caused an increase in the proportion of residual pores and loss of storage pores which can limit the water availability.

Compaction may be alleviated by biological and physical factors in the surface horizons but these factors are less effective in the deeper layers. At our study site, aspen growth coupled with physical processes may be progressively causing improved soil conditions in the 5-10 cm of non-ripped soils. Longer periods of time (e.g., > 10 years) may be required to stabilize the soil and to improve pore size distribution in on-going land reclamation sites receiving ripping management.

As indicated by MRC findings, some structural development may be occurring but aggregate formation appears to still be low. Increased OM inputs may actuate the formation of aggregate and structural development.

Previous research on the longevity of subsoil ripping has shown variable results. Main factors that affect the longevity of ripping include the depth of ripping, rainfall amount and intensity, and timing of subsoiling, soil texture and site activities following ripping. Increasing the depth ripping and reserving the timing of ripping to periods where the soil is at or near the field capacity can aid in extending the longevity of the effects of ripping. In general, lasting effects are more prominent in fine textured soils and when site activities such as planting are limited within the first few years after the ripping treatment when the soil strength it low. In areas where high and intense rainfall periods occur some re-compaction may be unavoidable in the surface layers. Addition of organic matter during ripping may reduce the degree of re-compaction in the soil by promoting flocculation and cementation of soil particles around organic materials.

2.8. TABLES

Table 1. Mean values for the van Genuchten parameters for soil moisture retention curves for ripped and non-ripped treatments at three depths. θs is the saturated volumetric water content, θr is the residual water content at -15,000 hPa tension, α is a negative inverse of the air entry potential and n is related the shape and smoothness of the curve.

Depth (cm)	Trt	α (-hPa)	n	θ_s (cm^3 cm^{-3})	θ_r (cm^3 cm^{-3})	RMSE (cm^3 cm^{-3})	Bulk density (g cm^{-3})	Porosity
5-10	NR	0.086	1.11	0.51	0.23	0.046	1.14±0.029	0.57±0.002
	R	0.030	1.60	0.51	0.27	0.038	1.18±0.046	0.55±0.002
15-20	NR	0.024	1.13	0.45	0.23	0.058	1.34±0.046	0.49±0.004
	R	0.013	1.51	0.49	0.22	0.055	1.28±0.036	0.52±0.003
30-35	NR	0.024	1.16	0.46	0.18	0.045	1.31±0.032	0.50±0.004
q	R	0.048	1.23	0.46	0.21	0.042	1.39±0.035	0.48±0.002

- van Genuchten parameters were calculated from the mean of the entire data set.

Table 2. Mass fractal dimension for ripped and non-ripped treatments. D_m is the mass fractal dimension with the standard error, k_m is the aggregate mass of unit diameter, L is the lacunarity and is a measure of the heterogeneity of the D_m.

	Ripped			
	D_m	**k_m**	**L**	**Breakpoint (cm)**
d	3.018 ± 0.029	0.89	0.025	
d≤ db	2.700 ± 0.44	0.90	0.027	1.21
d > db	3.034 ± 0.45	0.89	0.025	
	Non-ripped			
	D_m	**k_m**	**L**	**Breakpoint (cm)**
d	3.004 ± 0.021	0.95	0.025	
d≤ db	3.03 ± 0.026	0.96	0.025	6.10
d > db	2.64 ± 0.82	2.03	0.0033	

Table 3. Parameters from the calibration of the 5TM moisture probes for ripped and non-ripped soils at 10 cm and 30 cm depths.

		Parameters for calibration curves			
Depth (cm)	**Treatment**	**Slope**	**Intercept**	**R^2**	**Bulk density (g cm^{-3})**
10	Ripped	0.10	-0.030	0.28	1.27
30	Ripped	0.094	-0.025	0.41	1.50
10	Non-ripped	0.18	-0.35	0.73	1.39
30	Non-ripped	0.11	-0.089	0.37	1.41

2.9. FIGURES

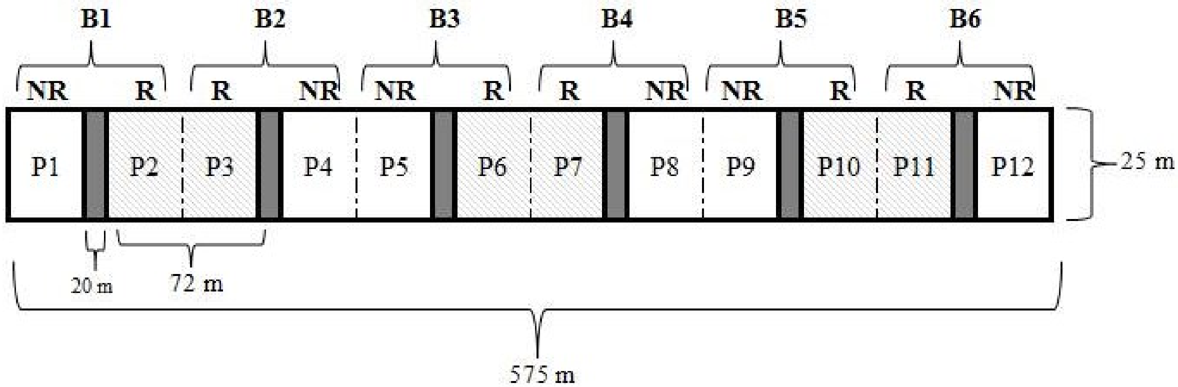

Figure 2-1. Experimental design layout of the study site showing ripped (R) and non-ripped (NR) soil plots. B is block and P is plot. Solid gray blocks represent buffer areas between ripped and non-ripped treatments.

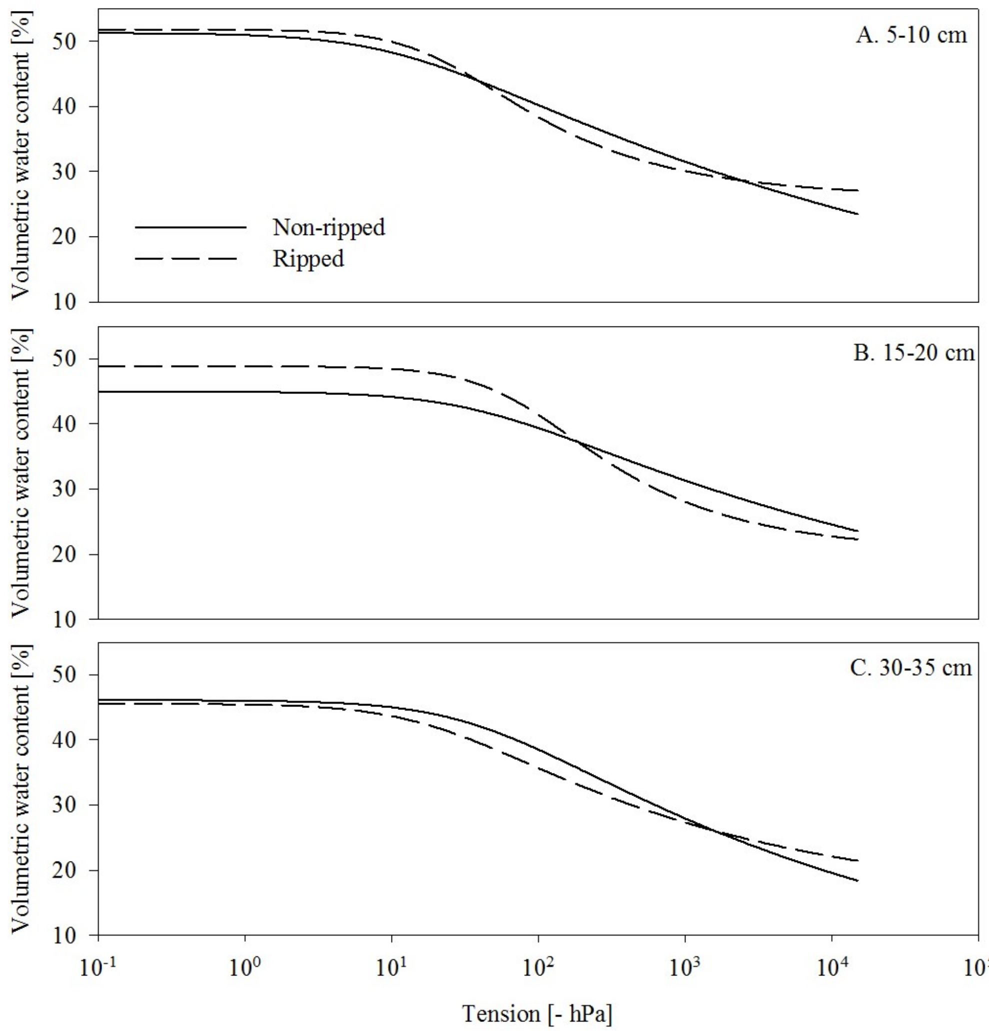

Figure 2-2. Soil moisture retention curve for ripped and non-ripped soils at 3 depths (A. 5-10, B. 15-20, and C. 30-35 cm) fitted to the van Genuchten model for moisture.

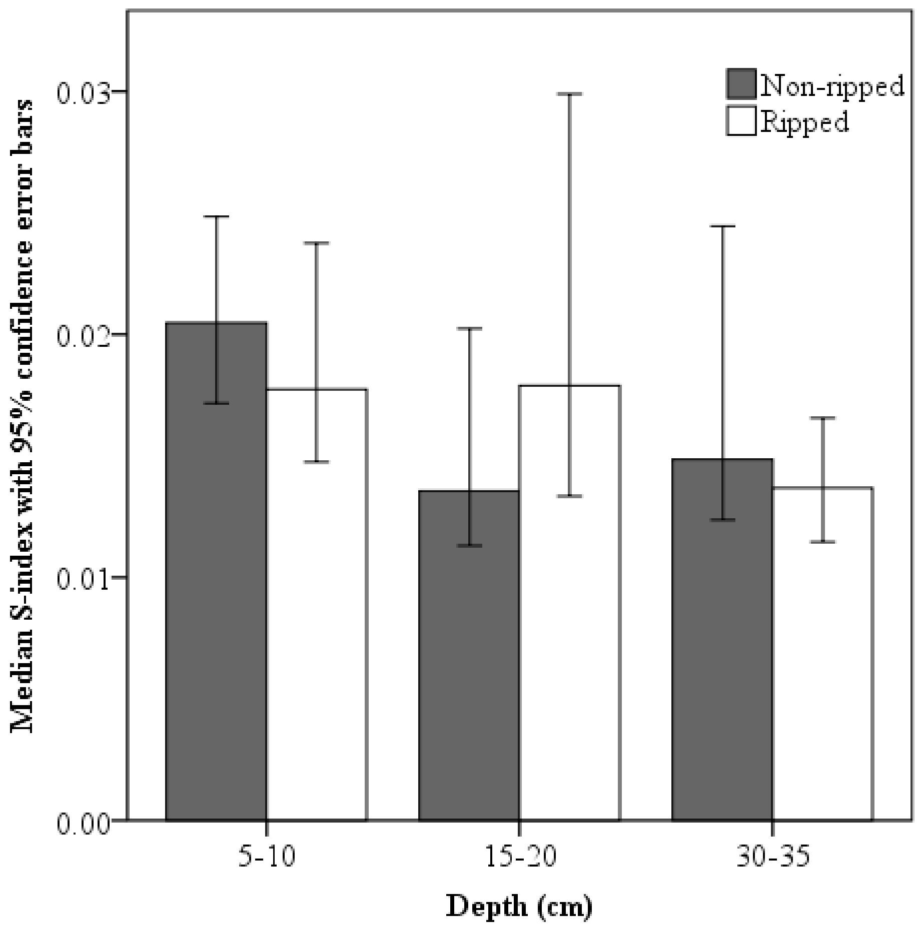

Figure 2-3. Median S index for ripped and non-ripped soils at 3 depths (5-10, 15-20, and 30-35 cm) calculated with gravimetric water contents. S index is a measure of the slope of the moisture retention curve at its inflection point.

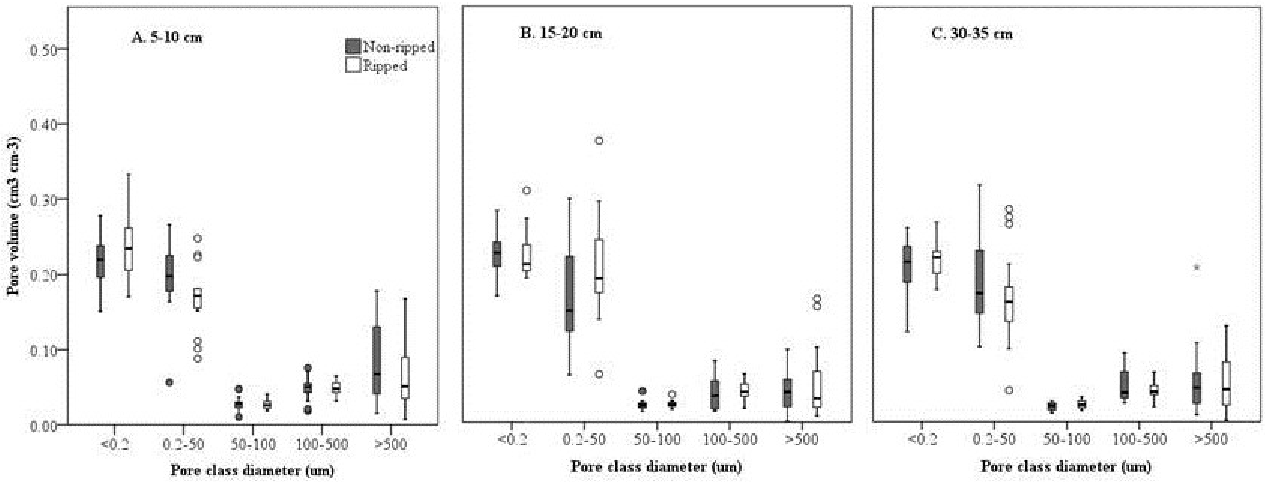

Figure 2-4. Pore size distribution for ripped and non-ripped soils at 3 depths (5-10, 15-20, and 30-35 cm) for five different pore classes (< 0.2, 0.2-50, 50-100, 100-500, and > 500 um).

