Considering groundwater recharge and flow in urban development planning

A case study from Torshovdalen (Oslo, Norway)

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Abstract

Humans have been altering the environment for centuries. Expanding cities and urbanization is yet another constant change in a dynamic environment in the advent of Climate change. Recent research has focused on the hydrogeological responses to climate change in urban areas, and the use of site assessment and modified Thornthwaite water balance as a tool for storm water management.

This project addresses the potential impact of climate change on groundwater recharge and flow in Torshovdalen and its associated catchments in Oslo, Norway. This case study is a part of a larger interdisciplinary project initiated by the Agency for Water and Sewerage works in the municipality of Oslo which encompassing disciplines such as engineering, landscape architecture, economy, and hydrogeology.

Site assessment of Torshovdalen was performed in form of geotechnical investigation, infiltration testing and field observation to explore the geological units and underground structures (i.e. heterogeneity below ground), infiltration capacity and hydrological response. A water balance was estimated for both Torshovbekken, and Torshovdalen catchment to evaluate necessary implementations of stormwater solutions. Previously recorded data, and new field data was collected to improve the geological understanding and define hydraulic properties of the unsaturated and groundwater zones.

Site assessment suggests a shallow sediment layer of 0-3 m which is covering the Cambro-Silurian formation in the northern half of the valley and demonstrates an average silty loam texture. Southern half of Torshovdalen revealed a much deeper sediment-bedrock interface of ~13 m. Here, soil textural properties exhibited a sandy loam texture in the upper 0-4 m, and a silty clay texture in the lower 4-6 m. Infiltration tests demonstrated a mean rate of 3-30 cm/h which suggests that the silty loam affect the hydraulic properties and thus do not infiltrate enough water.

Estimated water balance of Torshovdalen from the hydrological year of 2017/2018 showed abnormal values with high water deficit compared to the mean normal (1961-1990) and previous hydrological year of 2016/2017. The soil water balance of 2017/18 was also compared to expected climate change of an annual increase in precipitation of 10% and temperature of
0.5°C. Estimated runoff suggested a higher surface runoff from 19 l/s*km² (2017/18) to 21 l/s*km² (Climate change) for Torshovbekken catchment implying the future consequences of climate change.

This study offers fundamental research for future stormwater management projects and will be a supplement to water resource officials in proceedings regarding sustainable urban drainage planning in Oslo. Methods used herein exemplify a simplistic approach to stormwater management and uncertainties may apply. Stormwater management is becoming more and more important as weather patterns are changing, offering greater seasonal contrasts.
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1 Introduction

Recharge patterns have been altered since the mid-20th century, and according to the Intergovernmental Panel on Climate Change (IPCC) there is a correlation between human influence and changes in the global water cycle. Extreme weather and global ice sheet melting are only some on the consequences in warming of the atmosphere and oceans (IPCC, 2013).

Natural land cover and land use are diminishing due to densification of cities, which implicate the urban hydrogeological cycle (Cui and Shi, 2012). Impermeable surfaces and structures are preventing water from penetrating the ground, forcing water to find another course. This may not cause immediate damage; however, long term stormwater problem will collectively affect the drainage system and distress the infrastructure. Therefore, Stormwater management and green infrastructure has become an important line of work in growing cities (NOU, 2015).

A report on Climate adaptive stormwater management from Norsk Vann is highlighting the importance of preventing water overflow by implementing stormwater solutions such as: reopening rivers, retention pools, rain gardens etc. (Lindholm et al., 2008), collectively called sustainable drainage systems (SUDS) (Fletcher et al., 2015).

Stormwater solutions are implemented to reduce runoff and control infiltration. Urban green surfaces are often compacted due to foot traffic and blockage of pores, which may impact infiltration rate. Changes in metropolitan parcels with movement of soils, compaction and filling and shifting of soils results in a mixed soil profile that is characteristic of urban soil (Environmental Protection Agency, 2011) the great variation on content will have a great impact on the behavior of water. Therefore, infiltration rates, subsurface investigation and the water balance will be necessary to evaluate (Ahmed et al., 2014, Solheim, 2017, Lindholm et al., 2008).
1.1 Objectives

- Characterize and assess the geological subsurface material, infiltration potential and determine the hydraulic properties of Torshovdalen sub-catchment.

- Demonstrate the applicability of weather data to estimate hydrological parameters and assess how future climate change may influence the soil water balance.

- Evaluate relative effects of changes in temperature and precipitation and how future climate conditions are affecting the groundwater recharge and flow.

- To identify stormwater management actions for Torshovdalen sub-catchment.
2 Theoretical framework

This chapter describes the major concepts about the occurrence, movement, quality and origin of groundwater. These concepts are discussed as they apply to the knowledge for parameters included in this master project.

2.1 Hydrogeology and hydrology

Hydrology is the geoscience in which addresses the occurrence, distribution, movement and chemistry of the earth’s fresh waters. Moreover, hydrogeology is the geoscience in which encompasses the interaction of geological material and water within earth’s crust. Even though they incorporate different aspects of the hydrological cycle, the two geosciences are largely interconnected (Fetter, 2001, Dingman, 2015).

2.2 Hydrological cycle

The hydrologic cycle is the system of which water is recycled between the earth’s surface and the atmosphere (Ødegaard et al., 2014, NOAA, Updated 2019, Fetter, 2001, Soliman et al., 1998, Dingman, 2015). The hydrological cycle involves several major processes such as: Evaporation, transpiration, precipitation, runoff, condensation, infiltration and percolation. (Figure 2-1).

These processes make up the major storage and flow components in the hydrological cycle. It is the continuous circulations of water from the oceans to the atmosphere through evaporation and subsequently through precipitation in which brings the water back to the earth (Soliman et al., 1998). When water falls as precipitation it may either be re-evaporated, stored as snow and ice or become groundwater. Through infiltration, water may enter the ground where it is drawn through roots, replenishing growing plants that will ultimately transpire water back into the atmosphere. Alternatively, water will continue to flow laterally as groundwater in the vadose zone, or further percolate downwards (Fetter, 2001).
Solar energy is the driving force for the hydrologic cycle. When water molecules change its state between vapor, liquid and solid, there is an accompanied heat change occurring. Latent heat of vaporization, condensation, fusion and sublimation are a few of the processes that requires input of energy. This energy may vary depending on the latitudinal location. Where equator may experience a more constant heat balance, the pole regions are highly variable (Fetter, 2001). The hydrologic cycle may be grouped into three disciplines encompassing the aforementioned processes: Meteorology, Surface hydrology and hydrogeology (Soliman et al., 1998).

![The hydrologic cycle](image)

*Figure 2-1: The hydrologic cycle Department of water resource (2003)*

### 2.2.1 Water balance

To evaluate the hydrological cycle, a simple bookkeeping of the parameters can be quantified by the means of a water budget. The water budget demonstrates the hydrogeological cycle based on the applicable parameters (Healy et al., 2007). Determining the availability of water for people and the environment, makes calculating the water budget relatively uncertain, although the concept of a water budget seem simple.

A water budget is the rate of which water that is stored in an area, such as a watershed, is balanced with water that enters and leaves the said system. Precipitation is a leading factor in the water balance (XU and Singh, 1998). Precipitation can fall as both snow and rain, making
the input and output for a water balance highly variable. Snow will accumulate when temperatures are < 0°C, thus capturing and holding water until it either vaporizes into the atmosphere or melt into the ground. Precipitation and potential evapotranspiration are important factors in Thornthwaite water balance calculation as wetness and dryness factors cannot be determine without the other (Thornthwaite and Mather, 1957).

Calculations of water budgets include parameter from catchment inputs and outputs such as Eq 2-1. This approach introduces hydrological settings and give understanding to parameters such as groundwater recharge and flow (Fish, 2011). This section focuses on the Groundwater system in an urban setting. Variables, parameter and boundaries are important factors to determine a catchments yearly water balance. The water balance is based on the generalized conservation equation considering conservation of mass, Newtons first law, and the first law of thermodynamics (Dingman, 2015):

\[ Amount\ in - Amount\ out = Change\ in\ storage \]

\[ Eq\ 2-1 \]

Where Amount in is the sources of water input and Amount out is all factors of water leaving the system. At an annual perspective, change in storage is considered negligible (Healy et al., 2007). Consequently, this equation can be specified into its different parameters. The Following equation is the water budget equation and its relating components according to (Dingman, 2015):

\[ P + gw_{in} - (Q + RET + gw_{out}) = \Delta S \]

\[ Eq\ 2-2 \]

Where P is precipitation, gw_{in} is water flow into the watershed, Q is runoff, ET is evapotranspiration (the sum of evaporation from soils, surface-water bodies, and plants), \( \Delta S \) is change in water storage, and gw_{out} is water flow out of the watershed.

Approximating correct inflow and outflow parameters may be difficult. Water balance boundaries are often delineated to a set watershed boundary, and thus making values assumed to be zero (Fish, 2011). Considering these changes, the equation becomes simplified as follows:

\[ Q = P - ET - \Delta S \]

\[ Eq\ 2-3 \]

Where P is precipitation in (L), ET is the evapotranspiration in (L), \( \Delta S \) is the change in storage in (L) and Q is the surface water runoff. Using this equation with monthly and annual climatological parameters, a water balance can be estimated.
2.3 Meteorological factors

Meteorology includes the energy balance, circulation of water through the atmosphere, cooling of air, condensation of water molecules to form water droplets. Fundamental parameters in climate estimations are temperature, pressure, humidity, wind, solar radiation, and precipitation (Soliman et al., 1998, Dingman, 2015).

2.3.1 Precipitation

The occurrence and distribution of precipitation is heavily influenced by demographic and topographic features (Fetter, 2001). Precipitation that falls on the earth will go through various segments of the water cycle, and precipitation may occur as: Rain, snow, hail, drizzle and dew (Shukla, 2014, Fetter, 2001, Hiscock and Bense, 2014). When the available water droplets have become of critical size, it falls back to the earth due to gravity and may be intercepted by vegetation canopies, infiltrate the ground or collect to form overland flow (DeWalle and Rango, 2008).

For precipitation of any kind to occur, it is dependent on the available water vapor in the atmosphere. Water vapor are H₂O molecules interacting with molecules of other gases. In conjunction with water vapor, vapor pressure represents the maximum of water vapor in the atmosphere, at temperature T (Dingman, 2015).

2.3.2 Evaporation

Evaporation is a function of solar radiation, humidity, and wind. The removal and addition of energy is driven by the sun and the energy balance. Evaporation/condensation act both as a latent heat flux and heat transfer between a surface and the atmosphere. (Dingman, 2015, Egger, 2003). Evapotranspiration encompasses all processes which water is transformed from a liquid or solid to atmospheric water vapor and includes; transpiration form within plants, sublimation from ice and snow, and the interception loss from surfaces (Dingman, 2015). Phase changes for water is explained in (Figure 2-3). The exchange between the atmosphere and the land surface of water and energy is a governing process of the hydrologic cycle (Dingman, 2015).

Energy

Global climate and the hydrologic cycle is reliant of energy in form of electromagnetic radiation from the sun (Dingman, 2015). Latent heat is the energy that is absorbed or released when there is a phase change of a given mass (Dingman, 2015). Latent heat does not rely on temperature change when losing or gaining heat. Water’s latent heat is relatively large
compared to other substances because of the energy required during vaporization and melting by breaking of the hydrogen bonds, and the creation of the hydrogen bonds by freezing and condensation. Latent heat of fusion is the amount of heat energy required for melting and freezing. Freezing and thawing is an important seasonal process. For temperatures reaching zero, heat must be conducted to or from the substance in order for melting of freezing to occur. Latent heat of vaporization is the amount of heat energy required for vaporization and condensation. For vaporization to occur, complete breakdown of hydrogen bonds, and the latent heat decreases with temperature (Dingman, 2015, DeWalle and Rango, 2008).