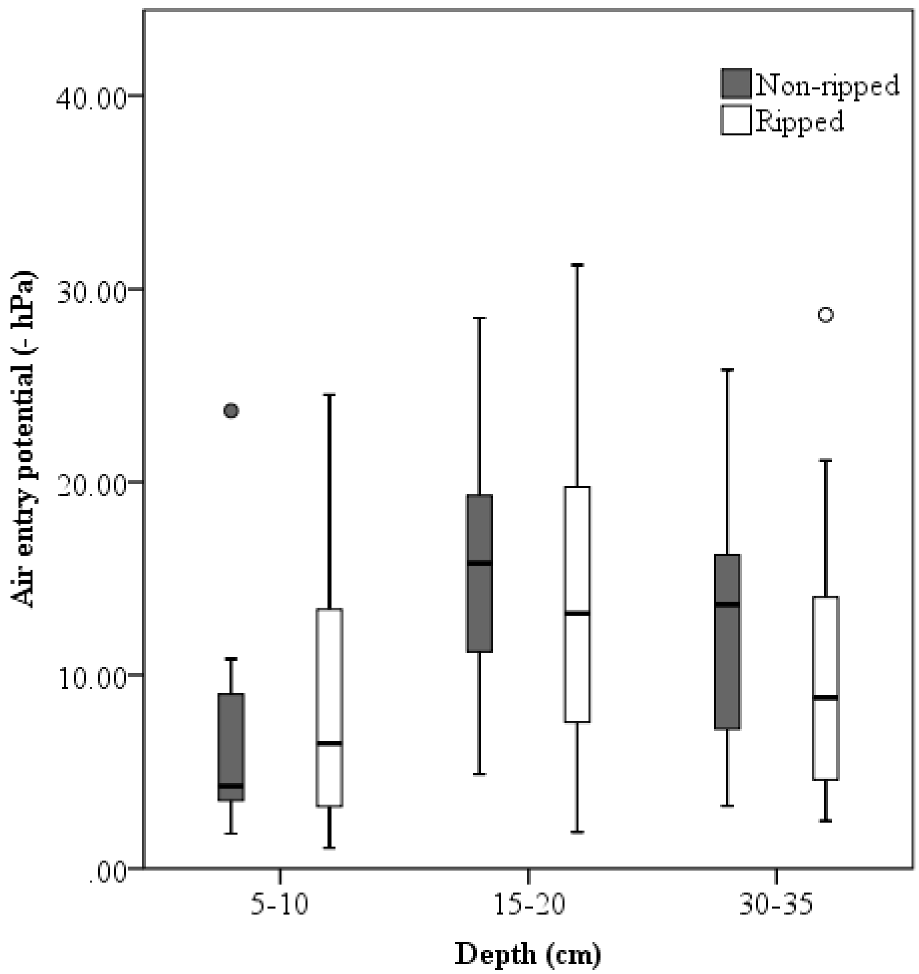

Figure 2-5. Air entry potential (-hPa) for ripped and non-ripped soils at 3 depths (5-10, 15-20, and 30-35 cm).

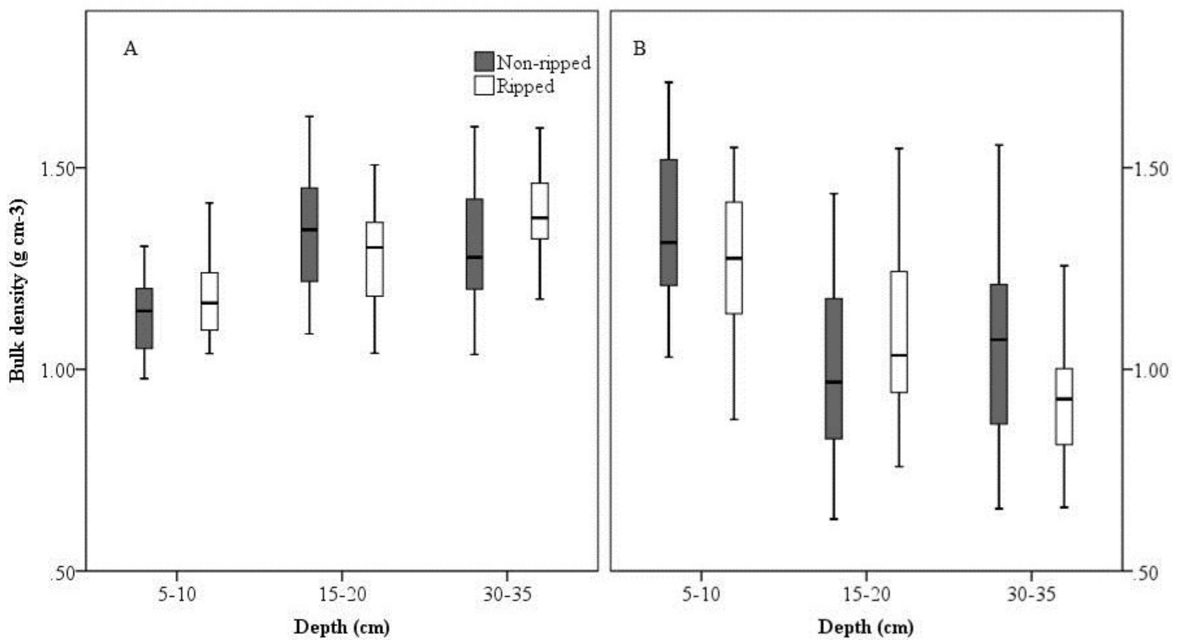

Figure 2-6. Bulk density (g cm^{-3}) (A) and void ratio (B) for ripped and non-ripped soils at 3 depths (5-10, 15-20, and 30-35 cm). Soil core method.

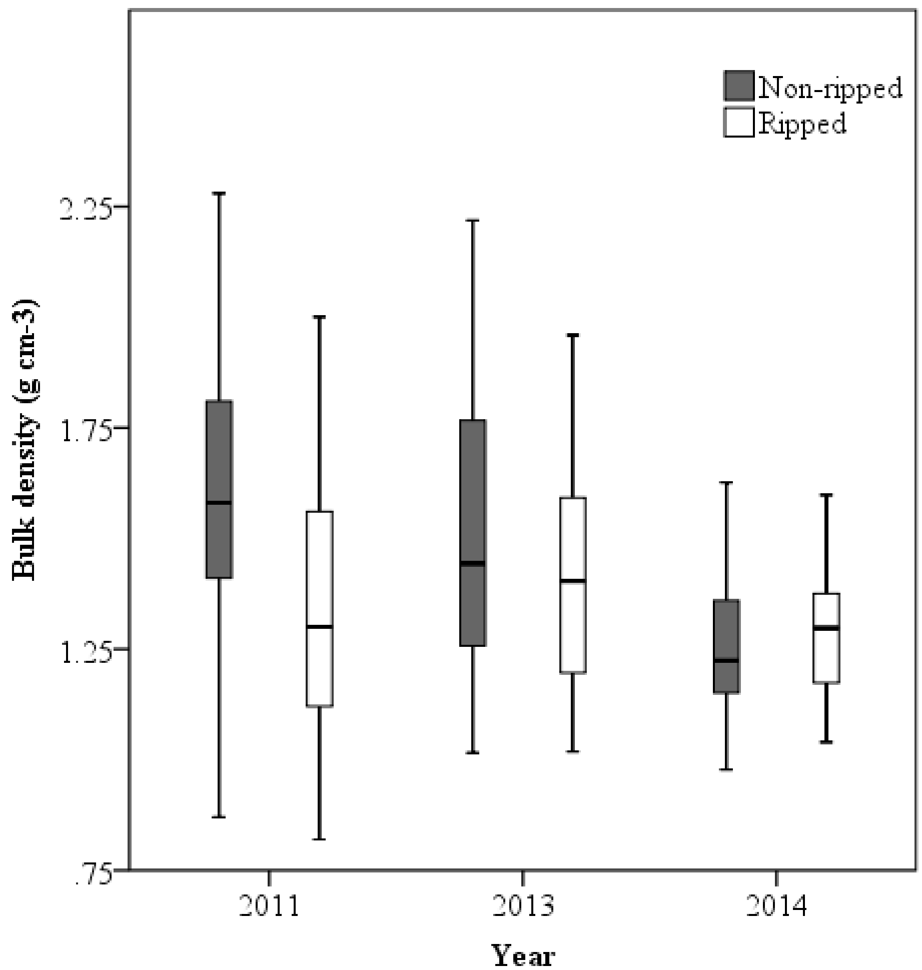

Figure 2-7. Average soil bulk density (g cm^{-3}) over four years in the 0-35 cm depth increment. Soil core method.

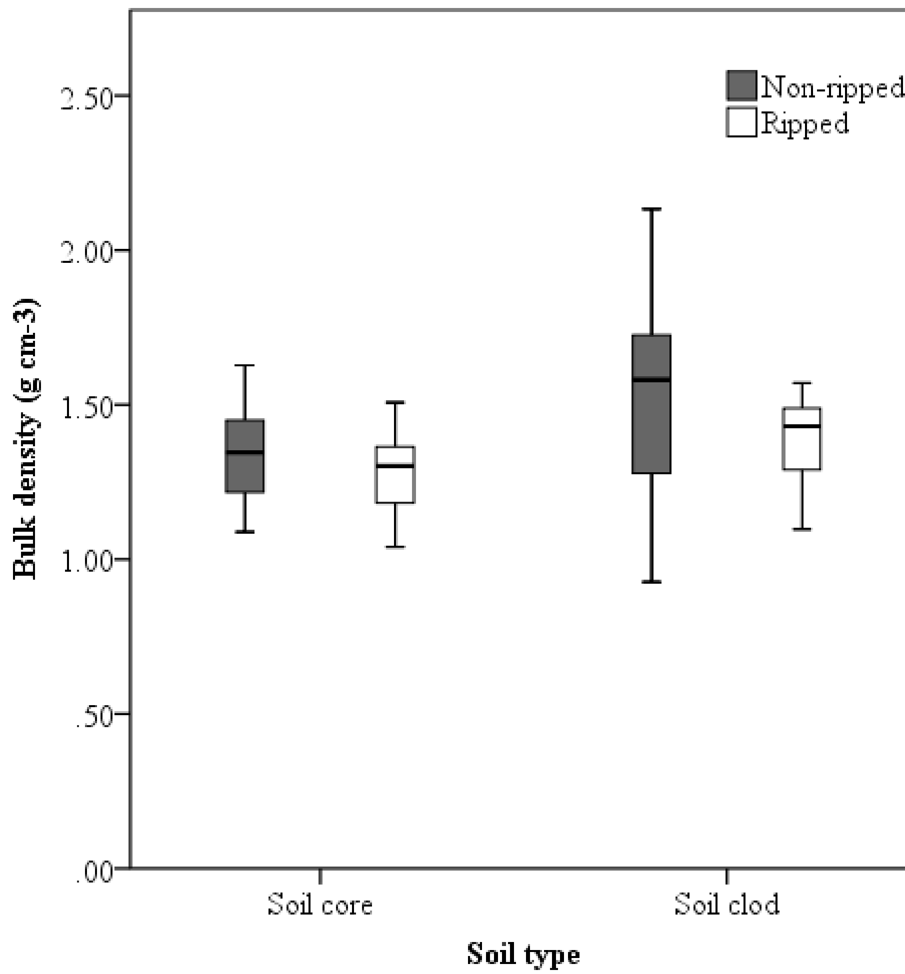

Figure 2-8. Bulk density (g cm^{-3}) and porosity measured from soil clods by the laser scanner and from soil cores from the Hyprop at the 15 cm depth.

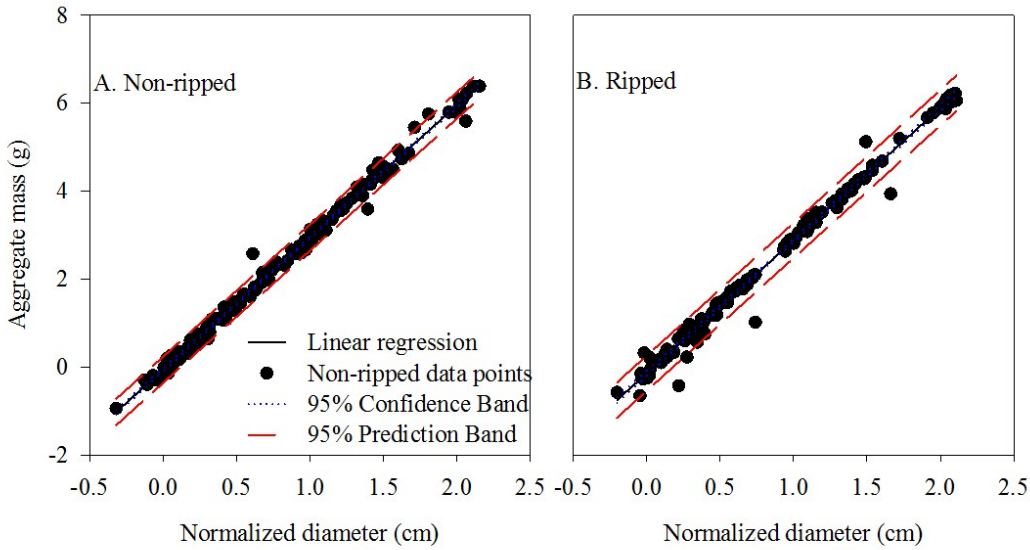

Figure 2-9. Fractal scaling of the natural log transformed aggregate mass and diameter for the non-ripped (A) and ripped (B) treatments. Normalized diameter is the cubic root of the aggregate volume.