The exchange between liquid and air until an equilibrium is reached and the vapor pressure is the vapor exchange. Usually, the molecules of water are attracted to other molecules within the body of water by hydrogen bonds. When Temperature increases molecules that have enough energy available will break the hydrogen bonds and escape to the near surface air (Figure 2-2). Number of molecules that separate this bond will increase with increasing temperature. Here, some molecules may stay in the air and some will return to liquid. Evaporation is governed by the latent heat of vaporization (Dingman, 2015)

Sublimation of snow is the process of water going from solid snow crystals to vapor the phase into the atmosphere. Sublimation occurs when vapor pressure near the snowpack is close to that of the surrounding air pressure (Svoma, 2016). The energy balance is defined as the sum of all energy fluxes at the surface (Hock, 2010).

Vapor pressure gradient determines whether or not the snow will melt due to latent heat vaporization or vaporize due to the latent heat of fusion and vaporization (Dingman, 2015). Surface vapor pressure is suggested by literature to be constant at 611 hPa when temperatures
at surface is 0°C (Strasser et al., 2008, DeWalle and Rango, 2008, Hock, 2010). Either processes will occur under following statement:

\[ \text{If Air vapor pressure} > \text{Surface Vapor Pressure: Melting will occur.} \quad \text{Eq 2-4} \]

\[ \text{If Air vapor Pressure} < \text{Surface vapor pressure: Evaporation will occur.} \quad \text{Eq 2-5} \]

For snow to accumulate is must hold an average surface temperature of ≤ 0°C (Dingman, 2015). Accumulating snow is possible when net energy is negative (away from the surface), however, as soon as available energy and net inputs are positive (towards the surface) snow begins to melt (Strasser et al., 2008). The snowpack energy balance is the net flux of energy as a function of available energy multiplied by the change in the internal energy in the snowpack over time (Dingman, 2015). Snowmelt is governing the annual runoff, and thus losses from sublimation will greatly affect the hydrology of high altitude and latitude regions.

![Phase Diagram for water](image.png)

*Figure 2-3: Phase Diagram for water. Diagram show the relationship between Liquid, gas and solid with respect to pressure (atm) and temperature (°C). Pearson (n.d).*
2.4 **Surface hydrology: Groundwater- and surface water interaction**

Surface hydrology encompasses the water that is at the surface of the earth: water that is stored in streams, lakes and rivers (Fetter, 2001). Hydrology is closely related to meteorology, soil science and fluid mechanics. Surface water is the circulation, distribution, including the chemical and physical properties, and the reaction of water to the environment (Maidment, 1992).

As precipitation reaches the land surface, infiltration occurs as water penetrates the soil by the influence of gravity and capillary forces (Shukla, 2014). Infiltration capacities are dependent on the type of soil and the rate of precipitation. Available pore space in between grains of sediment, will greatly affect at which rate water is moving downward. Infiltration is important part of the hydrologic cycle as it is not only feeding the groundwater and nearby bodies of water, but also the local vegetation (Shukla, 2014).

Rainwater that does not infiltrate the ground will become surface runoff. Surface runoff is a contributor to streams, intermittent or perennial flow. This occurs when the rate of precipitation exceeds the rate at which the soil will absorb and infiltrate water (Healy et al., 2007). Surface runoff can be divided into three concepts: surface runoff, interflow and groundwater discharge. Where surface runoff is the overland flow until it reaches a stream and becomes streamflow. The interflow is the runoff that laterally flows through the top soil and may yield ephemeral springs or become perched aquifers (Soliman et al., 1998). Water that becomes runoff after a rain event or the melting of snow is considered direct runoff and is the sum of immediate runoff and subsurface runoff. In mountainous regions, direct runoff is only considered to be overland flow as little water is able to infiltrate (Soliman et al., 1998).

Groundwater and surface water interaction are key elements in the hydrological cycle: Water and chemicals are constantly exchanged between surface water and groundwater. For example; Water that flows in streams may either feed the groundwater (effluent) or be fed by the groundwater (influent) by transporting water through the bed of the water body the zone is also called the hyporheic zone. Thus, water will either recharge or discharge depending on the current interactive circumstances. Changes in this natural cycle will affect both processes, and thus active water management is important for both groundwater and surface water (Department of water resource, 2003).
Water enter the streams through various events (e.g., rain or snowmelt) are called event flow. Contrary, base flow is water enters the streams at a slow rate with long residence time and is responsible for maintaining the streamflow between precipitation events (Dingman, 2015). The hyporheic zone extends both horizontally and vertically, the hyporheic zone have a great influence of water quality, aquatic organisms, and the biodiversity. In concordance with the hyporheic zone, the hyporheic flow is the flow and water exchange between the stream bed and groundwater. The flow is produced by the topography of the streambed, and the difference in pressure-head (Dingman, 2015, Department of water resource, 2003).

2.4.1 Infiltration

Infiltration is the process in which water is entering the soil. Water will enter the soil through unfilled pore space in the porous subsurface materials. The unsaturated zone is the zone at which the pore space or the soil is partially filled by air and water (Dingman, 2015). Infiltrating water will move vertically by the force of gravity and is either stored or will continue to flow in several ways: 1) Water will percolate vertically at a slow rate through pore space and open path, 2) water will remain in the soil and be taken up by vegetation, and 3) water may become groundwater recharge (Dingman, 2015, Hiscock and Bense, 2014). Groundwater flow may occur due to gravity and the pressure gradient between the subsurface and the atmosphere.
Soil tend to infiltrate at a higher rate at the when precipitation is introduced. Continuous water input will result in an increase of water available for infiltrating thus causing the pressure gradient to decrease (Dingman, 2015). Factor affecting infiltration are water input, surface slope, and porosity. Other factors include chemical properties, human influence, compaction, frost, and soil textural properties (Dingman, 2015).

Subsequently the infiltration rate is the rate at which the water is infiltrating the soil from the land surface (Hiscock and Bense, 2014). Infiltrations rates will decrease rapidly at the start of a rainfall event, as the water is entering the unsaturated zone, and the new water is now filling the available pore space (Figure 2-4). When the soil if fully saturated, the infiltration rate will reach a constant value at which the water has reached its maximum infiltration rate and thus reached its infiltration capacity (Hiscock and Bense, 2014, Dingman, 2015).

Saturated hydraulic conductivity ($K_{sat}$) is often considered the equivalent of a constant infiltration rate at field capacity. The field capacity is the volumetric moisture content of the soil after the soil has been allowed to drain and an equilibrium is reached. If precipitation continues and the water input to the soil exceeds its capacity, surface ponding and surface runoff will occur (Hiscock and Bense, 2014).

2.5 Physical hydrogeology and subsurface water

Hydrogeology is the science of subsurface water; however, surface- and groundwater are a closely linked system. Hydrogeology encompasses the water that exists, move and interact with the geology within the earth crust (Hiscock and Bense, 2014). Subsurface water is held between interconnected pore space and voids in the soil matrix. Water may also be stored in interconnected water bearing fractures in bedrock (Guymon, 1994). The subsurface is divided into two zones; The unsaturated and saturated zone, and is divided in aquifers of confined, unconfined and leaky.

Water moves relatively slowly through the unsaturated zone, and may remain in this zone until plant will transpire water, or it will discharge into a nearby watercourse (days) or move downward an recharge the lower aquifers, this may take years (Healy et al., 2007).

The unsaturated zone can be further divided into three zones: root zone, intermediate zone and capillary fringe zone. Unsaturated zone if found just below the surface, where the pores between grains are filled with both water and air. The root zone is the depth of which roots and plants are found and supports plant growth. The root zone is commonly understood to be one
meter thick (Guymon, 1994). Voids in the soil is also accompanied by decayed roots of previous plants, animal and worm burrows. Porosity and permeability is usually much higher than that of the lower soil profiles. (Heath, 1983).

Capillary fringe is the boundary or subzone between the unsaturated and the saturated zone before the water table. This boundary is governed by capillary forces which draws water upwards in between the open pore space commonly called capillaries. Attraction between water and rocks results water behaving as a vail around the rock or sediment particle and thus rises through the pores against the pull of gravity. Water in this zone is experiencing negative hydraulic pressure, meaning, the pressure exerted on the water is less than that of atmospheric pressure (Heath, 1983). Sizes of available pores determines the zones thickness, while gravel and sand may have a relatively short capillary fringe zone, clay and silt, because of the lack of interconnected pores, may have a fringe zone stretching several meters.

Impervious soil may create pockets of saturated soil which in turn saturate the overlying pervious soil. Local pockets of saturation are commonly called perched aquifers. (Guymon, 1994). Alternating layers of fine grained sediments and coarser grained sediments creating aquifers and aquitards (Heath, 1983). Perched aquifers are often found in glacially affected areas where lenses of clay are formed due to glacial outwash. They may also occur were there has been volcanic activity. (Fetter, 2001).

The saturated zone is where the soil is fully saturated by groundwater that makes up an aquifer. An aquifer is defined as a geologic unit which water is transmitted or stored to a satisfactory degree for supplying wells and springs (Fetter, 2001). This zone is recharged by percolating water from the unsaturated zone, it may also be recharged by lateral flow or the upwards motion of water (Fetter, 2001).
2.5.1 Aquifer

There are two main types of aquifers: Alluvial and fractured rock. Alluvial aquifers consist of sediment deposits of sand and gravel and fine grains materials such as clay and silt. Coarse material such as gravel and sand provide good conductance of water through the available pores. Clay and silt however, restrict water from flowing due to the lack of available pore space and is thus commonly called aquitards. Alluvial Aquifers are porous unconsolidated sediment aquifer that transfers water through the pores in between sediment grains (Department of water resource, 2003). Bedrock, crystalline rock, basement and hard rock are also denoted fractured rock. Fracture rock aquifer transports water through joints and fissure in a normally non water bearing surrounding, thus has limited storage capacity (Kirk, 2016). Rocks that transports water through pores and fractures are commonly denoted semi consolidated such as types of limestones and sandstones (Heath, 1983). Alluvial or fractures rock aquifers depend on different physical properties in order to transport water through its media. Two of the most important factors governing the groundwater flow is porosity and hydraulic conductivity. Water in the alluvial aquifer yields >>> water than a fractured rock aquifer.
A confining aquifer has restricted water movement and low hydraulic conductivity (Heath, 1983). It is underlaying a confining layer which hinders downward movement of water to some degree. An Aquitard is an example of a confining layer of extremely low permeability which is able to store groundwater or slowly transmit water through the layer, this is also called a leaky confining layer (Figure 2-5) (Fetter, 2001). Fluid pressure is greater than atmospheric pressure (Guymon, 1994). Recharge may occur where parts of the geologic unit is outcropping at the surface or by percolation through a leaky confining layer. Water in a confined aquifer is under pressure, and water will rise above the top of the aquifer if the confining layer is penetrated by a well, this water level is called potentiometric surface. In the case the potentiometric surface is above the surface, water under pressure will create an artesian effect.

An aquifer where water freely fills and drains the pores in accordance with a fluctuation of the water table is an unconfined aquifer (Heath, 1983). This aquifer is close to the surface and contain several layers of materials of various but high permeability. This aquifer is also called the water-table aquifer or phreatic surface (Fetter, 2001). Fluid pressure equals the atmospheric pressure. Unconfined aquifers are readily contaminated if located beneath industrial or urban areas as chemicals are easily transmitted through the soil from local spillage (Guymon, 1994).

2.5.2 Porosity

Porosity is the ratio of voids of a soil or rock of a given volume, or the fraction of a soil or rock volume that is occupied by voids. It is measured in the amount of water said material is able to store and it is often expressed as a percentage (Heath, 1983, Hiscock and Bense, 2014). Porosity may be found by laboratory procedures, from knowing the bulk density ($\rho_b$) and particle density ($\rho_s$), thus porosity can be calculated as follows:

$$n = 1 - \frac{\rho_b}{\rho_s}$$

Eq 2-6

Unconsolidated sediments are composed of both angular and rounded particles have a larger porosity then that of consolidated or indurated sediments. Crystalline rock has a very low porosity as micropores are within the crystal lattice. Shape, sorting, compaction, fracturing are some controlling factors of porosity (Hiscock and Bense, 2014). Porosity is dived into primary and secondary porosity. Primary porosity is the natural character of a soil or rock that developed during its formation. Subsequently, secondary porosity is the porosity established after physical
or chemical weathering along bedding planes or joints. In fractured rock aquifers have a typically low permeability and the porosity is associated with the water contained within the fracture. Fractures are decreasing proportionally to the porosity as depth increases (Hiscock and Bense, 2014).

2.5.3 Darcy’s law

Movement of water through interconnected voids, the groundwater flow is governed by Darcy’s law. French engineer Henry Darcy studied the movement of water through a porous media (Hiscock and Bense, 2014). Darcy tested the behavior of water using a tilted column and suggested that flow (Q) is proportionate to the change in height of the water (h₁-h₂) and the cross-sectional area of the outlet (A), however, inversely proportionate to the length (L).

Darcy’s law is used under the conditions that soil textural and hydraulic properties are averaged and thus bulk flow is described (Dingman, 2015, Root, 2016). Darcy’s law can be written as follows:

\[ Q = -KA \frac{dh}{dl} \]

2.5.4 Hydraulic conductivity

Hydraulic conductivity is the ability to transmit water of sediments rock, it is often interchangeable with permeability. Properties of a geologic material such as size, shape pore space, migrating fluid and specific weight will all greatly influence the hydraulic conductivity.

\[ K = k_i \frac{\gamma}{\mu} \]

Where \( \gamma \) is the Specific weight \((\gamma = \rho g)\), \( \mu \) is the viscosity and \( k_i \) is the intrinsic permeability (Hiscock and Bense, 2014). As a parameter used in Darcy’s equation, Modifying Darcy’s law we can rewrite hydraulic conductivity as specific discharge that has the same dimensions (L/t).
\[ q = \frac{Q}{A} = -k \frac{dh}{dl} \]  \hspace{1cm} \text{Eq 2-9}

Where \( q \) is the specific discharge (L/t), and \( Q \) is the rate of volume water (L\(^3\)/t), \( A \) is the area (L\(^2\)), \( k \) is the saturated hydraulic conductivity (L/t) and \( dh/dl \) is the hydraulic gradient. Groundwater flow within a rock formation may be expressed as the specific discharge defined as the flow per unit cross sectional area perpendicular to the flow (Gustafsson and Morosini, 2002, Dingman, 2015).

### 2.5.5 Aquifer characteristics

Transmissivity is the rate of water transmitted horizontally through an aquifer area at a given thickness (\( b \)) with a specific hydraulic gradient (Fetter, 2001, Hiscock and Bense, 2014). Transmissivity is closely related to the potential yield of an aquifer which is dependent on the saturated thickness and is thus defined as the product of hydraulic conductivity and saturated thickness of the aquifer (Department of water resource, 2003). For a confined aquifer following equation apply:

\[ T = kb \]  \hspace{1cm} \text{Eq 2-10}

Where \( T \) is transmissivity (L\(^2\)/t), \( k \) is the hydraulic conductivity (L/t), and \( b \) is the saturated thickness of the aquifer (L). Storage coefficient or Storativity also depends on the aquifers confining or unconfining characteristic. Storativity of a confined aquifer is the volume of water that is absorbed by a permeable unit or expelled from storage area per change in head. Storativity is commonly called the storage coefficient and is the product of specific storage and saturated thickness. Storativity for a confined aquifer is calculated as follows:

\[ S = S_s b \]  \hspace{1cm} \text{Eq 2-11}

Where \( S \) is storativity (unitless), \( S_s \) is the specific storage (1/L), and \( b \) is the thickness (L) (Hiscock and Bense, 2014).

Specific storage \( S_s \) (1/L) is the amount of water of water stored per unit volume of a saturated formation of confined aquifer and the release of water from storage under a hydraulic gradient. Due to increasing stress and the expansion of water due to
decreasing pressure, compressibility factors must be defined: $\alpha = 10^{-8} - 10^{-10}$ m$^2$/N, and $\beta = 4.4 \times 10^{-10}$ m$^2$/N (Kruseman and De Ridder, 1994). Porosity in fractured shales are between 0.05-0.5 (Freeze and Cherry 1979 as mentioned by (Hiscock and Bense, 2014)).

\[ S_s = \rho g (\alpha + n\beta) \quad \text{Eq 2-12} \]

Where $\rho$ is the density of water (M/L$^3$), $g$ is the gravitational acceleration (L/T$^2$), $\alpha$ and $\beta$ are the compressibility of the aquifer skeleton and water respectively (1/M/LT$^2$) and $n$ is porosity (L$^3$/L$^3$).

In an unconfined aquifer, water will fluctuate depending on the amount of saturation in the aquifer. Water will either drain from the pores as water level drops or stored as water level increase. The process of storing and releasing water is due to specific yield ($S_y$) or depending on the specific storage of the aquifer. Specific yield is sometimes equivalent to porosity in soil and rock, with values ranges from (0.01-0.30).

\[ S = S_y \quad \text{Eq 2-13} \]

Storativity is a fraction of total volume that will drain from a saturated media. Values for a confined aquifer ranges from 0.005-0.00005, and for an unconfined aquifer storativity ranges from 0.02-0.30 (Fetter, 2001, Hiscock and Bense, 2014). When water is held against gravitational forces it is denoted Specific retention ($S_r$). The sum of Specific retention and specific yield ($S_y$) is porosity $n$. Grain size is decreasing with increasing specific retention, indicating that finer grained sediments may have higher porosity than specific retention (Fetter, 2001).

\[ n = S_y + S_r \quad \text{Eq 2-14} \]

2.5.6 Water table

There are many factors that are due to groundwater table raising and falling. Leakage from septic systems, precipitation, however it is only if significant volumes of water that there might rise the regional water table (Howard and Gelo, 2002). Groundwater fluctuation are generally due to a change in volume of water stored in the aquifer: the addition or abstraction
of water. Unconfined aquifers are mostly affected as they are in direct contact with the percolated water. Groundwater recharge increases the water table one inch of precipitation moving into the ground may rise the groundwater table with to more than an inch. Because of the available void spaces in the aquifer porosity of 5% will theoretically raise with 20 inches.

The water table in an unconfined aquifer is at constant atmospheric pressure, and thus the hydraulic head remains the same (Price, 1996). On the contrary, a confined aquifer experiences pressure of the overlying bed and the atmospheric pressure, which is reinforced by the water and the aquifer framework. Thus is the atmospheric pressure would increase, the pressure is conducted to the confining layer resulting in a higher pressure in the confined aquifer (Price, 1996). If sufficient water is added to storage to the point of maximum capacity, water will leave the aquifer in form of a spring of seepage towards streams and rivers. An aquifer is composed of different soil and sediment textures that occupy a certain volume of the aquifer. The texture of the soil governs the amount of water that are available in the storage of an aquifer. Theoretically, if 1 mm of water is reaching the ground, and assuming no runoff nor infiltrating occur, the water will produce a layer of water that is 1 mm deep. However, if said water infiltrates the ground having a specific yield of 0.1 (10%), the water table may rise by 10 mm due to: 1) amount of groundwater recharge, 2) Changes in atmospheric pressure, and 3) changes caused by aquifer deformation (Price, 1996). Following equation explains this theory:

\[
\text{Rise in water table} = \frac{\text{Infiltration}}{\text{Specific yield}}
\]

Where rise in water table (L), Infiltration (L), and specific yield (fraction). An Aquifer having a very low specific yield may experience more rise in the water table, than that of an aquifer with high specific yield (Price, 1996).

2.5.7 Soil and rock properties

Soil texture and soil surface play a major role in hydrogeology. Especially in the response to precipitation and snowmelt. Soil surface determines the infiltration and runoff coefficients as well as soil moisture influences the energy balance and moisture exchange with the atmosphere. (Dingman, 2015). There are three general properties of the soil that influences water movement: Propensity for infiltration of water, ability to transmit or retain infiltrated water, depth to the water table or to an impermeable surface (Dingman, 2015).
The subsurface zone near the land surface in which there is interaction between the atmosphere, the biosphere and pedosphere, and the near surface portion of the hydrosphere and the lithosphere is called the critical zone. The critical zone encompasses several soil horizons and entails the entire region from the vegetative ground to the bottom of the aquifer.

![Pedologic soil horizons](Dingman, 2015)

Soil is often defined as a medium for holding and filtering water and other chemicals as well as contribute to plant growth (Shukla, 2014). Soil is a result of many natural processes such as weathering and erosion and contains both inorganic and organic material. Soils may be divided into different horizons of different thickness showing distinctive layer properties in a soil profile. Top soils are highly organic whereas the following horizons will resemble the parent material decreasing with depth (Figure 2-6). Generally, Texture structure, color and density are four soil properties that are important to evaluate. Urban soils may be difficult to determine as it has been drastically changed from its original form (Craul, 1999).

Soil hydraulic properties will govern the water flux and the infiltration capacity, and will be limiting to water infiltrated from other sources and produce water flowing through the soils from higher elevation (Haghnazari et al., 2015). Effects of temperature on infiltration have
been discussed in several studies (Constranz and Murphy, 1990, Haghnazari et al., 2015, Braga et al., 2007). Warmer temperature may impact the infiltration as the viscosity of water decreases and faster infiltration is expected. Similarly, colder water leads to slower infiltration. Understanding the properties of the soil can help determine the amount of water that a soil can store. Porosity, texture, grainsize and other characteristics are factors affecting how much water can be retained or drained (Brooks et al., 2012).

Changes In pore space within the soil will affect the infiltration rate (Helalia, 1993). Subsurface drainage is determined by the layer of least permeability. Pore space in soil will affect the flow of water and is related to the particle diameter. Sandy coarse soils have larges pores, fines grain size decrease in pore space. Climate conditions govern the processes influencing the soil freezing such as surface water heat and soil water phase changes. Reduction of permeability and thus changes in the hydraulic properties of the ground is a direct result of freezing (Oztas and Fayetorayb, 2003).

The frozen water will expand and continue to pull water closer until it has developed into a greater mass of ice. Depth of the frozen soil is dependent on the snowpack and temperatures (Lilleøren, 2018). Freezing and thawing is a physical process affecting the aggregate stability. It is suggested by (Schouse et al., 1990) that aggregate stability may be corelated to grain size distribution, and observations from the study showed that silt and clay content. Because of this, will the risk of freezing and thawing of the lower clay layers not only expand and collapse the surrounding soil, but also be a risk for the settlement of soils. (özgan et al., 2015). Frozen ground (tele) is a important parameter in in urban development. Frozen ground may cause implication underground and may alter the infiltration capacity tremendously.

Humus is organic matter which have undergone decomposition and is naturally being blended into the soil (Craul, 1999). Organic matter has a great impact on the water holding capacities in the soil, the increase in water holding capacity is proportional to the humic content. However too much organic matter can lead to an increase in the water holding capacity and soil may remain wet for longer periods. This will create anoxic condition unfavorable for soil organism and decompositions is stalled (Craul, 1999). Organic material in nordic soil is dominated by old but stabile humic substance. According to (Breland, 1992) The Quality of Organic matter is more important than the quantity. Stabil humus is important for naturally occurring chemical processes. Humus is a unique source of nitrogen, phosphorous and sulfur
for productive plant growth, aeration of the soil, drainage and water holding capacity. Soil rich in humus is thus more productive and beneficial for nature conservational purposes (Stevenson, 1972). Also, as oxygen is produced in the ground it increases the respiration by plants, and ultimately affect the total evapotranspiration in an area (Kirkham, 2005). High organic matter content will absorb and hold water, and then distribute available water to plants during episodes of no precipitation (Bhadha et al., 2017).

2.5.8 Urbanization and stormwater management

According to United Nations (2015), more than 50% of the worlds inhabitants are living in an urban areas, and 66% is predicted to be living in cities by 2050. Urban transition is a change of settlement location from rural areas to urban cities and is often concentrated in urban areas that emphasizes on industrial and service activities (United Nations, 2015). Urbanization is destroying the natural environment, by decreasing natural land and open spaces, and increasing impermeable surfaces such as pavement, rooftops parking lots among others. Contaminated stormwater runoff and catchment areas is a direct result of urbanization (Chithra et al., 2015). Climate change is influencing the volume of runoff and consequently, risking pollution entering clean bodies of water and thus making changes to both quality and the quantity of storm water (Jotte et al., 2017, Chithra et al., 2015). The effect of imperviousness is apparent on urban areas as 15% of the annual rainfall is expected to infiltrate compared to natural areas where 50% is expected to infiltrate Figure 2-7.