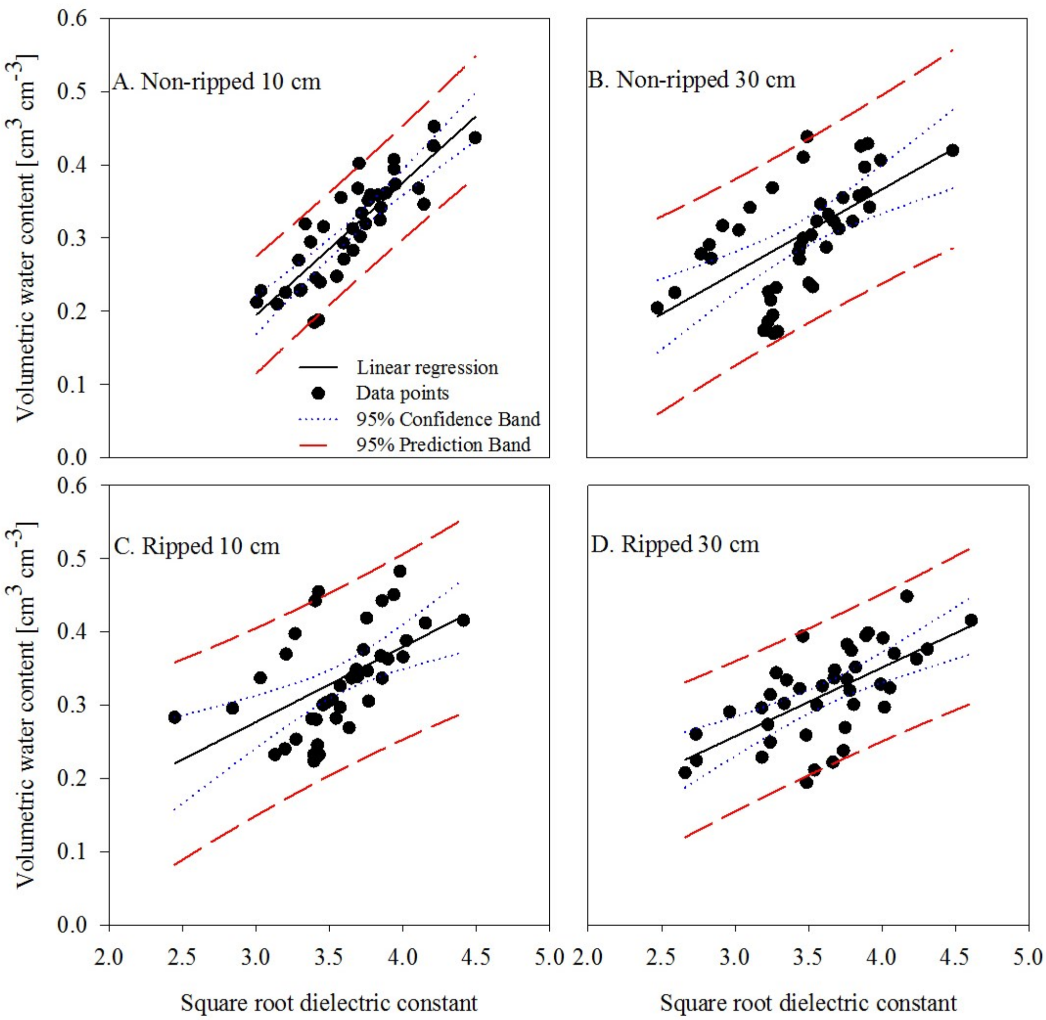

Figure 2-10. Calibration curves for 5TM moisture sensors for ripped (C & D) and non-ripped (A & B) soils at 2 depths (10 cm and 30 cm).

3. CHAPTER 3 – LONGEVITY OF DEEP RIPPING ON DYNAMIC SOIL WATER PROPERTIES IN A RECLAIMED MINE SITE

3.1. ABSTRACT

Mining causes significant alterations to soil properties, and reclamation aims to improve soil physical quality to provide a suitable growing media. Compaction, resulting from repeated movement of heavy machinery and equipment, negatively affects the pore volume, connectivity and distribution altering the soil water dynamics including infiltration, percolation (i.e., hydraulic conductivity), drainage, and evapotranspiration. Previous research has shown subsoiling can improve soil physical quality; however, long-term studies on the effects on water transport and storage are limited. The main objective of this research was to quantify the longevity of subsoil ripping on saturated and un-saturated hydraulic characteristics of a reconstructed soil. Subsoil ripping was completed to a 60 cm depth. Results show beneficial medium-term effects of ripping on infiltration and saturated hydraulic conductivity compared to the control. Hydraulic barriers to percolation were likely reduced by ripping as indicated by a saturated hydraulic conductivity 3 and 2 times higher in the 15-20 and 30-35 cm depths, respectively. Saturated hydraulic conductivity was 0.037 cm min^{-1} and 0.025 cm min^{-1} in ripped soils and 0.011 cm min^{-1} and 0.012 cm min^{-1} in non-ripped soils in the 15-20 and 30-35 cm depths, respectively. Subsoil ripping can be an effective reclamation tool to improve water flow in compacted reconstructed soils.

3.2. INTRODUCTION

Anthropogenic activities such as coal mining cause significant degradation of terrestrial ecosystems, such as the boreal and aspen forests. Surface mining is one of the most significant forms of anthropogenic disturbance to an ecosystem through the removal of vegetation and degradation of soil quality through overburden extraction and stockpiling (Shrestha and Lal 2011). Alterations to the topography cause changes to the hydrologic and energy cycling on these disturbed landscapes. Reclamation of disturbed lands focuses on improving the land value and quality for future use, as well as increasing the ecosystem function and services including biodiversity, water filtration, nutrient cycling, climate regulation and aesthetics (Prach and Hobbs 2008). The resulting reclaimed areas are expected to have equivalent land capability to prior to the disturbance or to adjacent landscapes. Likewise, post mining sites should reflect natural forest ecosystems and have specific characteristics of native plant communities (Alberta Environment 2006 and 2009). Initial reclamation often involves using non-native species to rapidly establish ground cover to improve soil conditions. These species have different ecosystem structures and functions, and may not provide the same ecosystem services as natural forest ecosystems (Evans et al. 2013). Reclamation to forested lands is critical to reduce fragmentation in boreal and aspen forest stands and to create a self-sustaining ecosystem capable of supporting multiple end land uses (Shrestha and Lal 2011).

Following disturbance due to open-pit mining, recovery of forested ecosystems can be often quite slow as a result of poor soil conditions. A common issue in reconstructed soils is the high degree of soil compaction following reclamation is common (Shrestha and Lal 2011). Compaction causes an anisotropic pore system resulting in lower macroporosity, increased tortuosity and low soil pore connectivity (Horn et al. 1994; Schaffer et al. 2007a). Reduced macroporosity and pore connectivity can lead to lower soil permeability (Horn et al. 1995; Richard et al. 2001a). Poor soil aggregation can result in surface sealing when particles are displaced and deposited into the pores at the soil surface (Li et al. 2009; Pagliai et al. 1998). These factors affect soil water dynamics by altering infiltration, drainage, redistribution and evapotranspiration. Increased runoff and erosion can occur in compacted soils as a result of altered soil water dynamics. Soil permeability is an important factor to consider when assessing soil compaction. Permeability may be a better indicator for the degree of soil compaction than bulk density as it can reveal details on the structural properties of the soil (Abu-hamdeh 2003b; Startsev and McNabb 2000). Startsev and McNabb (2000) found that a 12 trafficking cycle by a skidder caused a significant increase in the soil bulk density which caused reductions in the unconfined infiltration rate and saturated hydraulic conductivity of a medium-textured soil in the foothill and boreal forest regions of Alberta, Canada. After 3 years, the infiltration rate and conductivity of the soil had not returned to pre-compaction levels indicating poor recovery of hydraulic properties in compacted soils. Although a significant increase in bulk density was reported, in some studies an increase in the infiltration rate of the soil is not correlated to a decrease in the soil bulk density (Baumhardt et al. 2008). This indicates the presence of a few hydraulically important pores that do not contribute significantly to a decrease in the bulk density but are important in flow processes (Guebert and Gardner 2001).

Subsoiling is an effective method to improve soil water dynamics by increasing the soil pore volume and connectivity. Subsoiling can improve infiltration, redistribution and evapotranspiration (Travis et al. 1990) as deep ripping creates an open pore system with discrete pore classes increasing the water flow through the soil (Pagliai et al. 2004). Compaction results in hydraulic conductivity reductions by several orders of magnitude but deep ripping can increase the permeability of the soil by improving the soil structure. Following timber extraction in Alberta, Canada, tillage practices have been found to increase the recovery of compacted soils when coupled with freeze-thaw cycles (McNabb 1994). Intense freeze-thaw cycles can also increase the proportion of cracks or fissures at the surface which promotes water flow to deeper in the soil profile (Guebert and Gardner 2001).

Soils subjected to subsoil ripping are susceptible to re-compaction due to their weakened internal structure (Horn et al. 1995). Particle deposition (i.e., fine particles such as sand, silt, clay, and colloids), consolidation or site management (i.e., weeding, planting) can cause the ripped soil to counter-productively become more compacted than it was prior to ripping. This can cause the hydraulic conductivity be lower than a non-ripped soil (Azooz et al. 1996). Conversely, other studies have shown that deep ripping increases the infiltration rate and saturated hydraulic conductivity for several years (Abu-Hamdeh 2003b; Chong and Cowsert 1997; Sojka et al. 1993). Abu-Hamdeh (2003b) found the infiltration rate to be higher 3 years after a compaction and deep ripping treatment. Infiltration increased by 11 % and 9 % on 6.35 and 15.42 ton load plots in a clay loam soil. Collectively, the apparent discrepancies in these earlier reports can suggest that subsoil ripping effects depend on site-specific interactions driven by local climate, landscape characteristics, and preexisting soil properties such texture and organic carbon content. This notion substantiates the need to undertake new studies addressing the usefulness and duration of ripping effects.

Soil hydraulic properties can be measured with in-situ (field methods) or ex-situ (laboratory) methods. Many of these methods characterize the saturated or near-saturated hydraulic characteristics through measurement of infiltration rates. Although there is value on using both field and laboratory methods, field methods are often preferred to laboratory methods as measurements are more representative of the true soil conditions and do not discriminate between the spatial variability of hydraulic properties (Bagarello et al. 2006; Buczko et al. 2006; Stolte et al. 1994). Differences in the results of the measured conductivity between lab and field methods have been observed in many studies (Buckzo et al. 2006; Lee et al. 1985; Stolte et al. 1994).

3.3. RESEARCH OBJECTIVES

Research literature focusing on the longevity effects of subsoil ripping on key soil properties after multiple years following treatment is very limited. The objective of this chapter was to determine the longevity of a subsoil ripping treatment on alleviating compaction on a reconstructed soil following surface mining in central Alberta, Canada. Examination of dynamic soil hydraulic properties is necessary to determine if subsoiling will improve soil properties to establish aspen forests as part of land reclamation efforts. Specific research objectives for this chapter are as follows:

- To compare the infiltration rate near saturation in non-ripped versus ripped soil plots
- To determine if hydraulic barriers have developed in non-ripped and ripped soil plots
- To assess the variability of saturated hydraulic conductivity between ripped and non-ripped soils at varying depths
- To evaluate the longevity effects of deep ripping on unsaturated hydraulic conductivity and moisture retention properties
- To compare field- and laboratory-measured saturated hydraulic conductivity on ripped and non-ripped plots by contrasting steady and un-steady state flow methods

3.3.1. Hypothesis

- Saturated hydraulic conductivity will be significantly greater in the ripped plots with increased variability between ripped plots

- The slope of the unsaturated hydraulic conductivity curve for ripped plots will decrease more steeply with increasing matric potential and will eventually become less than unripped plots, using the van Genuchten-Mualem model (1980) shape parameter m as the indicator

3.4. MATERIALS AND METHODS

3.4.1. Site and climate description

The site was described previously in Chapter 2 section 2.4.1. Briefly, the Genesee Prairie Coal Mine (53°20'14.6" N 114°16'09.2" W) is located 70 km west in the Dry Mixedwood Natural Subregion and the Central Parkland Subregion. Prior to mining the site was used for agriculture and was located on Orthic Gray Luvisols. Mining began in 1990 and was completed in 1992. Reclamation began with soil reconstruction in 1991 and was completed in 2001, where annual forage was completed until 2003. In 2004 the site was seeded with oats, but climatic variables and wildlife resulted in complete crop loss. Annual cropping was completed from 2005 to 2009. (Navus Environmental 2010)

Soil temperature data at 5 cm depth was taken from a permanent weather station situated in Tomohawk, Alberta which is located 60.6 km west from the study site. Data was downloaded from the Alberta Agroclimatic Information Service network Agriculture and Rural Development from 1 January 2011 to 31 December 2014 (Agroclimatic Information Service 2014).