Urban stormwater management is the application of remediate stormwater runoff on urban developed land (Barbosa et al., 2012). It is the practice of implementing solutions within the urban environment to help decrease the risk of flooding and pollution. This natural approach is by redirecting runoff into a natural infiltrating and water retaining environment such as rivers, rain beds, swales, detention pools, and away from the sewer network (Fletcher et al., 2015). These methods are often called sustainable urban drainage systems (SUDS) and is implemented to control stormwater quality quantity as well as integrating a recreational value to the approaches (Jotte et al., 2017). Traditionally, Stormwater management only measure was to divert runoff into local bodies of water or watercourses to avoid flooding. However, quality and Quantity of runoff have been in focus as is has become an increasing problem. Source control has become an integral part of the modern stormwater management (Staalstrøm and Røed, 2016).
Urban hydrogeology

Urban hydrogeology entails the hydrogeological cycle and how the cycle changes in urbanized areas (Ødegaard et al., 2014). The natural hydrogeological cycle is broken or changed when introduced to impermeable surfaces. Infiltration is inhibited and thus creating a risk of abnormal quantities of surface runoff (Figure 2-7). Implications such as flooding of basements and sewer pipes may occur, and it can be both costly and detrimental (Solheim, 2017). According to (Jotte et al., 2017) 60-70% of the annual precipitation is estimated runoff volume in developed areas. The hydraulic properties of an urban areas may be difficult to predict, and local climatological data and subsurface knowledge play an important role in water resource management (Niemczynowicz, 1999). According to (NOU, 2015), In Oslo, costly damages due to surface water and poor storm water management can range from 1.6 to 3.6 billion every year.
However, with increased population, developed areas, and changing climate, the problem will amplify the cost up to 45 to 100 billion kroner in 40 years (NOU, 2015).

Urban catchments have poor water quality because of contamination from stormwater from roads and roofs, and contains a great deal of environmental pollutants (Krystad, 2017). Road salting is one of the greater contributors to Sodium chloride contamination in urban soils and may severely damage vegetation and affect downstream watercourses (Li et al., 2014, Hofman et al., 2012). High concentration of fecal bacteria correspond to urbanized areas compared to rural areas (Krystad, 2017).

2.5.10 Climate change

Increase in global mean temperature is a concern among researchers and it is estimated that already ~1°C has increased compared to pre industrial temperature, and is predicted to increase to 1.5°C if current global warming rate continues (Masson-Delmottem et al., 2018). Also, global precipitation is likely to increase with increasing temperature, and more pronounced at higher latitude and longitude regions. It is suggested by (National Research Council, 2011) that for every 1-4°C, there will be 5-10 % more precipitation in northern latitudes.

2.5.11 Strategy for storm water management in Oslo

A Stormwater management strategy has been implemented in the city of Oslo. By utilizing open and local storm water solutions, Oslo municipality is working toward becoming more climate adaptive and introduce sustainable structures. Stormwater of acceptable quality, infiltration and retention of water will diminish implications due to surface water and urban flooding (Oslo Kommune, 2016). The water directive framework of EU, imposes strict regulations on the quality of water, and runoff is a contributor and conveyer of poor quality water that ultimately pollutes other water sources (Oslo Kommune, 2013). With this in mind, a three-step strategy has been suggested for implementation in Oslo. Following the strategy, Oslo municipality hopes that both newly developed areas and older development is implementing sustainable and climate adaptive measures (Oslo Kommune, 2016). Figure 2-8 illustrates the three-step strategy and the different focus areas. The three-step strategy is based on the amount of precipitation that may occur at any time. Firstly, during normal precipitation events, the main goal is to catch and infiltrate the rainwater thus reducing the risk of flood and maintain a normal water balance. Secondly, if there is an increase in precipitation, the next step will be to delay
and retain the water by implementing areas that will capture and hold excess water. Lastly, to be able to handle large amounts of precipitation, it is thus important to redirect the stormwater safely until it discharges in the fjord. Here, reopening rivers and smarter waterways for extreme weather conditions are proposed as action plans.

Reopening rivers have become an important stormwater management practice in Oslo. According to (Oslo Kommune, 2015), a sustainable way to handle climate change and increased precipitation is to reopen the many buried streams in Oslo. Hovinbekken and Teglverksdammen are examples of successful projects, located at Ensjø in Oslo. Hovinbekken, although relatively small, is one of the greater watercourses in Oslo that historically was redirected into the sewer system (Krystad, 2017). Teglverksdammel in conjunction with the ~650 m opening of Hovinbekken, works as a natural cleaning system containing vegetative and sedimentation basins (Krystad, 2017, Kristiansen, 2019). Teglversdammen is built to remediate fecal contamination, pathogenic microorganisms, and various other contaminations.

Figure 2–8: Three-step strategy and storm water managements solutions. 1) Catch and Infiltrate 2) Delay and Retain and 3) Safe Floodways. Modified from (Oslo Kommune, 2016).
### 3 Study Area

#### 3.1 Geographic location and landscape

Torshovdalen is a green recreational area located in Oslo municipality, in the district of Sagene. The park is ~0.15 km², elongated, stretching ~ 850 m and gently sloping in SW-direction. Torshovdalen is located between the dense city development of the city center of Oslo to the south and the natural forested area of Grefsenkollen to the north. It also marks the boundary between the inner and outer municipalities together from Bjerke in the north to Grunerløkka in the south. According to (Storrvik et al., 2012), along the western side of the park, the terrain is quite steep, thus larger trees along the hillsides marks the boundary between the park and residential and commercial areas. The eastern side of the park has been reworked into having a more gently sloping terrain. Torshovdalen is also considered archeological important place, as they have found memorabilia from early stone age (Byantikvaren, n.d).

The city of Oslo is both a county and a municipality, and the city of Oslo is one of the oldest and largest cities of Norway. Oslo is ~454 km² with almost 680 000 inhabitants making it the most populous region in the country as per 2018 (Thorsnæs, 2018). Only 150 km² is considered
urban and is approximately 1/3 of the entire Oslo region. Protected by law against urban development, is the ~300 km² forested areas surrounding Oslo, collectively called “Oslomarka” (Eriksson et al., 2016, Thorsnæs, 2018). By 2040, Oslo is expected to have an increase in population to 815514 inhabitants (SSB, n.d). Surrounded by rocky hills to the North, West, and East, Oslo is a secluded city contiguous to the inner Oslo fjord.

3.2 Geological framework.

3.2.1 Bedrock

Bedrock of Torshovdalen is a part of the Oslo region that was established in the lower Paleozoic ~542 million years ago (Worsley and Nakrem, 2008). The Oslo region is between 40-70 km thick and encompasses the lower Paleozoic succession. The lower Paleozoic is affected by thrusting and faulting and by upper Paleozoic plutonic activity (Bruton et al., 2010).

The Caledonian folding and Permian block faulting created the Oslo Graben. Oslo consists mainly of Precambrian basement rock, Cambro Silurian sedimentary rocks, Permian plutonic rocks and dykes (Løset, 2014). Depositions from the ferro-Scandinavian sea, created a thin layer of sediments over the Precambrian basement, which was later removed by erosion. The Cambro-Silurian succession is more than 2 km thick sedimentary rocks intertwined with Permo-Carboneferous magmatic rocks (Figure 3-1) (Worsley and Nakrem, 2008). During the Caledonian orogeny, the Cambro-silurian rocks were rifted, thrusted and folded and affected by igneous activity (Nakrem and Rasmussen, 2013). Cambro-Silurian is an acronym for the three geological periods in the Paleozoic Era: Cambrian, Ordovician and Silurian.
The Cambrian succession left dark shales denoted alum shale with alternating bitomus limestone with concretions of alum shale. Following the Cambrian succession, the Ordovician succession favored fossiliferous alternating limestone and greyish shale units with characteristic calcareous nodules (Nakrem and Rasmussen, 2013, Worsley and Nakrem, 2008). Beds of bentonite was also deposited from volcanic activity, intermittently deposited in-between beds of shale and limestone during middle Ordovician. (Worsley and Nakrem, 2008). These alternating shale and limestone were exposed to faulting that shifted units and changed the topography (Nakrem and Rasmussen, 2013).

Figure 3-2: Sedimentary rocks of the area of Torshovdalen. Black dotted line mark the rock boundary separating the sedimentary rock by a fault. Modified from (Graversen et al., 2017).
Surrounded by intrusive igneous rocks, from the Oslo paleo rift, Torshovdalen witness mostly rocks sequence of middle, and upper Ordovician age (Nakrem and Rasmussen, 2013). The sedimentary rock of Torshovdalen is divided by a bedrock boundary, a fault cutting diagonally through the middle valley. As described in Figure 3-2, the left-hand side of the fault constitutes the intermitted shale with nodular limestone beds. The right-hand side is denoted ungrouped sedimentary rocks, and also includes alternating shale and limestone. In general, they are both composed mainly of marine deposits belonging to Osen-Røa complex within the Oslo group, Voll- and Elnes formation (Figure 3-2)(Graversen et al., 2017). Elnes formation is characterized by a siliclastic strata: Fossiliferous mudstone with small ripple marks. Succeeding the Elnes formation, is the Vollen formation. This formation is characterized by a typical sequence of alternating shales and limestones (Bjørlykke, 2004, Nakrem and Rasmussen, 2013). Stratigraphic map of the Oslo-Asker region Torshovdalen (Figure 3-4) show the sequence of which the sedimentary rocks are deposited, Elnes an Vollen formation is emphasized by a red rectangle to indicate the expected formation in Torshovdalen.

The sedimentary sequence is folded with a SW-NE strike with varying dip angles. Formations in the Cambro-Silurian rocks are heavily affected by Permian eruptions (299-252 million year ago), where syenite porphyry, syenite, diabase and rhomb-porphry are common intrusive rocks Norconsult (2018). The igneous formations may follow the local structural layering or may be diagonal. The Cambro-Silurian sedimentary rocks have two dominant fracture directions. Fractures and joints display a common SW trending direction parallel to the strike of the local folding, spacing in between joints ranges between 0.05 cm in heavy fractures zones and 2-3 m apart. Joints har moderate to strongly affected by folding Fractures SE is highly angular, and often filled with calcite, and partly filled with clay, typical Spacing in between joints ranges between 0.05-3 m (Norconsult, 2018). Figure 3-3 illustrate the two dominant joint sets and how they occur in conjunction with the Cambro Silurian Sequence. Different zones of a fault lineament and fold hinges may experience high permeability compared to the surrounding rock (Morgan, 2004). Conductivity along fractures system depends on fracture connectivity (Braathen and Gabrielsen, 1998).
Figure 3-3: Conceptual model of the cambro-silurian sedimentary rocks. Alternating shale and nodular limestone demonstrate a strike in the SW-NE direction. Two dominant joint set are running perpendicular and parallel to the strike of the formation. Igneous intrusions is illustrated as running parallel to the strike. Modified from (Makurat et al., 1990).

Figure 3-4: Stratigraphy of the lower Palaeozoic of the Oslo-Asker Area. Red rectangle represent the formation of which Torshovdalen constitutes. B=Bentonite. Adapted from Nakrem and Rasmussen (2013)
3.2.2 Quaternary deposits

Much of Norway is dominated by exposed bedrock and/or with a thin layer of quaternary deposits. However, the areas located east and northeast of Oslo, have a larger sediment cover containing till, glaciofluvial, fluvial and marine sediments. Here, sediments are heavily affected by the postglacial marine limit, in which have altered the topography by fluvial erosion and clay slides (Figure 3-5) (Olsen et al., 2013).

Torshovdalen has variable thickness of marine deposits from 0.5 to several meters deep. Deposits also includes evidence from avalanches of quick clay (Nordahl-Olsen, 1993). Unconsolidated sediments are composed of fine sediments with few larger clast, poorly sorted. Bedrock is overlain by ~1m thick clay layer. Sediments above clay layer, is from variable sources. Upper ~1 m of Torshovdalen is mostly composed of sediments possibly not native to the area as it stems from the development of the underground sanitary sewer line that is found along the eastern side of the park. The topsoil is heavily affected by the recent urban development, and wires, concrete and glass have been found on site. According to (Nordahl-Olsen, 1993), concluded that grainsize distribution and variable thickness and sediment content indicate a bad infiltration rate and no groundwater potential within the uppermost unconsolidated layer.

![Figure 3-5: Map of Quaternary deposits in Oslo County and the adjacent marine limit (blue dotted line). Black dotted circle indicate location of Torshovdalen. Torshovdalen bedrock is overlain by thick marine deposits (light blue) and weathered materia (pink). Adjacent to Torshovdalen is Akerselva in which fluvial deposits are dominant (yellow) (Nordahl-Olsen, 1993).](image-url)
3.3 Torshovdalen sub-catchment

Figure 3-6: Nordvassdraget catchment hierarchy. Here, Akers elva Watershed (white), Torshovbekken catchment (grey), and Torshovdalen sub-catchment (Yellow rectangle). Akerelva is shown in blue, and the historic Torshobekken waterway is shown in red.
Torshovdalen is part of a larger catchment hierarchy of Nordmarksvassdraget specifically in the catchment of Akerselva (NVE, n.d). According to (VAV, 2018), Nordmarkvassdraget is 272 km², which encompasses ~256 km² of forested areas to the north, and 17.2 km² of urban areas to the south, whereas 11.6 km² is considered impermeable surfaces (VAV, 2018). Akerselva watershed is 14.84 km² with an average runoff of 17.65 L/s*km². Akerselva recharges from Maridalsvannet and discharges into Hovinbekken before it enters the Oslofjorden by Bjørvika (Figure 3-6) (NVE, n.d).

Torshovdalen is a sub-catchment within Torshovbekken catchment Figure 3-6. The watershed is ~5.15 km², and consist of 62 % urban development and 6.9 % forest (NVE, 2019). Average surface runoff is estimated to 17.6 l/s*km² (NVE, 2019). This watershed includes the south-facing part of Grefsenkollen following downgradient in the southwest direction encompassing Torshovdalen and discharges into Akerselva around the intersection of Thorvald Meyers gate and Trondheimsveien at the bottom of Grünerløkka.

3.3.1 Torshovbekken

Torshovbekken is a continuation of Disenbekken. In 1875, the work began with planning the closing of Torshovbekken. That has its original waterway trough Torshovdalen, over Rodeløkka and Grünerlokka and out into Akerselva. A health act from 1860 demanded that rivers and streams were to be closed due to the great health risk after waste water was dumped into the waterways. In addition, groundwater from Sofienbergparken would also bring deposits from the decomposed bodies from the cemetery. Several streams that ended up in Akerselva was lead in underground pipes, and among these were Torshovbekken (Moland, 2014).

The almost 3km long Torshovbekken was proposed culverted in an oval shaped brick sewer in the upper 490 m, and in a concrete-culvert in the bottom 530 m (Moland, 2014). The rest of Torshovbekken is possibly lead through either a brick sewer or a concrete sewer which was the popular sanitary sewer around 1890-1910 (Særgrov, 2013, Skaar, 2013).

3.3.2 Sanitary sewerage system in Oslo

Increasing urbanization in the 1800s suffered due to bad sanitary sewer infrastructure. Rivers and streams in Oslo were previously used as a place for sewer and waste and quickly became a health risk for the city’s inhabitants (Hvoslef, 2010). Rivers and streams have been culverted from 1600s to mid-1900s, and in Oslo, 67 % of the cities smaller rivers and streams were closed as of 2013, and 27 % of the main waterways. Some rivers were also partially closed and may
be seen at certain parts along the watercourse. The closing of the water courses had not only health benefits but also freed space for further development as rivers now were hidden in underground culverts (Oslo Kommune, 2015). A common practice was to combine the river and the sanitary sewer into a combined system were stormwater and sewer is placed in the same culvert. (Oslo Kommune, 2015) Today, a combined and a separated system is in place in Oslo leading the water away from residential and commercial areas to the waste water treatment plants (Figure 3-8)(Ødegaard et al., 2014). Table 1 show the extent of the sanitary sewer system within the Akerselva catchment , and Table 2 show the capacity of volume of water at which the sewer can handle.

![Conceptual model on combined and separate sanitary sewer system. Modified from Inspiregreen](image-url)

<table>
<thead>
<tr>
<th>Extent of sewer in the watershed</th>
<th>Km</th>
<th>Percent</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Total Sewer</strong></td>
<td>320</td>
<td>100</td>
</tr>
<tr>
<td><strong>Combined sewer</strong></td>
<td>188</td>
<td>58</td>
</tr>
<tr>
<td><strong>Waste water</strong></td>
<td>60</td>
<td>19</td>
</tr>
<tr>
<td><strong>Surface water</strong></td>
<td>74</td>
<td>23</td>
</tr>
</tbody>
</table>

Table 1: Sewer capacity of nordvassdraget watershed. (VAV, 2018)
Figure 3-8: Sanitary Sewer lines as they are placed in Torshovdalen sub-catchment.

Table 2: Hydraulic calculation for pipe capacity - full pipe

<table>
<thead>
<tr>
<th>Sewer Dimensions (mm)</th>
<th>Flow Velocity (m/s)</th>
<th>Flow Rate (l/s)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1400</td>
<td>1.73</td>
<td>2668</td>
</tr>
<tr>
<td>1200</td>
<td>1.58</td>
<td>1781</td>
</tr>
<tr>
<td>500</td>
<td>0.907</td>
<td>178</td>
</tr>
<tr>
<td>360</td>
<td>0.735</td>
<td>74.8</td>
</tr>
</tbody>
</table>
3.3.3 Local climate

Norway is located high in the Northern Hemisphere, and although one would expect a colder climate, Norway is relatively temperate due to the Gulf stream ocean current (Eldevik et al., 2015). The climate in Oslo is highly affected by southern winds in the summer, and northwestern winds during the winter. Winds are moderate because the city is protected by surrounding hills. During the winter, the horizontal and vertical exchange of air is relatively small, resulting in a higher risk of air pollution. (Dannevig, 2019) Climate research has over a period of 30 years (1961-1990) estimated seasonal mean temperature (Figure 3-9). Mean normal temperature measured -3.8°C during the winter months (December, January, February), and 15.6°C during the summer months (June, July and August), and a yearly mean temperature of 5.7°C. Mean Yearly Precipitation is ~760 mm where approximately ~280 mm and ~480 mm is estimated to fall as snow and rain respectively (Eklima). Norway’s climate is affected by high precipitation, temperature fluctuations, solar radiation and available snow (Engeland et al., 2004).

Figure 3-9: Mean monthly values based on a thirty year period from 1961 to 1990. Data is gathered from metrologic station of Blindern (Eklima).
4 Methods and materials

Figure 4-1: Red lines indicate sample transect for field campaigns 2018. Yellow star indicate that a sample was collected from the specific location location.
Field work was conducted throughout the summer and early fall of 2018. Research and preparation were done during spring semester. Methods for field work were chosen to characterize and map groundwater in the study area and to establish a water budget. Data supplied by NGI was analyzed in their laboratory, and procedures were fulfilled according to international and Norwegian standards.

4.1 Well construction

Three environmental wells were installed in Torshovdalen early September 2018. Norwegian geotechnical institute (NGI) was engaged to investigate the geotechnical properties of the soil and installing wells for groundwater monitoring. The wells were installed at locations stretching across the entire park; North, South and an eastern location. Location of wells were decided in cooperation with the water and sewerage works of Oslo (VAV) and Bymiljøetaten (BYM). Wells were financed and approved by the water and sewerage works through chief engineers Tharan Fergusson and Julia Kvitsjøen with guidance and supervision from academic professionals Anja Sundal, Clara Sena from the university of Oslo. Wells were finalized with an NS-EN 124 certified 15.5 cm2 cast-iron square-framed manholes covering the installation, commonly called “gategutter” (Figure 4-5:Left). Manholes were chosen based on the classification of the areas foot-traffic and location. The frame of the manhole was fastened above top of casing, however, leveled to the ground surface. A wider hole in the top 10 cm was dug around the well casing in order to properly fit the manhole.

To collect data, total sounding was performed using a heavy-duty soil investigation hydraulic apparatus, Gm100GT drilling rig. The method combines rotary-pressure-sounding and rock drilling (Vegdirektoratet, 2018). In Norway, geotechnical specialists frequently utilize the total sounding technique and the method is particularly useful in and mapping soil

Figure 4-3: principle sketch of total sounding mechanism. Modified from Vegdirektoratet, 2018 #41@@author-yeFigure 4-3ar].
conditions and determining depth to bedrock (Haugen et al., 2016). The apparatus is used by vertically placing the drill tower above borehole location. Total sounding is performed by initiating a rotational movement and penetrating force and on a 2 m rod (Figure 4-3) (Vegdirektoratet, 2018). Fluid was flushed simultaneously through the rod to ease the drilling. Water was used as drilling fluid at well 1 (North) and well 2 (East). A mix of polymer and water was used at well 3 (South). At well 1 (North) and well 3 (South), two boreholes per location were made: one for the installation of wells, and one for geophysical wireline logging. At well 2 (East), only one borehole was produced. Figure 4-4 display the process of drilling in Torshovdalen and how it occurred.

![Figure 4-4: Left: Trond and Audun from NGI performing total sounding in Torshovdalen. Audun is attaching new rods as the drill bit is going down. Middle: Gm100GT drilling rig uses water to flush the borehole when penetrating the ground in order for the rods to descend with ease. Right: A Large casing is fully submerged in the ground.](image)

Each well is constructed with a solid PVC casing placed in the unconsolidated sediment layer (Figure 4-5). Filter casing with perforations of 0.2 mm was placed in the bedrock at all locations. Each well was produced with a 63 mm drill bit. Well Grouting was placed in annular space between borehole and well casing. The grain size of the sand was 0.4-0.8 mm in diameter and placed in the sediment bedrock boundary. Well 3 (South) was not completed with grouting. In addition, data was also collected using a sonic rig. This method uses vibration to penetrate the ground. With a frequency of ~150 vibrations per seconds it is thus suitable for hard ground covers such as bedrock penetration (NGI, 2018). During well construction, samples from the bedrock was collected as cuttings from drilling was brought to the surface: One sample representing bedrock in well 1 (South) and two samples from well 2 (East).
Table 3. Well specificities.

<table>
<thead>
<tr>
<th>Well</th>
<th>Drilling rig</th>
<th>Technique</th>
<th>Drill bit Diameter (mm)</th>
<th>Drilling fluid</th>
<th>Casing (mm)</th>
<th>Sand and Bentonite</th>
</tr>
</thead>
<tbody>
<tr>
<td>1 (North)</td>
<td>Eijkelkamp sonicsampdrill</td>
<td>Vibration Total</td>
<td>102</td>
<td>Water</td>
<td>217</td>
<td>X</td>
</tr>
<tr>
<td>2 (East)</td>
<td>Gm100GT</td>
<td>sounding Total</td>
<td>67</td>
<td>Water</td>
<td>115</td>
<td>X</td>
</tr>
<tr>
<td>3 (South)</td>
<td>Gm100GT</td>
<td>sounding</td>
<td>67</td>
<td>Polymer mix</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

Figure 4-5: Left: PVC well casing is assembled for illustration. Middle: Well casing is being put together by Trond and Audun from NGI. Right: Finalized construction with well cap inside a Cast-iron squared frame. Manholes commonly called “gategutter”
4.2 Geophysical methods

4.2.1 Instrumental well logging

Boreholes for instrumental wireline logging were produced next to the wells at the northern and southern location. Geophysical logging was performed by NGI. Geophysical methods is used to map sediment layering, thickness of the sediments, bedrock stability and subsurface weaknesses (Vegdirektoratet, 2014). Wireline logging was performed at location well 1 (North) and well 3 (South) and was obtained after the investigative drilling, however, before the installation of well 1(North) and after the installation of well 3(South).

The instrument constituted of a sensor that was lowered to the bottom of the well. Wireline logging continuously record the walls of the borehole as the instrument is retrieved back to the surface. As the instrument is moving upward, it uninterruptedly measures the properties of the surrounding rock or sediment layers (Boggs, 2011). Several instrumental logs were used, such as: Acoustic Image log, Optical tele viewer, electrical resistivity, and Natural gamma (Table 4). These procedures measure the variation in the rock properties and will reflect features such as mineralogy, fluid content, porosity, and lithology (Boggs, 2011).