3.4.2. Experimental design

The experimental design was discussed previously in Chapter 2 section 2.4.2. Briefly, in 2010, an experimental research site was established at the Genesee Prairie Mine, in the reclaimed portion of NW¼-20-050-02. An area of 575 m X 25 m was delineated and stratified into six blocks each containing 2 – 36 m X 25 m plots separated by 20 m buffers. In the fall of 2010, a McNabb Winged Subsoiler attached to a D7R XR Caterpillar completed ripping to a 60 cm depth on 1 plot per block resulting in 6 replicates of control and ripped treatments. Plots were further subdivided into 4 vegetation treatments of no vegetation, a grass (*Bromus inermis*), aspen (*Populus tremuloides*), or both grass and aspen.

3.4.3. Field sampling and measurements

Soil cores were collected at 5-10, 15-20, and 30-35 cm depths, at three randomly selected locations per plot, for analysis of hydraulic conductivity. Sampling was completed using a stainless steel cylindrical core, 80 mm in diameter and 50 mm in height, which was driven into the soil following removal of soil above the sampling depth with a shovel. After insertion, cores were excavated using a shovel. Following excavation, samples were sealed with plastic caps and placed in bins lined with bubble wrap for transport and were stored at 0-4° Celsius prior to measurements.

Saturated hydraulic conductivity (K_s) is a soil hydraulic property describing the ease at which water is transmitted through a permeable porous media. Air entrapment in pores can cause K_s to be less than if all the pores were saturated, this is referred to as field saturated hydraulic conductivity (K_{fs}). Field measurements of K_{fs} began in May 2014, using a Guelph Permeameter and Tension Infiltrometer (TI). Six locations were randomly chosen in the rows between planted aspen in the ripped and unripped plots. Three measurements were made using each method; 36 measurements per method and depth were completed for a total of 36 and 144 measurements made with the tension infiltrometer and Guelph Permeameter, respectively. As texture is relatively homogenous throughout the profile infiltration rates were measured at 0, 15, 30, 45, and 60 cm depth intervals. To manage for interaction between successive measurements in one location, each depth measurement was spaced 1 m apart in a randomly chosen direction.

3.4.3.1. Field measurement of saturated hydraulic conductivity – Guelph Permeameter

A Guelph permeameter is a device that facilitates an easy and reliable, constant head determination of in-situ hydraulic conductivity in the ranges of 10^{-1} to 10^{-4} cm min^{-1}. The procedure for operation and measurements with a Guelph permeameter is outlined in Soil Moisture Equipment Corp. (2012). Three measurements at each depth per plot were completed. A borehole was excavated using a sizing auger 10 cm in diameter to a specified depth. To reduce smearing a well prep brush was inserted and lifted one or twice in the borehole. Prior to measurements the soil surrounding the borehole was saturated to produce a saturation bulb.

The Guelph permeameter and tripod were assembled following the procedure outlined in the Guelph permeameter Operating Instructions (Soil Moisture Equipment Corp. 2012). Once assembled the reservoir value was turned to face the notch upwards to connect the inner and outer Reservoirs. The fill plug was removed from the reservoir cap and water was poured into the recess. The reservoir was filled completely until no air bubbles materialized from the fill hole. The tripod was positioned directly above the borehole and the permeameter was slowly lowered into the opening. For measuring the 45 and 60 cm depths, the permeameter was removed from the tripod and slowly placed into the borehole until the water outlet tip rested on the borehole bottom. A tripod bushing was placed and secured at the top of the borehole to stabilize the permeameter.

The two-head procedure was used for determining water outflow. Before making a reading, the reservoirs were connected, the well height indicator and well head scale were seated down, the fill plug was replaced in the reservoir cap and the vacuum tube closed off. A well head (H_1) height was obtained by raising the air tube on the well height indicator to 10 cm in the 15 cm depth borehole and 20 cm depth in the 30, 45, and 60 cm boreholes on the Well Head Scale. The second well height, H_2, was set at 10 cm for the 15 cm depth and 20 cm for the 30, 45, and 60 cm depths. Water level height was recorded in selected intervals to calculate the rate of fall of water calculated by:

$$R = \frac{\Delta water\ level}{Time\ Interval} \qquad [1]$$

When three consecutive readings were produced the steady state rate of fall was determined, R (cm s^{-1}).

The calculations for determining K_{fs} are described below. A soil texture-structure category of unstructured loam soils was assigned for the shape factor, given by:

$$C = \left(\frac{H_i/a}{1.992+0.091\left(H_i/a\right)}\right)^{0.683} \qquad [2]$$

Where C is a shape factor, H_i is the water head height (cm), a is the borehole radius (cm). Field saturated hydraulic conductivity (K_{fs}) was calculated using a modified Darcy's Law:

$$Q = R + \ 35.22 \qquad [3]$$

$$K_{fs} = \frac{C*Q}{\left(2\pi H_i{}^2 + \pi a^2 C\right)a^* + 2\pi\left(\frac{H_i}{a^*}\right)} \qquad [4]$$

where R is the steady-state fall of water in reservoir at H_i (cm s^{-1}), Q is the flux rate in reservoir at H (cm^3 s^{-1}), and K_{fs} is the soil field saturated hydraulic conductivity (cm s^{-1}). The calculation was repeated for H_1 and H_2, and the values for K_{fs} averaged to obtain a final measure of conductivity.

3.4.3.2. Field measurement of infiltration rate

Tension infiltrometers measure water infiltration under a known negative pressure head that is imposed at the soil surface. A tension infiltrometer was used to measure the infiltration rate at 3 locations in each plot for a total of 36 measurements. Methods for the tension infiltrometer are outlined by Soil Measurement Systems (2013). Prior to placing the infiltrometer, debris was removed and the soil surface was levelled. A 20 cm diameter ring was placed on the soil surface and filled with a 2 mm layer of fine sifted sand to obtain good hydraulic contact between the infiltrometer and soil surface. Prior to measurements the surface was pre-wet.

Tension infiltrometers are composed of a 20 cm diameter infiltration disc connected to a water tower through a tube, and a bubble tower used to control the water tension. An infiltration disc was placed in a plastic bin to fill with water. After filling, the water tower and bubble tower were filled with water and the caps were replaced while leaving one outlet open until the caps were closed. Air entrapped in the disc or attached tube was removed by lifting the tower above the disc and gently shaking air bubbles to the top of the tower. To set the tensions, the bubble tower air inlet tube was moved to -5, -10, and -15 cm ((+) -4 cm for calibration) below the water level in the bubble tower. Experiments began by placing the disc on the sampling location, opening the tubing clamp between the bubble tower and water tower, and the tubing clamp on the top of the water tower. Infiltration measurements were conducted for 40 min at each tension and water levels were recorded manually at 2 min intervals. .

Applying the Ankeny et al. (1991) method, the infiltration rate was calculated using two different tensions, given by:

$$Q(\varphi_1) = \left[\pi r^2 + \frac{4r}{A}\right] K(\varphi_{1)} \quad [5]$$

$$Q(\varphi_2) = \left[\pi r^2 + \frac{4r}{A}\right] K(\varphi_{2)} \quad [6]$$

where $A = \frac{K(\varphi)}{\emptyset(\varphi)}$ is a constant (cm^{-1}), $Q(\varphi_i)$ is the steady infiltrating rate ($cm^3\ min^{-1}$), $K(\varphi_{i)}$ is the field saturated hydraulic conductivity ($cm\ min^{-1}$), $\emptyset(\varphi)$ is the matric flux potential at the soil surface (cm), and r (cm) is the radius of infiltrometer disc.

3.4.4. Laboratory measurement of saturated hydraulic conductivity – Falling head method

The falling head method (un-steady state flow) is designed to measure saturated hydraulic conductivity in the ranges of 10^{-1} to 10^{-4} cm min^{-1}. The procedure for determining conductivity with the falling head method is described by Klute and Dirksen (1986). Three undisturbed soil cores at each depth (5-10, 15-20, and 30-35 cm) per plot were measured.

Soil samples were kept in the stainless steel cylindrical cores during analysis to prevent lateral flow of water. Cores were covered on both ends with cheesecloth, held by elastic bands. Cores were placed top down in a wetting bin with 2 cm de-aired water for six hours. Additional water was added to 1 cm below the top of the sample to allow for complete saturation. A baseplate and top-plate were secured to the cores with an o-ring placed between the core and plate to seal and prevent lateral flow of water. A burette was attached to the top of the plate using a pvc tube. The burette was filled to 0 ml and the rate of water fall was recorded at regular intervals. When three consecutive measurements were made the experiment was complete.

The saturated hydraulic conductivity was determined using an equation adapted from Darcy's Law by Klute and Dirksen (1986), it can be expressed as:

$$K_s = \frac{aL}{A(t1-t0)} \cdot ln\left(\frac{b0+L}{b1+L}\right) \quad [7]$$

where K_s is the saturated hydraulic conductivity (cm min^{-1}), a is the cross sectional area of the standpipe, L is the soil sample length (cm), A is the cross sectional area of the sample, $t1$ is the time for the water level in the standpipe to fall from height one to height two (min), $t0$ is the initial time (min), $b0$ is the initial water level height (cm), and $b1$ is the water level height after time x (cm).

3.4.5. Laboratory measurement of unsaturated hydraulic conductivity – Evaporation method

Unsaturated hydraulic conductivity measurements were conducted by Schindler's method by natural evaporation from soil columns (Schindler et al. 2010). A HYPROP system (UMS and http://www.ums-muc.de/en/products/soil_laboratory.html) measures the evaporation rate and tension to quantify hydraulic conductivity under natural drying conditions (Schindler et al. 2010). Unsaturated hydraulic conductivity was measured at three depths (5-10, 15-20 and 30-35 cm) per plot. Methods were based on procedures by Schindler et al. (2010) and UMS (2012). Briefly, a HYPROP unit consists of a sampling ring holding tensiometers connected to sensor units containing the pressure transducers. The hyprop measures the soil water matric potential with tensiometers at two elevations within cores that are connected to pressure transducers. Two holes were augured into the soil cores to 1.25 and 3.75 cm depths where the ceramic cups of the tensiometers could be inserted.

In preparation for measuring the unsaturated hydraulic conductivity, the tensiometers and sensor unit were filled with de-ionized and de-gassed water and pressurized to -0.9 bar for 24 hrs. Soil cores were saturated by placing cheesecloth and perforated attachment to the top of the sample. Soil was saturated slowly over a period of 2-3 days with de-gassed water at an initial level of 2 cm to a final height of 1 cm from the top of the core. Tensions were automatically recorded at 10 min intervals by the transducers which were connected to the TensioView software through a bus cable. Water content and fluxes were calculated by manually weighing the hyprop unit 3-4 times daily. Measurements were completed in 7-9 days in standard laboratory conditions (20-22° Celsius and 98-103 kPa atmospheric pressures). Cores were subsampled for use in pressure plate experiments outlined in section 2.4.4 of Chapter 2. Finally, soil was oven dried at 105° Celsius for 24 hrs to determine the final weight.

Experiments assumed that the water tension and water content decrease linearly from the bottom to the top of the column and the water flux and hydraulic gradient increases linearly between the two tensiometers (Schindler et al. 2010).

3.4.5.1. Unsaturated hydraulic conductivity calculation

Uncertainties arise when calculating the saturated hydraulic conductivity caused by small hydraulic gradients at low tensions. Errors caused by the nonlinear tensions in the late stage of the experiment were found to be small (Schinder et al. 2010). Using Poiseulle's Law of soil capillarity, the volumetric water content and matric potential associated with the water-filled pore effective diameter or hydraulic radius was determined as outlined by Hernandez-Ramirez et al. (2014). The calculated matric potential and corresponding volumetric water content are equal to one point on the moisture retention curve. Hydraulic conductivity associated with meso-pore diameters was calculated for unsaturated hydraulic conductivity. Soil meso-pore diameters of 50-100 um were used to determine the soil water matric potential (h_m) in hPa. The surface tension (γ) was 0.0728 N m^{-1}, solid-liquid contact angle ($\cos(\theta_{siv})$) was assumed to be 0° at saturation, density of water (P_w) was assumed to be 0.998 g cm^{-3} at 20° Celsius and g is constant at 9.81 m s^{-2}.

$$Pore\ diameter = 2\ x\ \left[\frac{2\gamma \cos(contact\ angle)}{P_w\ g|h_m|}\right] \quad [8]$$

The hydraulic conductivity is calculated according to Darcy's Law (Schindler et al. 2010), and may be written:

$$K = \frac{\Delta\, V}{2\ A\, \Delta\, t i_m} \quad [9]$$

where $\Delta\, V$ is the total evaporated water volume (cm^3), A is the cross sectional area of the sample (cm^2), $\Delta\, t$ is the change in time interval (sec) i_m is the mean hydraulic gradient (cm), given by:

$$i_m = \frac{1}{2}\left(\frac{\varphi_{ti,upper} - \varphi_{ti,lower}}{\Delta z} + \frac{\varphi_{t2,upper} - \varphi_{t2,lower}}{\Delta z}\right) \quad [10]$$

$\varphi_{ti,upper}$ and $\varphi_{ti,lower}$ are the lower and upper tensiometer values.