Electrical resistivity yields 2D information on zones and direction of weaknesses in bedrock. The benefit of this approach is that it may detect a difference in density of the rock, providing information about layers of high and low permeability (Varhaug, 2016). Acoustic image logs use soundwaves to map stratigraphy and fracture orientation (Crain, n.d). This method uses a rotating receiver to record the travel time of sound waves. This method also incorporates the amplitude of said signal in which detects the resistance of the rock. The Optical tele viewer uses a camera to depict the walls of the borehole. A source of light aids the camera in capturing details in sediment and bedrock. For both acoustic and resistivity measuring, it is important that the borehole is sufficiently filled with water. For the optical tele viewer, if water is present, it must be optically clear in order for the instrument to capture clear images (Vegdirektoratet, 2014). Bulk density of rock is measured by emitting gamma rays onto the walls of the borehole. Electrons are scattered and ultimately colliding with each other. The collision of electrons provides insight of weather the formation is of high (low number of backscattered electrons) or low (high number of backscattered electrons) density. Natural gamma measures naturally occurring radioactivity, primarily from Potassium (K), Thorium (Th) and Uranium (U) usually found in clay minerals (Varhaug, 2016).
Most well logs are able to detect fractures (Crain, n.d). In particular, the acoustic tele viewer is helpful in detecting the presence of fractures. The acoustic tele viewer provides a 360° image of the borehole in which demonstrate fracture characteristics. Fractures and geological features are demonstrated as sinusoidal waves in the logs, from this wave it is possible to determine the strike and dip, and the aperture of the fracture (Laongsakul and Durrast, 2011). Roughly estimating the aperture of a fracture may be found by increasing the vertical scale using AutoCAD.

<table>
<thead>
<tr>
<th>Name</th>
<th>Description</th>
<th>Tool</th>
<th>Unit</th>
</tr>
</thead>
<tbody>
<tr>
<td>Amplitude</td>
<td>amplitude of reflected acoustic signal in 360°</td>
<td>BHTV</td>
<td>-</td>
</tr>
<tr>
<td>Caliper (Radius)</td>
<td>360°-hole radius, computed from acoustic travel time</td>
<td>BHTV</td>
<td>mm</td>
</tr>
<tr>
<td>Natural Gamma</td>
<td>Smoothed Total natural gamma, moving average over 20 cm</td>
<td>NGAM</td>
<td>cps</td>
</tr>
<tr>
<td>Natural Gamma</td>
<td>Total natural gamma</td>
<td>NGAM</td>
<td>cps</td>
</tr>
<tr>
<td>Optical</td>
<td>360° oriented optical image</td>
<td>OPTV</td>
<td>-</td>
</tr>
<tr>
<td>Resistivity (Deep)</td>
<td>Focused resistivity</td>
<td>Deep DLL3</td>
<td>ohm-m</td>
</tr>
<tr>
<td>Resistivity (Shallow)</td>
<td>Focused resistivity</td>
<td>Shallow DLL3</td>
<td>ohm-m</td>
</tr>
</tbody>
</table>
4.3 Data sampling

4.3.1 Auger sediment sampling

The Gm100GTT drill rig was used to aid the gathering of soil and clay samples from all well locations (0-4 m depth) using a 2 m long auger with a rotating helical screw blade which works as a material conveyor (Figure 4-6). The auger was attached to the drill rig and pierced into the ground. At retrieval of the Auger, materials were trapped in between screw blades as it was pulled from the ground. Sediments from approximately every 10 cm were stored in Ziploc bags and brought to the laboratory at the department of geology at the University of Oslo for grain-size analysis.

4.3.2 Cylinder core samples

Geotechnical properties of the soil and sediment may be determined by collecting undisturbed samples. Hollow cylinders of 80 cm length and 72 mm diameter were used to collect four undisturbed sediment sections at well 3 (South) location. Using the Gm100Gt Drilling rig, hollow cylinders were lowered into the ground in a closed position to avoid materials being collected. When satisfactory depth was reached, a stamp was released opening the tubes. When further penetrating the soil, materials was forced into the cylinder. Samples were later analyzed at the NGI laboratories. Three undisturbed cylinder samples from 4 -6.8 m depth was collected.

Density ($\rho$) is calculated as an average of the whole cylinder sample. Evaluation of sample properties were performed with respect to Norwegian/international standard, NS-EN ISO 17892-2:2014. Specific weight ($\gamma$) is measured as a relationship between weight and total volume, where $g = 9.807\text{m/s}^2$. Total length of the sample denoted (L), and Volume (V)(NGI, 2019). Specificities of the core samples are showed in Table 5.
Table 5: Cylinder core specificities as noted by NGI laboratory staff (NGI, 2019).

<table>
<thead>
<tr>
<th>Depth (m)</th>
<th>Diameter (mm)</th>
<th>Sampling</th>
<th>Lithology</th>
</tr>
</thead>
<tbody>
<tr>
<td>4.0-4.8</td>
<td>72</td>
<td>Piston</td>
<td>Clay, lost part of the lower section</td>
</tr>
<tr>
<td>5.0-5.7</td>
<td>72</td>
<td>Piston</td>
<td>Clay, lost 10 cm of the lower section</td>
</tr>
<tr>
<td>6.0-6.8</td>
<td>72</td>
<td>Piston</td>
<td>Clay</td>
</tr>
</tbody>
</table>

4.4 Laboratory experiments

4.4.1 One axial compression test

The axial compression apparatus tests the relationship between compression and strain and is estimating when soil sample cannot longer withstand the force acted upon it. On the contrary, axial strain was decreasing with depth. This is implying that Failure with decreasing strain indicate a lower water content, and that lower porewater pressure is significant in samples that reach a failure point with low strain (Mun et al., 2016). Overburden pressure is reducing the bulk volume by particle rearrangement. This will increase the intrinsic forces and thus shear strength is increasing with depth (Moore, 1964).

Sample was inserted into the instrument between a plate and a pressure-stamp Figure 4-7. Pressure is exerted on the sample to measure its ability to withstand strain. Failure point is reached when material cannot withstand the amount of pressure and a rupture point is established. Half of the strain after rupture yields the shear strength (Vegdirektoratet, 2016). Shear strength (Cu) is measured with a one axial compression (qu), and axial strain (ε). Tests are performed according to regulations: NS 8016:1988 and NS-EN ISO 17892-7:2017, specific for material with low permeability. Water content (w) is estimated according to ISO 17892-1. Whereas Specific weight (γ), Density (ρ), and dry-density (ρ_d) is calculated with respect to the international standard ISO 17892-2 (NGI, 2019).
Figure 4-7: Principle sketch of the One axial compression apparatus. Samples is placed between a plate and a stamp before being exposed to pressure. Modified from Vegdirektoratet (2016).

4.4.2 Cone penetrometer test

Undrained and remolded shear strength was measured using a cone penetrometer apparatus. Cone experiment is an empirical method to determine the strength of soil samples, this in order to classify the sensitivity of the clay. Shear strength is measure by the penetration of a metal cone of different weight and diameter into the sample. Test is performed twice: Once on the undisturbed sample and once on the sample after being remolded.

Figure 4-8 illustrates the specificities around the cone penetrometer test. Here, an instrumental cone was exerted on the soil sample. Results are measured as to how far the instrumental cone of a certain weight and degree angle will penetrate the soil sample. The tip-angle and diameter of the cone will affect the resolution of the results. The penetration of the cone is done at least three times, and the mean value determine the shear strength. Next step is to add water to the soil sample and stir until a homogenous mass. Cone experiment is then repeated on the remolded sample (Vegdirektoratet, 2016). According to NGI standard procedure index, quick clay is determined by its plastic characteristics to be < 0.5 kPa and with a cone penetration of > 20mm.
Cone penetrometer technique was performed according to NS-EN ISO 17892-6:2017 and Norwegian standard NS 8015:1988. Penetration range is according to ISO 17892-6 for 4-20 mm. Sensitivity (St) is evaluated according to Norwegian standard NS 8015:1988 (NGI, 2019).

![Cone penetrometer](image)

*Figure 4-8: Simple principle sketch of how the Cone penetrometer is applied. A cone is lowered onto the sample to measure to which depth it is possible to penetrate with respect to cone weight. Modified from Vegdirektoratet (2016).*

**Sodium chloride (NaCl) and electrical conductivity**

This method measures the electrical conductivity and the NaCl content in the soil sample within the range of 20 to 5000 µS/cm. Three samples from cylinder 1, 2 and 3 were inspected for its NaCl content from Well 3 (South). Weight of sample and percent water content was measured. NaCl equivalents and conductivity is found by inducing ion exchange. Water is added to the sample to dissolve the salts and other soluble substances in the sample. Electrical conductivity is measured based on the NaCl content from the dissolved water sample. NaCl equivalents are given in (g/kg). NaCl equivalents in porewater (g/l). and conductivity is measured in (µS/cm). Conductivity calculations is achieved with a sample-water relationship of 1:1 (Skår, 2019).
4.5 Supplementary data sampling

4.5.1 Disturbed sediment sampling

Figure 4-9: Left: Sediments are trapped inside the open-faced sediment catcher. Right: Sediment sample was easily molded demonstrating a high clay and soil moisture content.

Sampling of sediment was conducted using a 1.5 m long manual auger and was intended for shallow depth sampling for this study (Figure 4-9). The benefit of this approach is that it is easy to use and does not rely on heavy machinery. The instrument is an open-faced stony soil auger with a sediment catcher of 5 cm diameter. Sediment samples were collected at twelve locations along four transects, transects are depicted in. Approximately 300 g of disturbed material was collected at a depth of ~0.30 m at 3-4 locations per transect. Retrieval of samples was prosecuted by manually rotating and simultaneously pushing the auger into the ground. The open-faced sediment catcher would collect sediments as the auger was pushed further into the ground. Samples were processed in the laboratory at the department of geology at the university of Oslo: Grain size analyses was conducted for grainsize > 2 mm, and a particle size analysis was performed for all remaining grainsize < 2 mm. A Soil texture triangle produced by the National Resource Conservation Service was used to place samples in a soil texture groups (NRCS, n.d-a).
4.5.2 Modified Philip-Dunne infiltrometer

Method for determining rate of infiltration was performed using a Modified Philip Dunne (MPD) falling head infiltrometer. According to (Ahmed et al., 2014, Nestingen, 2007, Solheim, 2017), the Modified Philip Dunne infiltrometer is a simple yet effective method for point measuring of infiltration rates at the surface. The benefit of this approach and the reason why this method was chosen was because of its applicability and reliability in the field. Also, recent studies by (Solheim, 2017)and (Nestingen, 2007) concludes that the method is suitable for measuring the hydraulic conductivity at cite specific areas, such as rain gardens for the assessment of stormwater related issues. The flow geometry in the MPD, as seen in Figure 4-10, is resembling a capped sphere forcing the water to flow in a horizontal and vertical direction (Nestingen, 2007, Ahmed et al., 2014). According to Nestingen (2007), the depth at which MPD is buried have great impact on the infiltration rate.

Figure 4-10: Principle sketch of the MPD instrument. Water is poured into the cylinder to a certain height. As water is infiltrating, the water will flow in all directions through the soil. Modified by Julie Fjeldseth from Nestingen (2007).

Figure 4-11: Left: Ongoing Infiltration testing at Torshovdalen using a stopwatch to note the time of infiltration. Middle: MPD device filled with water. Water is murky due to lose sediments in suspension. Right: Equipment used when performing infiltration, soil moisture and sediment sampling in Torshovdalen
The MPD is an open ended, PVC cylinder with slanting edges for easy penetration through the soil. A 10 mm diameter hole-dozer was used to create an effortless path of insertion for the cylinder. The cylinder was equipped with a measuring tape that was fastened to the inside of the tube. The MPD is 50 cm long with an outer diameter of 10 cm.