3.5. RESULTS

Soils ameliorated with deep ripping had higher infiltration rates at each tension (Table 3-1). Median infiltration rate increased with decreasing tension; infiltration increased from 0.65 x 10^{-4} cm min^{-1} to 2.49 x 10^{-4} cm min^{-1} for non-ripped soils and 0.94 x 10^{-4} cm min^{-1} to 3.00 x 10^{-4} cm min^{-1} for ripped soils. The greatest increase in infiltration occurred between -10 and -5 cm tension for both ripped and non-ripped soils. Water flux increased with decreasing tension as the size and volume of water conducting pores increased. The water flux was higher in the ripped treatment at all three tensions and shows an increased proportion of water conducting pores with diameters of 200-600 um. Ripping had the greatest effect on water conducting pores between

300 to 600 um. Non-ripped soils shows that compaction has reduced the water conducting pores at all tensions and the greatest reduction occurred between -10 and -5 cm. Over time, compaction has had less effect on larger water conducting pores than the smaller pores. Variability within each treatment is low; the greatest variability was observed at the -5 cm tension for both treatments.

A cumulative distribution function (CDF) of the measured saturated hydraulic conductivities was used to analyze the variability between and within ripped and non-ripped soils. A CDF gives values for the probability that the conductivity will be less than or equal to a specific value. Observed differences between the treatments show that median K_s measured with the falling head method was higher in the non-ripped treatment at the 5-10 cm depth but was lower at the 15-20, and 30-35 cm depths (Figure 3-1). The greatest differences between treatments were found in the 15-20 cm depth; median conductivity was three times higher in ripped soils - 0.037 cm min^{-1} in ripped and 0.011 cm min^{-1} in non-ripped soils. The CDF over the measured depths (0-35 cm) indicates no variation between ripped and non-ripped treatments (data not shown).

At the 30-35 cm depth, it would be expected that the non-ripped treatment would have a higher conductivity due to its lower bulk density than the ripped soils (Table 3-2). Lower conductivities at the 15-20 and 30-35 cm depths may be a result of dead-end structural pores in the non-ripped soils. Conductivity decreased with depth in ripped soils with the greatest difference between the 5-10 and 15-20 cm depth. This corresponds to the observed increasing bulk density with depth (Table 3-2). Conductivity of non-ripped soils decreased between the 5-10 and 15-20 cm depth and relatively unchanged to the 30-35 cm depth (Table 3-2).

Data from the evaporation experiment was used to plot the log-transformed conductivity as a function of tension/matric potential (hPa) and volumetric water content (cm^3 cm^{-3}) (Figure 3-2 and Figure 3-3). Results showed unsaturated hydraulic conductivity was not different between ripped and non-ripped soils (Figure 3-2). Unsaturated hydraulic conductivity plotted as a function of the volumetric water content shows that for conductivities greater than 10^{-5} cm min^{-1} ripped soils had a higher volumetric water content (Figure 3-3). Conductivities below 10^{-5} cm min^{-1} show volumetric water content was higher in the non-ripped soils at the 5-10 and 15-20 cm depth, but was not different at the 30-35 cm depth. Non-ripped soils showed a highly linear relationship between the volumetric water content and the logarithm of unsaturated hydraulic conductivity at all 3 depths. Ripped soils showed a more S-shaped curve with an inflection point around hydraulic conductivities of 10^{-5} cm min^{-1}; the curve flattened at this conductivity for the 5-10 and 15-20 cm depths.

Modelled moisture retention curves (MRC) revealed some differences between the pore size distributions of ripped and non-ripped soils (Figure 3-4). MRC for ripped soils show that the slope is relatively flat to approximately 30 to 60 hPa where the slope becomes linear until 1000 to 1500 hPa in the 5-10, 15-20, and 30-35 cm depths, respectively. This indicates the formation of discrete pore classes 2 to 100 um in diameter. Non-ripped soils had a more even distribution of pores as indicated by the flatter curve across the measured tensions.

Correlation analysis for both ripped and non-ripped soils over all measured depths shows that conductivity (or log conductivity) was positively correlated to all pore size classes with the exception of pores < 0.2 um (Table 3-3a). Most of the larger pore classes were positively correlated with each other. However, smaller pores showed opposite correlationships; for instance < 0.2 um were negatively correlated to all other pore fractions, and pores 0.2-50 um were negatively correlated to the large pores > 500 um. Ripped and non-ripped soils showed some differences in the correlation between K_s and the pore classes (Table 3-3b and 3-3c). Pore classes of 50-100, 100-500, and > 500 um in non-ripped soils and pore sizes 50-100 and 100-500 um in ripped soils had a significant positive correlation. Both soils showed a negative correlation for pores < 0.2 um to other pore classes; this was more significant across different pore classes in non-ripped soils. Results show K_s was negatively correlated to pores < 0.2 um in non-ripped and ripped soils.

In-situ (Guelph Permeameter) saturated hydraulic conductivity was analyzed using cumulative distribution functions (CDF). Results obtained for the saturated hydraulic conductivity showed some variation between ripped and non-ripped soils (Figure 3-5). In the 30, 45, and 60 cm depths the non-ripped soils have a slightly higher saturated hydraulic conductivity than ripped treatments. In the 15 cm depth the ripped treatments had greater variability indicating some of the cores have higher conductivities than non-ripped soils. The Guelph Permeameter is designed to make accurate readings of conductivity to 10^{-4} cm min^{-1}. In the 45 and 60 cm depths the soil permeability was too low for readings with the Guelph Permeameter.

All methods are seen to yield similar trends for the measurement of saturated hydraulic conductivity for ripped and non-ripped soils (Figure 3-6). Differences between the conductivity in ripped versus non-ripped soils were often negligible. Comparing the results from the methods revealed some differences. In general, the conductivity calculated with the hyprop is one and two orders of magnitude greater than the falling head and Guelph Permeameter measurements, respectively. Greater variability was observed in the Guelph permeameter and falling head measurements with values ranging over three orders of magnitude (Table 3-5).

3.6. DISCUSSION

3.6.1. Comparison of hydraulic methods: field and laboratory

Data showed some differences in the measured saturated hydraulic conductivity amongst methods (Figure 3-6 and Table 3-5). In-situ and ex-situ methods may provide different results in the saturated hydraulic conductivity and often the correlation between these methods is low (Buckzo et al. 2006). Comparative studies have shown mixed results between methods (Lee et al. 1985; Mohanty et al. 1994; Paige and Hillel 1993; Richard et al. 2001v; Stolte et al. 1994).

Multiple methods can be used to assess the hydraulic properties of soils. The Guelph Permeameter (GP) is a useful, in-situ tool to measure the saturated hydraulic conductivity. However, stable readings may only be an approximated value due to slow declining rates of water fall (data not shown). In low conductivity soils, the decline in water level in the instrument reservoir may be too slow to obtain accurate measurements of true steady state conditions to obtain K_s with the GP. This may have caused some discrepancies between the falling head (FH) and GP methods in our experiments. Conductivity measurements made with the GP include both vertical and horizontal conductivity of water whereas water flow is forced to be entirely vertical in FH measurements. Compaction causes an increase in the horizontal planar pores; therefore, non-ripped soils may have a higher conductivity due to higher flow of water in the both directions. In our falling head experiment, flow is predominantly vertical non-ripped soils would have a lower conductivity because horizontal flow is restricted by the core.

Guelph permeameter measurements are subject to errors arising from preparation of the borehole well in which smearing can occur. Where long measurement intervals occur there is the risk of deposition of particles into pore spaces (Mohanty et al. 1994). Air entrapment is not accounted for in field measurements which may induce differences between lab and field measured K_s. Field measurements often have lower values than those obtained in the lab (Bodner et al. 2013). This was evident in our results as the GP measurements were lower than both the FH and Hyprop measured conductivities. Also, it may be hard to discriminate between differences in spatial variability and the methods since GP measurements were not conducted in the same locations where cores were excavated from (Stolte et al. 1994).

The Hyprop measures the evaporation rate and hydraulic gradient to quantify the saturated hydraulic conductivity. Near saturation the hydraulic gradients are often too low to obtain a true measure of the hydraulic conductivity. Therefore, saturated and near saturated hydraulic conductivities calculated with the hyprop may overestimate the K_s as the data is extrapolated to zero tensions (to saturation).

Field methods are often preferred to lab methods as true soil conditions are captured. Macro-pore topology may be damaged during soil core extraction and may not be representative of field conditions. Soil cores may reflect greater variability than field methods as some cores may be inclusive to macro-pores whereas others are not. Our results show that variability between the FH and GP was similar for both ripped and non-ripped soils. However, in our study analysis was based on the results of the falling head experiment because the prevailing conditions (humidity, soil moisture, air temperature, and atmospheric pressure) were held relatively constant for all analyses. Field conditions were variable over the course of the experiment and could result in some variability between other measured parameters. To make accurate comparisons of the effects of the ripping treatment, as many variables as possible are carefully controlled to ensure that the results were caused by the treatment and not extraneous factors.

Despite some differences amongst methods, the results do not show large differences in K_s between the ripped and non-ripped treatments. It should be noted that most differences were observed in the lower quartile of the CDF between the FH and GP methods. In general, this observation can indicate a good agreement in the results of the ripped and non-ripped treatment for each method. Treatment differences of K_s will be based on FH measurements in the next section.

3.6.2. Ripping longevity effects on water movement and soil pores

Results show that ripping is having a positive effect on the infiltration rate of the soil four years after its application (Table 3-1). Ripping was likely able to increase the cracks or fissures at the surface that promotes macro-pore development. In addition, variations in temperature and volumetric water content are generally greatest at the soil surface. In this experiment the climatic factors helped maintain the increased porosity from ripping. Creation of soil cracks from ripping allows for greater frost action and shrinking-swelling than in non-ripped soils because of increased infiltration into surface cracks. Soil temperature in the 5 cm depth showed fairly consistent results across 4 years. In addition to spring thaw and fall freeze, soil temperatures fluctuated above and below 0° Celsius only 5 times over since the ripping treatment (data not shown). Also, root development in the created cracks, could likely aid in extending the longevity of the effects of ripping on macro-pores through maintaining and contributing to soil fracturing. As well, roots exert pressure on soil particles increasing the binding between individual particles, thereby increasing aggregate strength and structural formation at the site (Chen et al. 2014). Chong and Cowsert (1997) found that infiltration of a soil ameliorated with deep tillage to 80 cm decreased with time, although the improvements were still detected for 3 years. Improved surface porosity can reduce surface runoff and erosion in deep ripped soils (Sojka et al. 1993). Non-ripped soils show a higher percentage of large water conducting pores (Table 3-1). Compaction causes reductions in these pore sizes but surface pore development in compacted soils can be affected by natural processes (Bottinelli et al. 2014). Porosity at the surface of non-ripped soils may be the result of drying shrinkage around coarse fragments (i.e., rocks and roots). Guebert and Gardner (2001) found coarse fragments at the soil surface to be important pathways for the development and connectivity of macro-pore networks. Macro-pores at the surface increase surface storage and flow to greater depths which reduces overland flow and surface runoff. Infiltration in a minesoil was shown to increase after 3 years following reclamation in a silt loam soil as a result of fragments and roots in the surface soil (Guebert and Gardner 2001).

In this experiment coarse fragments (i.e., rocks and roots) may have improved large pore development of non-ripped soils by promoting physical processes. Differential expansion and contraction, as a result of temperature fluctuations, between adjacent coarse fragments at the surface are believed to be responsible for macro-pore development. This was evident in the 5-10 cm depth as well where results showed saturated conductivity measured with the falling head method to be higher in the non-ripped soils, although the differences between the treatments was small (Figure 3-1). Previous, studies have shown surface compaction (< 10 cm) may be alleviated through shrink-swell cycles and biological activity (Bradshaw 2000; Etana et al. 2013; Evans et al. 1996). Kreyling et al. (2007) defined one freeze-thaw cycle as soil temperatures crossing 0° C twice for a 48 hour period above and below 0° C. Accordingly, our results show that although freeze-thaw may have contributed to improved conductivity, it was likely not the only factor in compaction alleviation in non-ripped soils (Figure 3-1) as discussed above. Ripping can cause the conductivity to be lower than it was prior to the ripping treatment (Dexter et al. 2004). This has been attributed to a reduction in soil strength and loss of meso-structure by re-compaction following deep tilling. In our experimental site, ripping may have reduced the soil strength making the surface layer more susceptible to re-compaction by consolidation.

Differences between the water flux and infiltration rate of ripped and non-ripped soils were small and low variability was measured within each treatment at the soil surface (Table 3-1). Positive effects of deep tillage on infiltration were negated over time likely due to surface crusting when the vegetative cover is low (Chong and Cowsert 1997). In our experimental site, soil crusts may have developed in both ripped and non-ripped soils creating homogenous surface conditions for both treatments. Surface crusting and sediment deposition into the surface pores may be causing reduced pore variation and infiltration (Paglial et al. 2004). Poorly aggregated soils have weakened particles bonds; rain impact can cause the displacement and deposition of soil particles into surface pores. Further, site activities that required equipment and foot traffic including planting and weed management may have caused some degree of compaction at the soil surface.