Fourteen point-infiltration tests were performed along four transects (Figure 4-1), located throughout Torshovdalen. The MPD-tube was inserted 5 cm into the ground (Figure 4-10) and filled with water to the ~43 cm mark, which marked the beginning of the test. Approximately 5 L of water was needed per test. Each test was performed with a minimum of one-hour observation time. Height of cylinder is read off the measuring tape inside the tube as the water was infiltrating down. According to Munoz-Carpena, et al., 2002 as cited by (Nestingen, 2007), only three measurements are necessary to record in able to evaluate the infiltration rate, however, continuous measurement should be done to increase the precision (Figure 4-11). The method should be performed until a stable K_{sat} value is achieved (Solheim, 2017). Soil moisture content, conductivity, and temperature was recorded before and after each infiltration test using a Delta T HH2 moisture device with a WET-Sensor.

Data obtained from infiltration testing was used to estimate saturated hydraulic conductivity of the top soil. The change in height of consecutive measurements over the change in time yield the initial infiltration rate (Solheim, 2017):

\[ I_{ini} = \frac{\Delta h}{\Delta t} \]

Where \( I_{ini} \) is the initial infiltration (L/t), \( \Delta h \) is the change in height of the cylinder (L) and \( \Delta t \) is the change in time (t) corresponding to the change in height.

The infiltrating water will penetrate the ground producing a wetting front of a capped sphere. Adjustments must be made to account for the horizontal and vertical movement of water out of the MPD through different soil textures. Solheim (2018) in collaboration with the water and sewerage works of Oslo municipality (VAV), proposed a correction factor for which different soil textural properties are based. To increase the reliability of measures and to avoid overestimating \( K_{sat} \) values following equations were proposed: Eq 4-2 should be used to correct for soils with high clay content and silty material and initial infiltration rates ranges between 7-53 cm/h. Whereas Eq 4-3 should be used for soils with coarse grained silty-sand and sandy loam and initial infiltration rates ranges between 25-73 cm/h (Solheim, 2018, Solheim et al., 2017).
\[ K_{\text{sat}} = 0.6 I_{\text{ini}}. \quad \text{Eq 4-2} \]

\[ K_{\text{sat}} = 0.8 I_{\text{ini}}. \quad \text{Eq 4-3} \]

Solheim et al. (2017) identified several advantages for using correction factor in conjunction with MPD measurements. It was therefore decided that a correction factor of 0.6 should be adopted to this study for estimating \( k_{\text{sat}} \).

### 4.5.3 Ground penetrating radar (GPR)

Ground penetrating radar (GPR) was used to map the underground structures in the study area. Execution of the GPR was performed by Senior engineer, Trond Eiken from the university of Oslo. GPR was recorded at two different field campaigns: Fall 2018 and Winter 2019 (Figure 4-12). The objectives for this method was to determine depth to bedrock, in order to compute a hydrogeologic conceptual model. Different antennas were used with different frequencies: 50, 250 and 500 MHz. These antennas were able to capture images at different depths and with different resolution for evaluation of the subsurface and thickness of the unconsolidated sediment layer (Mussett and Khan, 2000). To capture as much details about the subsurface as possible both the 250 MHz and 500 MHz antennas were used throughout the park. Figure 4-12, illustrates the extent of the path of the GPR field campaigns.

GPR uses electromagnetic energy to detect subsurface structures and layers (Beres and Haeni, 1991). The electromagnetic energy is transmitted through the ground, from antennas carrying different frequencies. Some energy will reflect to the antenna when sensing different dielectric properties in the ground, the remaining energy will continue downward until it reaches another interface (Mussett and Khan, 2000). Ground penetrating radar produces continuous profiles of the subsurface. Results are amplified and converted into audio-frequencies. Reflected energy will yield total travel time of the energy from the antenna, through the subsurface, and reflection back to the antenna.
Figure 4-12: GPR profiles recorded in the fall of 2018 is shown in red, and field campaign of winter 2019 is shown in black. The entire park was covered with respect to well locations and pre-selected transect.

4.5.4 Well purging and groundwater level measurement

Water level measurements were conducted during- and after the installation of wells. Water level measurements and groundwater quality are parameters that are easily accessible in domestic and monitoring wells (OSMRE, 2012). To obtain accurate groundwater quality samples, pumping a specific amount of water out of a well before sampling is necessary. This will ensure that new fresh water will enter the well and water rebound rate data is readily available. Purging a well three times its volume is considered sufficient, however, having a high influx of water, further purging should continue (OSMRE, 2012).

To determine the flux and quality of water, an Eijkelkamp agrisearch 12 vdc peristatic pump was used to purge water out of the wells. The apparatus can be operated on both gas and liquid in which a flexible tube, a clamp and rollers control the incoming fluid. The pump will always be constricting the flexible tube, which means that even though the pump is not working, the flexible tube is still compressed by the rollers inhibiting liquid from flowing back into the tube. The rollers will reel the fluid in from out of the well, through the flexible tube and subsequently expelled. Rate of revolution will determine at which speed the water is extracted.
The flexible tube will rebound after the roller has passed and start the process of sucking in new fluid all over again (Eijkelkamp, 2018).

In this study, an 11 m long hose of 1 cm inner diameter and a thickness of 2 mm was attached to the pump (Figure 4-13:Left). Once the hose was lowered into the well below the water level, the apparatus was initiated. Well 1, 2 and 3 was purged one time its volume before timing the in-flux of water. A Water level meter with a built-in probe that detects fluid conductivity was used to measure the water level before and after purging of the wells to note the in-flux of water. Water level measuring was repeated three times to obtain a mean value (Figure 4-13:Right). Volume of water in the wells was calculated using following equation:

\[ V_{cyl} = \pi * r^2 * h \quad Eq \ 4-4 \]

Where \( r \) is the radius of the well (L), and \( h \) is the height of the water column (L). height of the water column is found by subtracting the depth to water by the depth of the well. The rate of influx is calculated by continuously measuring the water level over time. Rate of influx may be obtained by following equation:

\[ Rate \ of \ in-flux = \Delta water \ level / time \quad Eq \ 4-5 \]

Where \( \Delta water \ level \) is the change in water table measurements (L) over time (t).

Figure 4-13:Left: Ingrid is lowering the hose into the well to use the peristatic pump. Right: Manually measuring the water level using a water-level meter listening for the built in signalling sound indicating that water is reached. Photo: Clara sena
**Conductivity, temperature and depth (CTD) datalogger**

An Eijkelkamp CTD-diver datalogger was deployed in well 3 (South) for continuously monitoring of pressure (cmH2O), temperature (°C) and conductivity (mS/cm). The diver was fully submerged under water at approximately 1300 mm into the well. Water level was measured manually with a water-level meter with a built-in signaling system for when water is reached before and after the diver was submerged. A reading unit for diver was used to transfer data from the diver to the computer. The diver was pre-instructed to have a fixed-length memory and record interval every 15 minutes. Pressure range was 50mH2O, and conductivity range was 120 mS/cm.

The datalogger is housed in a 22 mm diameter sealed cylindrical ceramic casing of which makes up the diver. Measurements are made with respect to the predetermine time steps and ultimately stored in the internal memory of the diver. The purpose of the diver is to measure water pressure, temperature and conductivity. The diver is equipped with a pressure sensor to measure the water pressure and a semiconductor temperature sensor. In addition, the CTD diver measures electrical conductivity. Advantages of measuring temperature is the information it can provide about groundwater flow and determining the distribution of polluted water. The diver measures the absolute pressure to determine the height of the water column.

The figure below illustrates the parameters for estimating the water level from a well in which a diver has been installed. Where water level measurements are obtained by subtracting the water column (WC) by the length of the cable (CL). A barometric diver was not installed and the atmospheric pressure (pbaro) must be obtained manually from online resources. The diver measures the pressure exerted on the diver (pdiver) by the water column and the atmospheric pressure to establish the height of the water column.

\[
WC = 9806.65 \frac{p_{Diver} - p_{Baro}}{\rho * g} \quad \text{Eq 4-6}
\]
Where WC is the water column (cm), $P_{\text{diver}}$ is the pressure in (cmH2O) retrieved from the diver, $P_{\text{baro}}$ is the pressure in (cmH2O) retrieved from atmospheric pressure data, $g$ is the acceleration due to gravity (9.81 m/s$^2$) and $\rho$ is the density of the water (1,000 kg/m$^3$). The water level is estimated by subtracting the water column from the cable length using the following equation, Where WL is water level (L), $CL_{\text{measured}}$ is the manually measured length of the cable.

\[ WL = CL_{\text{measured}} - WC \]  \hspace{1cm} \text{Eq 4-7}

*Figure 4-14: Water level calculations.* $WL =$ water level, $WC =$ water column, $CL =$ Cable length. Modified from Schlumberger (2014).
4.5.5 Monitoring of snowmelt

Snowmelt was monitored every other week in 2018 (February, March, April). Pictures and notes were taken explaining the current conditions as well as climatic data. Snow remained in the valley until April 20th, 2018.

![March 14th](image1)
![March 25th](image2)
![April 13th](image3)

Figure 4-15: Pictures illustrate how the snow melted in Torshovdalen in 2018. Pictures show Torshovdalen from the southern part looking north.

4.5.6 Soil moisture

Soil moisture sampling was performed using a delta T soil moisture HH2 Wet-2 sensor kit. The sensor comprises of three rods, positioned in a row, which was inserted into the soil for measuring. The sensor has an overall dimension of 120 x 45 x 13 mm. The wet sensor collects data from the surrounding soil with a volume of 500ml. Soil moisture was measured at 10 cm depth every 2 m along four transects, and nine measurements in total were completed between May-Nov of 2018 (Figure 4-1). By inserting the probe into the ground every two meter, a profile data record of soil water content was obtained.

The Wet-2 sensor is mostly used for horticulture but also within soil science research to measure soil properties. The sensor measures soil water content, pore water, electrical conductivity and temperature. It can be calibrated with different mineral, sand and clay factors depending on location and assumptions about the soil (Delta-T, 2007). The sensor generates a 20 mHz signal applied to the central rod and produces a small electromagnetic field. The electrical field will respond to the dielectric properties within the soil. An HH2 moisture meter attached to the sensor, will receive and further calculate the data obtained by the responding dielectric properties (Delta-T, 2007).
Table 6: Parameter for measurements for the delta probe, modified from Delta-T, 2007.

<table>
<thead>
<tr>
<th>Measurement</th>
<th>Volumetric water content</th>
<th>Temperature</th>
</tr>
</thead>
<tbody>
<tr>
<td>Accuracy</td>
<td>± 0.03 m³.m⁻³ (1%)</td>
<td>± 1.5°C</td>
</tr>
<tr>
<td>Soil moisture measure</td>
<td>Full accuracy over:</td>
<td>0 to 50°C</td>
</tr>
<tr>
<td>range</td>
<td>0 to 1 m³.m⁻³</td>
<td></td>
</tr>
<tr>
<td>Sample volume</td>
<td>~500ml</td>
<td>Not available</td>
</tr>
</tbody>
</table>

4.6 Methods of analysis

4.6.1 Grain size analysis

The hand sieving method

Twelve sediment samples from along each soil moisture transect, and five samples from well 2 and 3 were taken for grain size analysis and was performed on samples for particles > 2 mm diameter.

Soil samples were poured into a stack of sieves with mesh sizes ranging 5 mm - 2 mm in descending order, and a solid bowl at the bottom to catch particles < 2 mm. The sieve-stack was manually shaken for three minutes for each sample. This would separate different particle sizes into sieve-fractions. Samples were removed from the sieves of its corresponding mesh-size and weighed. The data acquired was later combined with results from the particle size analysis. Samples were handled individually, and sieves were cleaned after each sample.