Surface controlled infiltration may result in both ripped and non-ripped soils when the delivery rate of water exceeds the infiltration rate of the soil. The saturated conductivity in both soils is greater in the 5-10 cm depth than the infiltration rate of the surface (Figure 3-1 and Table 3-2). A hydraulic barrier or bottleneck may develop which impedes infiltration into the profile (Hillel 1998). Further, reaching saturation may require longer periods of time or replenishment of the root zone may not occur. In our experiment, ripping increased the infiltration rate, and hence the rainfall intensity that is required to cause surface limiting conditions will in theory be also higher. Surface roughness is also increased with ripping and during surface controlled infiltration uneven surfaces may promote increased surface-storage capacity (Sillon et al. 2003). Ripping may increase ponded infiltration thereby reducing surface runoff at the site.

Longevity of ripping on soil K_s is evident in the 15-20 and 30-35 cm depths, with the greatest improvements in the 15-20 cm depth (Table 3-2). Increased conductivity may promote deeper water percolation at the site, which would increase the available water during drier periods where stored surface water (< 15-25 cm) is lost through evapotranspiration (Brady and Weil 2002). Improved percolation and redistribution would increase water recharge in the root zone. In our study, it is expected that greater conductivity would improve aspen growth by diminishing incidence of extreme conditions such as plant water stress and waterlogging conditions. Similar results were found by Travis et al. (1990) who found that deep plowing (76 cm depth) on a saline silty clay loam soil improved soil water dynamics (distribution, drainage, and infiltration). Soil water replenishment was more evenly distributed in ripped soils and the water table had irregular depressions and peaks. This indicated deep ripping had increased soil porosity which enhanced water transmission properties. Hydraulic barriers in non-ripped soils may prevail when water percolates and its flow becomes restricted at the 15-20 cm depth. Increased lateral flow may cause saturation and reduced aeration above this depth (Table 3-2). Ripping at the site did not improve the bulk density in the 30-35 cm depth, but the higher conductivity may indicate a few hydraulically important macro-pores responsible for the conductivity being twice as great as the non-ripped soils. This is supported by the positive correlation between K_s and pore classes >50 um (Table 3-3c). This may indicate greater connectivity and capillary flow in these pores. These results show similar trends to that of Drewry et al. (2000). In their study, surface settling caused no differences in the saturated conductivity of the 0-18 cm layer after 2.5 years. Conductivity was increased in deep ripped (ripping depth of 25-30 cm) plots in the 18-24 cm depth by up to two orders of magnitude in the silt loam soil. Subsoiling increased the connectivity of macro-pores in subsurface layers.

Natural processes are less effective at ameliorating compacted subsoils (McNabb 1994). McNabb (1994) suggested that Alberta winters are not sufficiently cold or variable enough to induce changes to subsoil properties and it may require decades for improvements by natural attenuation. This is shown in the non-ripped soils in the 15-20 and 30-35 cm depths (Figure 3-1). Our results of bulk density and void ratio closely correspond to the saturated hydraulic conductivity for non-ripped soils (Table 3-2), with the exception of the 30-35 cm depth. The lower bulk density did not result in a higher conductivity. Relict structural pores may be causing this trend. These pores are only accessible through the necks of lacunar pores and may not contribute to the water flow (Sillon et al. 2003). As well, saturated hydraulic conductivity is more influenced by macroporosity (> 100 um diameter) volume which was greater in ripped soils in the 30-35 cm depth. This is supported by the negative correlation of pores 100-500 um and > 500 um to the K_s at this depth ($r = -0.028$, $P < 0.921$; $r= -0.066$, $P < 0.82$, respectively). As expected the correlation between residual pores and K_s was low in both soils, as water in these pores is adsorptive rather than capillary and water flow is confined to thin films along the outer pore wall (Table 3-3a-c). Previous research has shown K_s of loam soils to be 0.107 cm min^{-1} (Kargas and Londra 2015) and 0.019 cm min^{-1} (Arvidsson 2001). According to these values, K_s for ripped soils at all three depths and non-ripped soils in the 5-10 cm depth are comparable to natural loam soils.

Unsaturated hydraulic conductivity was not different between treatments at any depths (Figure 3-2). But the volumetric water content was higher in ripped soils above conductivities of 10^{-5} cm min^{-1} (Figure 3-3). According to these results, ripping may have changed the volume, size and geometry of pores in the soil. Over time, the size of the pore necks may have decreased without changing the soil porosity (Richard et al. 2001a). Unsaturated conductivity would be similar to prior to ripping but the volumetric water content would remain higher in the near saturated zone. Non-ripped soils show evidence to suggest there is greater pore connectivity and reduced tortuosity above 10^{-5} cm min^{-1} or pores with a higher hydraulic radii. The shape of the moisture retention curves (MRC) indicates variability in the pore size distribution is affecting the unsaturated hydraulic conductivity (Figure 3-4). Pore distribution at tensions of approximately 50 to 100 hPa for ripped soils is small causing the slope of the MRC to be less steep. This corresponds to the unsaturated conductivity curve flattening at 10^{-5} cm min^{-1} because as these pores fill they do not contribute to an increase in the conductivity. Overall, pore size distribution was more evenly distributed for non-ripped soils at all three depths in our study as evidenced by a lower slope in the MRC and linear unsaturated and volumetric water content curve.

Discrete pore size distributions are evidenced in the ripped soils which may indicate soil structural development. Hierarchical pore development may be responsible for the steeper slope of the MRC. Ripping may be causing the formation of pores 2-100 um as indicated by the steeper slope on the MRC at approximately 30 to 1500 hPa (Figure 3-4). This is also observed in the correlation matrix which indicates that aggregate development may be occurring. Cracks (> 500 um) are negatively or poorly correlated to transmission pores (50-500 um); micro-aggregates may be forming as particles become more closely packed and there is a loss of cracks and increased inter-aggregate pore space. Ripping at the site may have promoted fissures attached to the crack openings along natural planes of weakness, as drying-shrinkage and biological processes occur this may cause increased aggregate formation in and around the large openings (> 500 um). Inter-aggregate pore spaces (50-500 um) may be better for soil quality as more water is retained and available for plant growth compared to cracks which can act as preferential pathways that could lead to loss of water or leaching nutrients (Etana et al. 2013). Although not always an occurrence, a reduction in storage porosity may be coupled with an increase in residual pores or vice versa. Residual pores may increase in diameter through microbial activity (i.e., burrowing or decomposition). This is supported by our finding of non-significant correlation between residual pores (< 0.2 um) and storage pores (0.2-50 um) (Table 3-3). In our study, soil structural development as a result of ripping is likely responsible for the decline in the slope of the conductivity and water content curve, indicating the formation of distinct pore sizes at these water contents.

Non-ripped soils have a fairly even distribution of pores as compaction causes a relatively homogenous pore system and there is a loss of large inter-aggregate pores and increase in medium size pores (Hillel 1998). Results show pore variability is greater in non-ripped soils. The significant positive correlation between pores > 50 um may be explained by how large pores develop via biophysical processes. Once plant roots establish in reconstructed soils this can promote faunal activity which can facilitate larger pore development. A decrease in the inflection point on the MRC of non-ripped soils may be the result of plant and microbial activity, where there is increased distribution of finer (< 500 um) pores. This result is supported by Daynes et al (2013) who found pore size distribution of soils with seeded plants and fungi to cause a decrease in the slope of the MRC. Bottinelli et al. (2014) explains that large macro-pores (> 240 um) will regenerate when faunal activity recovers.

3.7. CONCLUSIONS

The objective of this chapter was to assess the longevity of a subsoil ripping treatment on the dynamic soil water properties of a reconstructed soil following the disturbance associated with surface mining. Results show that improvements to hydraulic properties were evident four years after ripping to a 0.60 m depth in the 15-20 and 30-35 cm depths with the greatest improvements at 15-20 cm. Results show that soil structural development may be occurring in ripped soils as indicated by the MRC, and conductivity and volumetric water content curves. Both show that abundance of pore sizes 50-500 um may be increasing. Our study found evidence for the presence of hydraulic barriers as surface soil infiltration rates were lower than the surface layer; this was further accentuated by a further reduction in conductivity in the sub-surface layer (15-20 cm).

Deep ripping with a heavy duty rip plow is a method to improve the soil water dynamics on reconstructed soils when aiming at alleviating subsoil compaction. The increased water transmission, as a result of increased transmission pores (50-500 um), in the subsoil could suggest greater water redistribution and drainage. More importantly, increased water flow to deeper layers in the soil profile could potentially increase plant available water throughout the growing season. Structural development improves pore continuity and connectivity which may subsequently feedback into greater biological activity. In theory, biological activity would be expected to increase in the ripped soils further promoting aggregate development and stability. This would hypothetically lead to greater nutrient cycling and organic matter decomposition, further improving soil quality for aspen regeneration. Although infiltration rates are lower at the surface than below the surface layer, ripping can also create rough surface conditions that would trap more rainwater and snow as well as limit surface runoff compared to non-ripped soils. It would therefore be expected that erosion potential would be decreased by ripping the soil. Future research can address these various mechanistic hypotheses.

To improve the longevity of deep ripping, the use of a flocculating agent (i.e., gypsum) may be appropriate and further supplemented with addition of organic material (Hamza and Anderson 2003). Likewise, limiting site activities following deep ripping can reduce surface re-compaction.

3.8. TABLES

Table 3-1. Median water flux (cm^3 cm^{-2} min^{-1}) and infiltration rate (cm min^{-1}) for ripped and non-ripped soils at the soil surface.

Treatment	Tension (cm)	Equiv. pore diameter (um)	Water flux (cm^3 cm^{-2} min^{-1})	Water flux range (cm^3 cm^{-2} min^{-1})	Infiltration rate (10^{-4} cm min^{-1})	Infiltration rate range (10^{-4} cm min^{-1})
Non-ripped (n = 18)	-15	200	0.78	0.48 – 2.11	0.65	0.55 – 1.31
	-10	300	1.17	1.08 – 2.75	1.35	0.88 – 2.40
	-5	600	2.79	1.62 – 4.61	2.49	1.39 – 4.58
Ripped (n =18)	-15	200	0.99	0.62 – 2.33	0.94	0.60 – 1.82
	-10	300	1.49	1.06 – 3.76	1.68	1.05 – 3.07
	-5	600	3.14	1.92 – 6.23	3.00	1.84 – 5.62

Table 3-2. Median saturated hydraulic conductivity (cm min^{-1}), mean and standard error void ratio, bulk density (g cm^{-3}) and median macro-pore volume (cm^3 cm^{-3}) for ripped and non-ripped soils

Depth (cm)	Treatment	K_s (cm min^{-1})	Void ratio	Bulk density (g cm^{-3})	Macro-pore volume (> 100 um)[a]
5-10	NR	0.075	1.34±0.047	1.14±0.029	0.050
	R	0.052	1.26±0.075	1.18±0.046	0.048
15-20	NR	0.011	1.00±0.075	1.34±0.046	0.035
	R	0.037	1.09±0.059	1.28±0.036	0.041
30-35	NR	0.012	1.05±0.052	1.31±0.032	0.039
	R	0.025	0.92±0.058	1.39±0.035	0.041

[a] Macro-pore volume is expressed as the median volume fraction of pores with a diameter greater than 100 um

Table 3-3a. Correlation matrix for combined ripped and non-ripped soils for saturated hydraulic conductivity and equivalent pore diameter size classes (um) measured with the falling head method for all 3 depths.

	n	< 0.2	0.2-50	50-100	100-500	> 500
K_s (cm min^{-1})	98	-0.15	0.16	0.35**	0.38**	0.27**
< 0.2	102	1	-0.46**	-0.21*	-0.32**	-0.17
0.2-50	102		1	0.039	0.16	-0.31**
50-100	102			1	0.40**	0.24**
100-500	102				1	0.11
> 500	102					1

* significant at 0.05, ** significant at 0.01

Table 3-3b. Correlation matrix for non-ripped soils for equivalent pore diameter size classes (um) and the saturated hydraulic conductivity measured with the falling head method for all 3 depths.

	n	< 0.2	0.2-50	50-100	100-500	> 500
K_s (cm min^{-1})	48	-0.24	0.35**	0.40**	0.41**	0.28
< 0.2	50	1	- 0.36*	-0.25	-0.44**	-0.34*
0.2-50	50		1	-0.0017	0.17	-0.16
50-100	50			1	0.39**	0.28*
100-500	50				1	0.41**
> 500	50					1

* significant at 0.05, ** significant at 0.01

Table 3-3c. Correlation matrix for ripped soils for equivalent pore diameter size classes (um) and the saturated hydraulic conductivity measured with the falling head method for all 3 depths.

	n	< 0.2	0.2-50	50-100	100-500	> 500
K_s (cm min^{-1})	50	-0.019	-0.081	0.28	0.29*	0.25
< 0.2	52	1	-0.54**	-0.15	-0.21	-0.028
0.2-50	52		1	0.062	0.17	-0.46**
50-100	52			1	0.41**	0.16
100-500	52				1	-0.26
> 500	52					1

* significant at 0.05, ** significant at 0.01

Table 3-4. Range of values for saturated hydraulic conductivity (cm min^{-1}) for the falling head method (FH) at 3 depths (5-10, 15-20, and 30-35 cm), Hyprop (HY) at 3 depths (5-10, 15-20, and 30-35 cm) and Guelph Permeameter (GP) at 2 depths (15 and 30 cm).

	Method					
	Falling Head		**Hyprop**		**Guelph Permeameter**	
Depth	**NR**	**R**	**NR**	**R**	**NR**	**R**
5-10	0.0027 - 0.35	0.0078 - 0.30	0.054 - 2.09	0.0012 - 0.19	-	-
15-20	0.00069 - 0.17	0.00040 - 0.54	0.063 - 8.90	0.056 - 4.0	0.00014 - 0.080	0.00019 - 0.25
30-35	0.0048 - 0.41	0.00087 - 0.26	0.05 - 1.53	0.066 - 1.23	0.0002 - 0.085	0.000073 - 0.050

3.9. FIGURES

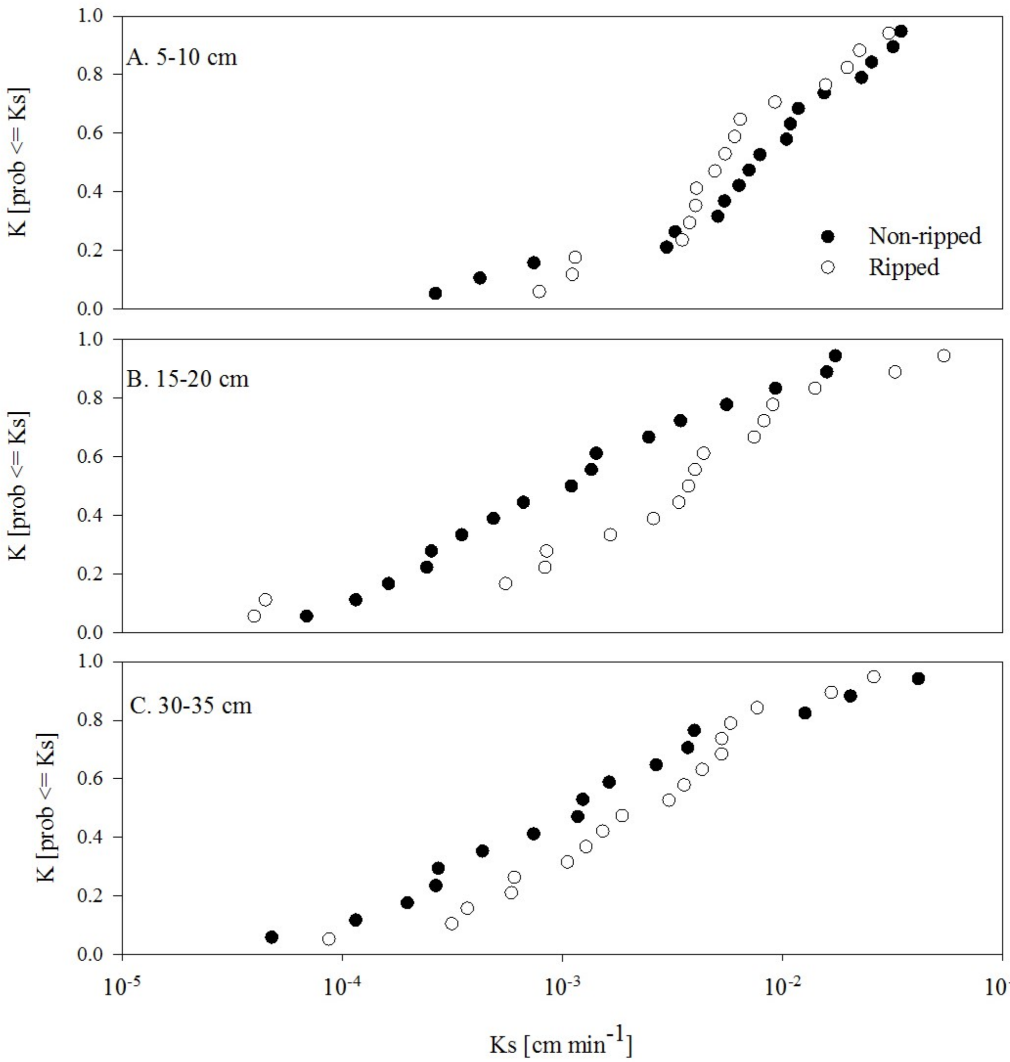

Figure 3-1. Saturated hydraulic conductivity (cm min^{-1}) for ripped and non-ripped soil treatments at 3 depths (A. 5-10, B. 15-20, and C. 30-35 cm) measured with the falling head method.

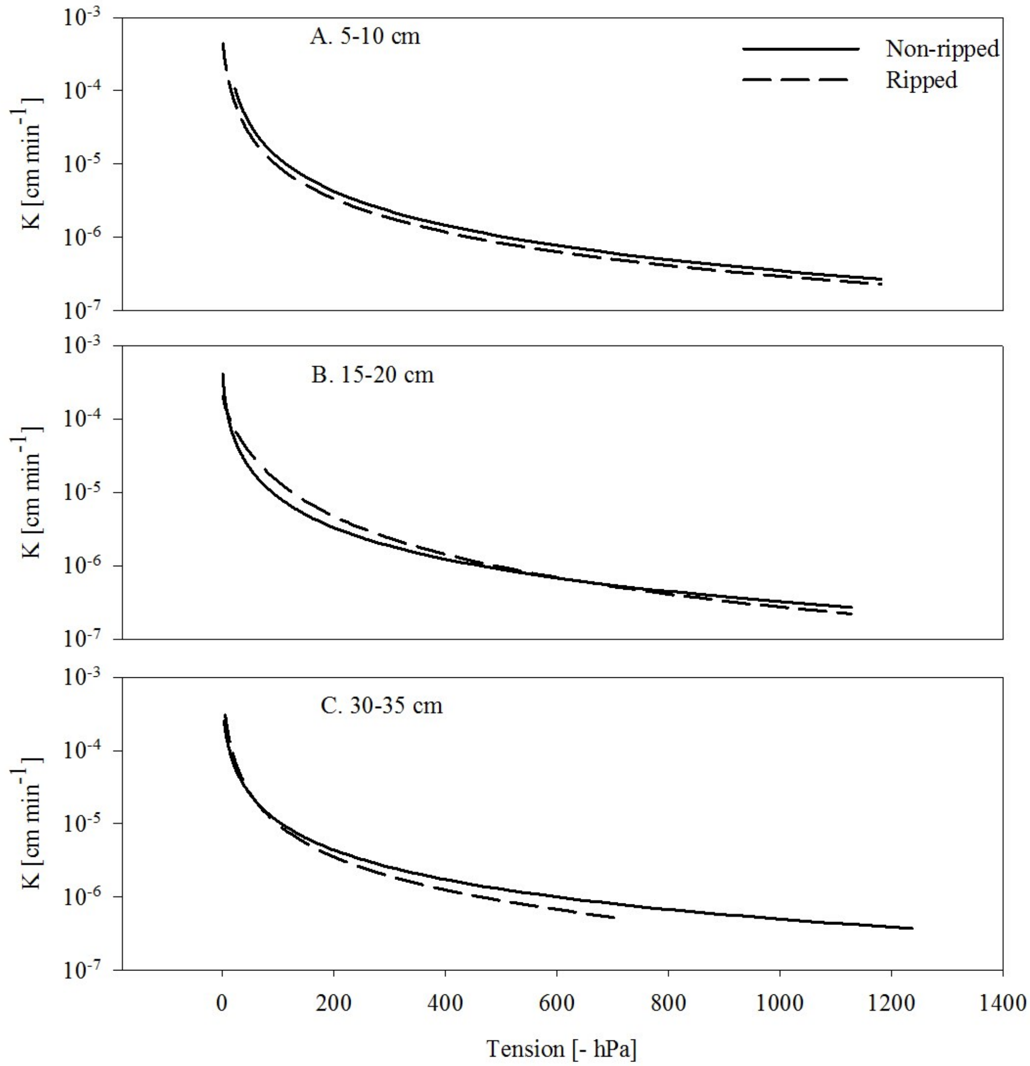

Figure 3-2. Hydraulic conductivity (cm min^{-1}) curve for ripped and non-ripped soils at 3 depths (A. 5-10, B. 15-20, and C. 30-35 cm). Fitted to the van Genuchten-Mualem model.

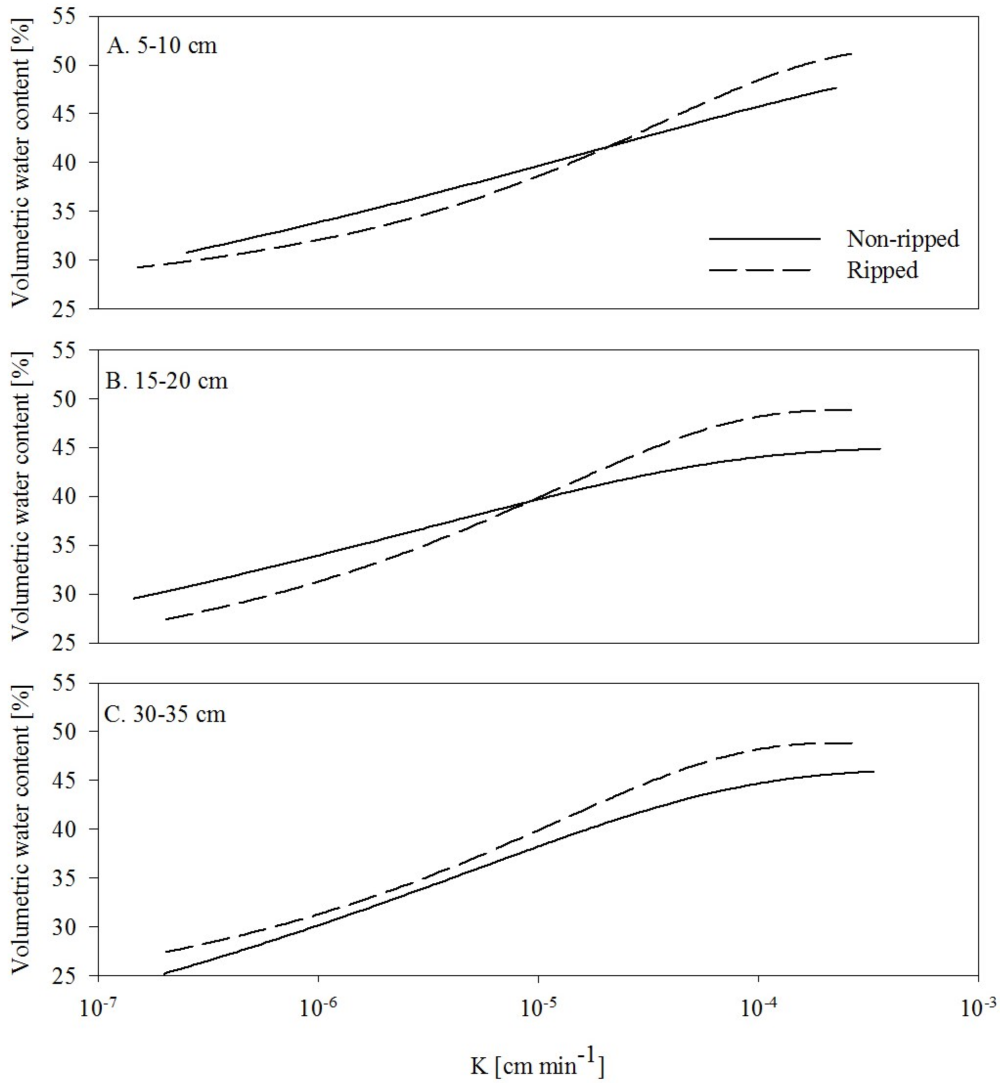

Figure 3-3. Hydraulic conductivity (cm min^{-1}) as a function of volumetric water content for ripped and non-ripped soils in 3 depths (A. 5-10, B. 15-20, and C. 30-35 cm) fitted to the van Genuchten-Mualem model.

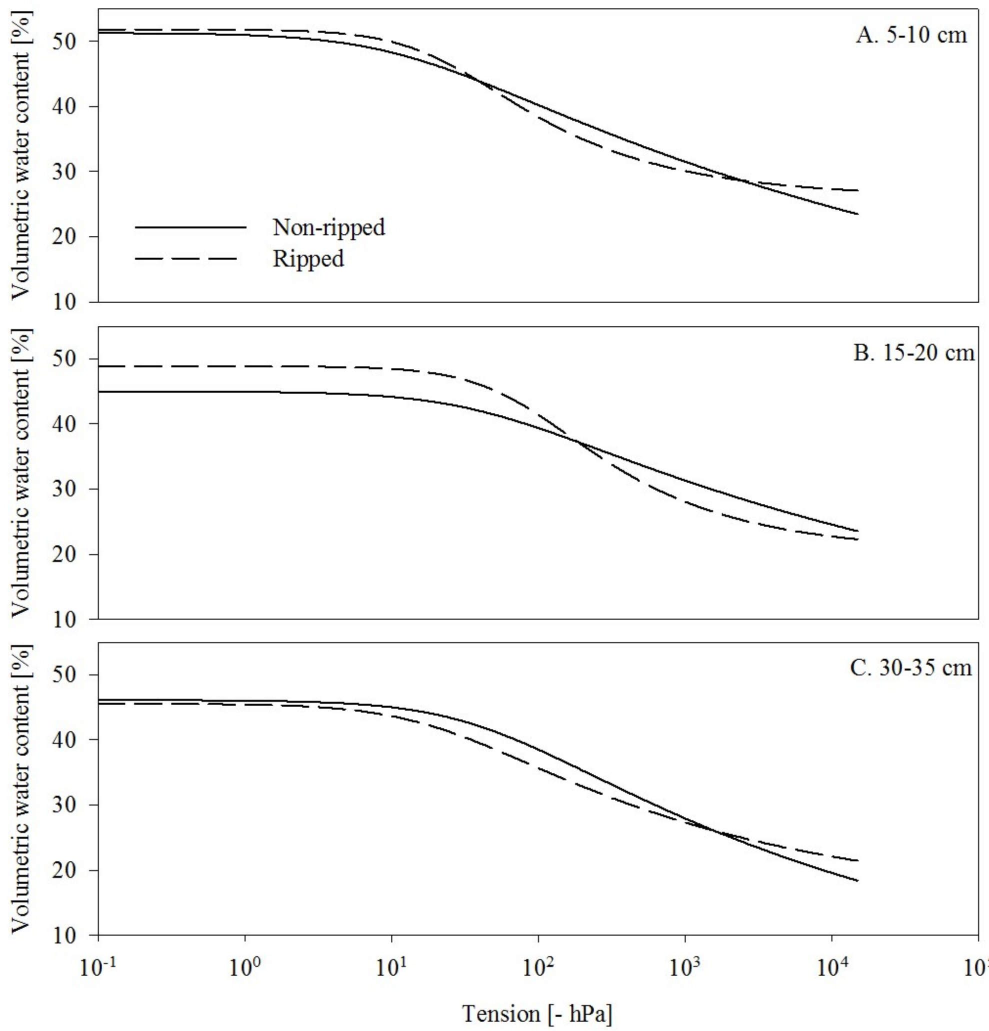

Figure 3-4. Soil moisture retention curve for ripped and non-ripped soils in 3 depths (A. 5-10, B. 15-20, and C. 30-35 cm) fitted to the van Genuchten model for moisture retention.

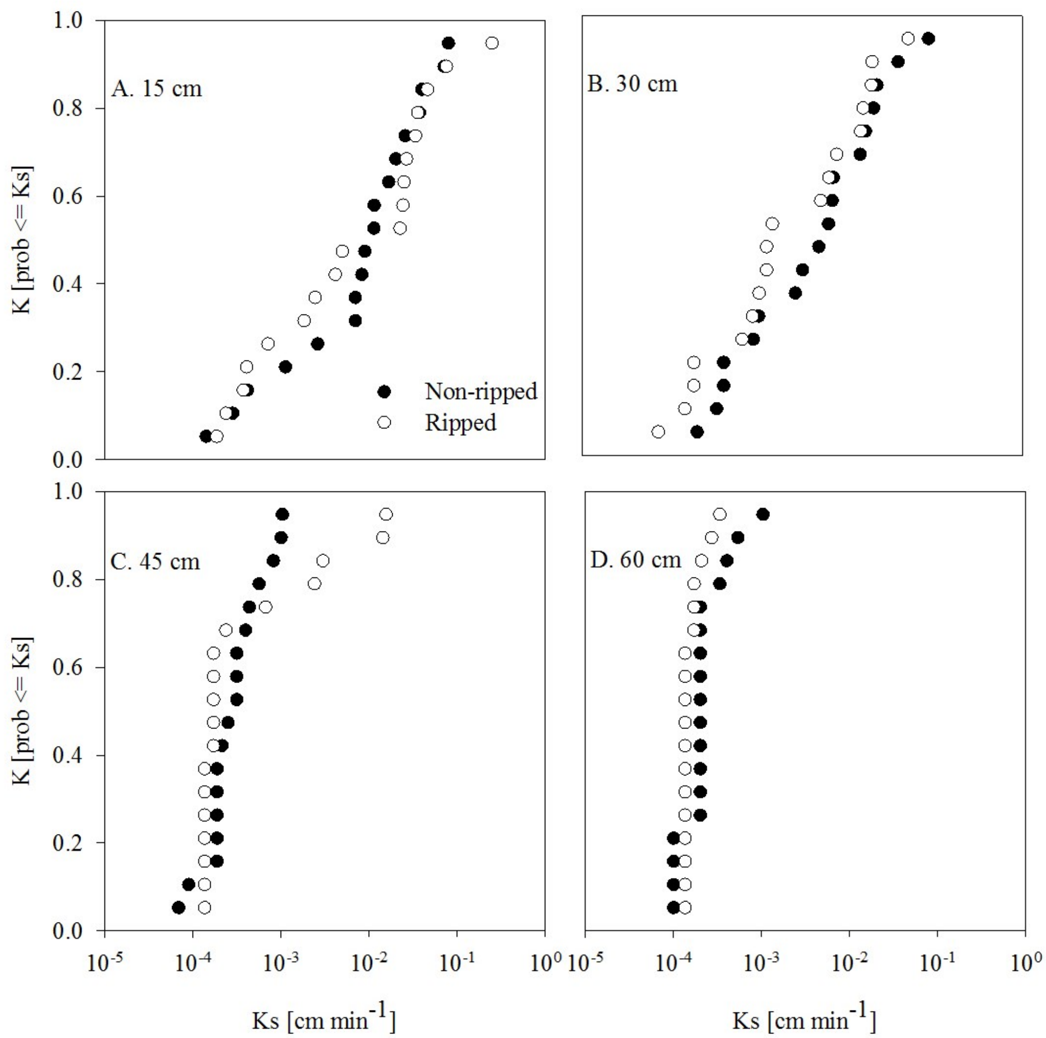

Figure 3-5. Field saturated hydraulic conductivity measured with the Guelph Permeameter at four depths (A. 15, B. 30, C. 45, and D. 60 cm).

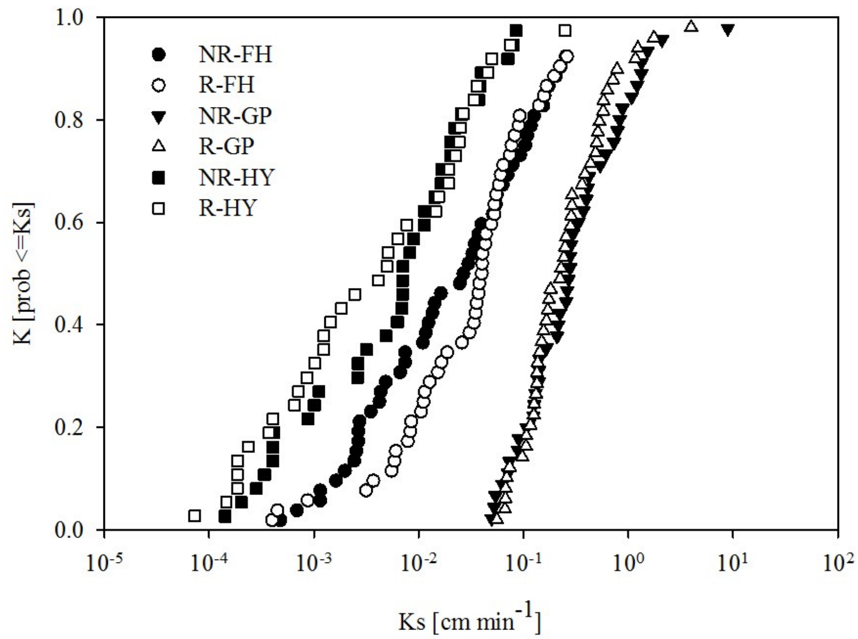

Figure 3-6. Cumulative distribution function curves of saturated hydraulic conductivity (K_s) for ripped and non-ripped soils measured with the falling head (FH) method at 3 depths (5-10, 15-20, and 30-35 cm), Hyprop (HY) at 3 depths (5-10, 15-20, and 30-35 cm) and Guelph Permeameter (GP) at 2 depths (15 and 30 cm).

4. CHAPTER 4 – GENERAL CONCLUSIONS

4.1. RESEARCH SUMMARY

Quantification of the vertical spatial (i.e., stratification with depth) and temporal variations in soil properties is necessary to evaluate the effects of subsoiling on reconstructed soils. The two main objectives of this research include: (1) assessing soil physical properties including pore size distribution and bulk density, to determine the medium-term (~ 4 yrs) influence of ripping on soil water storage; (2) determine the medium-term effects of ripping on dynamic soil hydraulic properties (i.e., hydraulic conductivity, infiltration rates). Overall, this research aims to inform land reclamation efforts as to whether deep ripping can effectively improve soil quality for re-vegetation purposes by increasing water infiltration, re-distribution, storage, drainage, and reducing evapotranspiration.

- Ripping induced changes to the moisture retention curve by reducing the volume of water retained in the soil with increasing tension from approximately 30/60 to 1000/1500 hPa. The observed differences in the moisture retention curves between treatments shows an increase in the slope of the ripped soils curve indicating the formation of discrete pore classes 2 to 100 μm diameter.
- Greater saturated water contents were observed in ripped soils in the 5-10 and 15-20 cm depths. Ripping likely increased the macroporosity of the ripped soils at these depths.
- Ripping affected the air entry potential of soils by increasing the volume of macro-pores in the 15-20 and 30-35 cm depths; this may indicate greater O_2 diffusion reducing root zone hypoxia.
- Over time there has been a simultaneous decrease in soil bulk density in non-ripped soils with no improvement or some increase in bulk density of ripped soils. However, our results show that bulk density may not be the most sensitive indicator to assess the medium-term effects of deep ripping with respect to determining changes in pore size distribution and saturated hydraulic conductivity which has a more significant influence on water storage and transport.

- Mass fractal dimensions were not significantly different indicating no hierarchical aggregate development in the compacted and non-compacted (deep ripped) soils. Low organic matter in the soil may be limiting aggregation.
- Compacted and non-compacted soils show changes in bulk density and pore size distribution may influence the parameters of the moisture sensor calibration curves.
- Deep ripping increased the infiltration rate of the soil by increasing the volume and proportion of large pores (> 200 μm diameter) at the soil surface thereby hypothetically reducing surface limiting conditions of ripped soils.
- Saturated hydraulic conductivities were improved in subsurface layers (15-20 and 30-35 cm) with ripping which is believed to be the result of increased pore continuity, increased pore hydraulic radii and volume or the presence of a few hydraulically important macropores.
- Results from this research indicate some medium term benefits of subsoil ripping on soil physical and hydraulic characteristics in compacted reconstructed soils. Over all, medium term effects of ripping are most evident in the subsurface layers (15-20 cm) as shown by greater saturated conductivity and water storage. Results from this research suggest that medium-term benefits are apparent from deep ripping reconstructed soils and soil conditions are improved for forest re-vegetation.
- Surface compaction (< 10 cm) amelioration may be reduced by physical and biological processes that aid in increased macro-pore formation and reduce surface bulk density.

4.2. RECOMMENDATIONS AND APPLICATIONS

Based on this research it would be recommended, as a reclamation practice, to utilize deep ripping to improve soil properties of compacted soils to greater depths. Non-ripped soils showed some improved conditions; however, subsoil compaction was evident and poor soil quality would likely hinder revegetation in non-ripped soils. Ripping has likely increased the rate at which soils will naturally ameliorate compaction by promoting structural formation. This has increased the infiltration rate as well as reducing hydraulic barriers which alters soil water dynamics.

It is recommended that to increase the longevity of ripping increased ripping depth and wider spaced shanks should be implemented. This can increase the volume of the soil and allow increased time to allow for increased soil strength and particle aggregation to greater depths.

To prevent surface compaction and reduce the potential for surface crusting organic amendments can be added to the surface layers to enhance aggregate formation, water holding capacity and prevent erosion and surface degradation by rainfall impact when a vegetative cover has not been established.

It is also recommended to limit site activities immediately following subsoil ripping as the soil is especially susceptible to re-compaction and settling within the first couple years (< 2 yrs). This includes limiting tillage operations, or use of heavy equipment for site management.

Minimizing the degree of compaction must be considered during reconstruction of soils. This includes timing activities to when the soil is drier than field capacity; selection of equipment with tracks to more evenly distribute the load and when possible use equipment with smaller loads; limit the number of passes made with equipment; minimizing soil loosening; and add organic matter to maintain the effects of ripping.

4.3. FUTURE RESEARCH AND RESEARCH LIMITATIONS

Further research is required to fully understand the influence of subsoil ripping on the physical and hydraulic properties of reconstructed soils. Long term monitoring is necessary to assess whether the ripping treatment had an influence on the trajectory of soil properties of compacted subsoil. Results from this study indicate some long term influence of ripping on soil physical properties and water characteristics. Potential areas for future research include:

- Conduct long term research (> 10 yrs) to assess differences in ripped and non-ripped soils. Results may indicate if significant re-compaction has occurred on ripped plots and determine if biological and physical processes are improving subsoil conditions of compacted non-ripped soils.
- The impact of ripping on the fractal dimension of surface soils (> 15 cm depths) to further our knowledge on the role ripping has aggregate hierarchy.

- Understanding the role ripping has on organic matter decomposition and aggregation, through determining water stable aggregates formation.
- Experimenting with the use of a flocculating agent (i.e., gypsum) on structural formation and longevity of ripping on compacted soils.
- Understanding the dynamics freeze-thaw and drying-shrinkage cycles on compacted and non-compacted soils, and determining whether ripping increases the number of cycles occurring per year.
- Comparing soil properties at greater depths (> 35 cm) to determine saturated hydraulic conductivity using different laboratory methods.
- The effect of ripping on pore morphology (i.e., hydraulic radii and geometry) through image analysis
- Conduct experiments on soil air permeability
- Monitoring erosion and sediment deposition in sloped landscapes in ripped soils
- Completing an economic assessment on costs of ripping and the value of improved vegetative growth

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