The wet sieving method

A Beckman Coulter LS13 320 apparatus with a Fourier lens was used to perform particle size analysis. The Beckman Coulter instrument uses laser diffraction to determine grain size distribution for particles between 0.4 µm- 2 mm (Coultier, 2017). This method measures scatter pattern diffracted by particles in suspension. Smaller angles will yield a higher angle of diffraction, whereas larger particles will yield a lower angle. The Fourier lens Figure 4-16, which is sensitive to the angle in which light is reflected, will group and categorized particles within ranges of the diffractive angles (Coultier, 2017).
A small fraction of sample, 0.18g, was poured into a small beaker containing 5-10 mL of 5% Calgon Carbon. The beaker was then placed in an ultrasonic high frequency pool for 3 minutes. Sample was poured into a compartment in the apparatus containing approximately 1 L of water. The desired amount of sample was adjusted on the fly depending on the sample texture. The refining process was highly sensitive to the amount of sample inserted. Finer particles needed less sample material to be used. Using guidelines from Table 7: Minimum amount of sample needed, depends on median diameter of the particle. Water inside the apparatus was continually flushed and rinsed between each sample. The process was repeated 2-3 times per sample to obtain a mean value. Measuring’s were continuously monitored by means of accompanied software to establish the accuracy and to decide whether sample needed additional processing.

<table>
<thead>
<tr>
<th>Median diameter (µm)</th>
<th>Weight (g)</th>
</tr>
</thead>
<tbody>
<tr>
<td>10</td>
<td>0.2</td>
</tr>
<tr>
<td>50</td>
<td>0.4</td>
</tr>
<tr>
<td>70</td>
<td>0.6</td>
</tr>
<tr>
<td>100</td>
<td>0.8</td>
</tr>
<tr>
<td>300-400</td>
<td>1.0</td>
</tr>
<tr>
<td>600</td>
<td>1.5-2.0</td>
</tr>
</tbody>
</table>

Figure 4-16: Details on the Fourier Optics in which the Beckman Coulter particle size analyzer. (Coulter, 2017).
Saturated hydraulic conductivity ($K_{sat}$)

Saturated hydraulic conductivity ($K_{sat}$) was evaluated using available soil properties such as grain size and particle size analysis (Abdelbaki et al., 2009). Pedotransfer functions (PTF) is a term used for transferring available data into useful empirical solutions from soil hydraulic characteristics (Abdelbaki et al., 2009, Wösten et al., 2001). PTFs works as predictors where there are missing parameters (Wösten et al., 2001). Several empirical formulas are available, however, two well know methods have been chosen for this thesis: Hazens method which emphasizes on coarse grained soils and Puckett 1985 for fine grained soils.

Hazens method is a well-known method for estimating $K_{sat}$. This method is appropriate especially for soils with grain size between 0.1 -3.0 mm (Fetter, 2001, Koenig, 1911), and is relating the hydraulic conductivity to grain size diameter where the 10th percentile of the particles are of finer material denoted $d_{10}$ (Shukla, 2014).

$$K = C(d_{10})^2 , \quad d_{10} \text{ is between } 0.1-3.0 \text{ mm} \quad Eq 4-8$$

Where C is an empirical coefficient based on soil characteristics, Table 8, $d_{10}$ is the 10th percentile (cm), and K is in cm/s.

<table>
<thead>
<tr>
<th>Material</th>
<th>Coefficient (C)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Very fine sand, poorly sorted</td>
<td>40-80</td>
</tr>
<tr>
<td>Fine sand with appreciable fines</td>
<td>40-80</td>
</tr>
<tr>
<td>Medium sand well sorted</td>
<td>80-120</td>
</tr>
<tr>
<td>Coarse sand, poorly sorted</td>
<td>80-120</td>
</tr>
<tr>
<td>Coarse sand well sorted</td>
<td>120-150</td>
</tr>
</tbody>
</table>

Another empirical formula to use for estimating $K_{sat}$ was an equations developed by Puckett 1985 as cited by (Alexander, 2009). This formula was created with respect to the percent clay content in the sample, and $k_{sat}$ may be determined by using the following formula:

$$K_{sat} = 4.36 \times 10^{-3} e^{-0.1975 \text{CL} \%} \quad Eq 4-9$$

Where CL% is the clay content in percent (%), and $K_{sat}$ is in cm/s.
Two empirical methods for calculating $K_{sat}$ (Hazen and Puckett 1985) were used to compare possible deviations. Samples analyzed displayed a high fine soil content which complicated the use of Hazen’s formula which is suitable for sandy soils with an effective grain size ($d_{10}$) falling between 0.1-3.0 mm range (Fetter, 2001). Reading off the cumulative plot, all samples except L3-T3, L3-T2 and all samples from well 3 (South), had an effective grain size < 0.002 mm. However, the $d_{10}$ value may be found mathematically by assuming a linear increasing trend. By solving for $a$, one may establish the effective grain size for the 10th percentile. Where $a$ is the 10$^{th}$ percentile (mm), $y$ is the lowest cumulative grain size (%), and $x$ is the lowest grain size (mm) when $b=0$.

\[ y = ax + b \hspace{2cm} Eq \ 4-10 \]

\[ a = \frac{y}{x} \times 10 \hspace{2cm} Eq \ 4-11 \]

**Frozen soil**

The risk at which the soil is susceptible to frozen ground can be evaluated based on grain size distribution Figure 8-10. Sediments are categorized into four risk zones based on their content different soil texture: T1, T2, T3, and T4, where T1 has no risk, and T4 has a high risk. Risk of frozen soil is specified as the soil's vulnerability for capillary forces to pull water into the frost zone and increasing the surrounding frost lenses (Lilleøre, 2018, Johnsrud et al., 2013, Andersland and Ladanyi, 1994). Frozen ground was determined using the grain size distribution curve (Johnsrud et al., 2013). See appendix H for further information.

4.6.2 Sample preparation

*Table 9: XRD and XRF was performed on following samples. Ten samples were evaluated: Six from Well 1 (North) three from well 2 (East) and one from well 3 (South). * indicates that a clay analysis was performed on the samples.*

<table>
<thead>
<tr>
<th>Sediment</th>
<th>Well 1</th>
<th>Depth (m)</th>
<th>Well 2</th>
<th>Depth (m)</th>
<th>Well 3</th>
<th>Depth (m)</th>
</tr>
</thead>
<tbody>
<tr>
<td>B1-L1.5</td>
<td>0-1.5</td>
<td>B2-L1</td>
<td>0-1</td>
<td>B3-L4*</td>
<td>3-5</td>
<td></td>
</tr>
<tr>
<td>B1-L2.5</td>
<td>1.5-2.5</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>B1-L3.5</td>
<td>2.5-3.5</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Bedrock</td>
<td>B1-F1</td>
<td>3.5</td>
<td>B2-F1*</td>
<td>3-4</td>
<td></td>
<td></td>
</tr>
<tr>
<td>B1-F1.5</td>
<td>3.5-4.5</td>
<td>B2-F4</td>
<td>4.5</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>B1-F4</td>
<td>4.5-5.5</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>
Freeze-dry sediments

For certain analysis with sediment samples containing finer material, it is important that the sample is freeze dried rather than oven dried. This will ensure that important sediment properties are undamaged (Flink and Knudsen, 2002).

The freeze-drying method is a process of three steps. First step includes the sample being placed inside a freezer at -25 °C for a couple of days to ensure that the water is frozen solid. Second step involves removing the ice from the sample, which was completed in a Christ Alpha 1-4 LD lyophilizer. The process of freeze-drying uses sublimation to gently exhaust the sample so thermally sensitive material are preserved. This allows the water content of the sample to go from solid phase to gas phase. Final step constitutes of further converting the sorbed water into vapor by means of desorption (Flink and Knudsen, 2002).

Ten unprocessed samples (Table 9) were poured on an individual sheet of paper and thoroughly mixed. Then, each sample was divided into four equal parts, Figure 4-17, in order to obtain a representative sample, only 1/4 was used for freeze drying. Samples were kept in labelled plastic containers; however, the solid lids were replaced with punctured lids for when the samples were moved into the lyophilizer apparatus (Naoroz, 2018).
Coarse comminution by ring-milling

After being freeze dried, particle sizes was reduced using high abrasion milling instrument (Thompson and Nathanail, 2003). For quantitative and semi quantitative analysis such as XRD and XRF, samples must be reduced to a very fine mass (Thompson and Nathanail, 2003). Ring-milling entails the reduction of sediments by high energy collision and abrasion between the walls of a milling vessel and its inner loose elements. Excessive percussion processing must be avoided, in order to preserve the mineral crystalline structure (Thompson and Nathanail, 2003). The milling vessel for this study was made of pure agate with a stainless-steel outer casing. An agate ring and puck was inserted in the vessel for crushing purposes. As a large amount of sample will risk inefficient crushing, a small amount of sample was processed at a time and loaded in between the ring and the puck (Naidoo, 2018).

Before crushing the samples, the agate vessel was loaded with quartzite, and milled for 10 minutes. Twelve samples of ~15 g per sample were milled using this exercise and stored in Ziploc bags. A small amount of sample was added to the vessel before it was placed on the mill and oscillated for 10 min at 18 rpm. The agate vessel was run on low velocity as the shaking tends to generate heat and the agate may be damaged by the abrasion (Naidoo, 2018). To monitor the progression, samples were checked every 5 minutes. Between each sample, the mill was cleaned with ethanol to avoid cross contamination.

4.6.3 Fine comminution by micronizing mill

Samples prepared for XRD was milled to finer particles of <0.5 mm, using a McCrone micronizing mill. Ten samples of 3 g were prepared and added to a container containing 48 cylindrical agate percussion elements. The micronizing mill functions as a shaker, and the abrasion of the agate elements will reduce the particle size. ~10 mL of ethanol was added to the container to ensure an ample aqueous mix. The samples were milled for ~10 min each. After milling, the liquid samples were poured into a new plastic container and placed in an oven overnight at ~50°C for the liquid to evaporate from the sample.

4.6.4 Heat treatment

Loss on ignition (LOI) was performed on six samples from well 1 (North), three samples from well 2 (East), and one sample from well 3 (South). Experiment was performed in collaboration with (Kristiansen, 2019). This exercise was performed in conjunction with XRD and XRF sample preparation.
Soil sample was sequentially introduced to different high temperatures. By introducing the sample to high temperatures, volatile substances may escape. By measuring the changes in weight, we are measuring the loss of water, loss of organic content and loss of volatiles through the different heating steps. The elements that is lost during the fusing method cannot be otherwise measured by the XRF spectrometer, and it is therefore necessary to validate this through the LOI process beforehand (Naidoo, 2018).

Samples were carefully measured using an analytical balance (2 g) and added to small porcelain crucibles (Figure 4-18:Right). Samples were then loaded into an oven at 105°C overnight. This procedure will yield the loss of water using equation:

\[
\text{Loss of Water} \% = \frac{\text{Weight}_{\text{ini}} - \text{Weight}_{105^\circ}}{\text{Weight}_{\text{ini}}} \times 100 \tag{Eq. 4-12}
\]

Next, samples were placed on a fire safe tray to be inserted into a bench mounted chamber furnace, CWF1300 Carbolite at 550°C for 4 hours. After 4 hours, samples were taken out of the oven to be measured by weight to obtain the loss of organic material:

\[
\text{Loss of Organic Matter} \% = \frac{\text{Weight}_{105^\circ} - \text{Weight}_{550^\circ}}{\text{Weight}_{105^\circ}} \times 100 \tag{Eq. 4-13}
\]

When the furnace reached 1100°C, samples were placed back inside the oven for 2 more hours. Samples were then removed from the oven to cool before further processing. This heating process will help obtain the loss on ignition.

\[
\text{Loss on Ignition} \% = \frac{\text{Weight}_{550^\circ} - \text{Weight}_{1100^\circ}}{\text{Weight}_{1100^\circ}} \times 100 \tag{Eq. 4-14}
\]
4.6.5 Fusion beads

The fusion bead technique creates an accurate solution used in the analytical XRF process (Willis, 2010). Creating fusion beads, a non-crystallized amorphous homogenous solid is formed in form of a glass bead, which is ideal for XRF (Willis, 2010). Samples were carefully measured and mixed with a binding additive, Fluxana HD Elektronik FX-X65-2, that contains Lithium tetraborate 66.5% / Lithium metaborate 33.5%, to produce fused glass beads, Figure 4-19 show is a principle sketch of steps in the preparation process. Samples were prepared in the laboratory at the University of Oslo.

Using an analytical balance, samples were measured in a ratio of 6 g Flux and 0.6 g sample mixed together in a platinum crucible. An error value of ±0.0001 g was used for precision when mixing the samples. The platinum crucible was carefully inserted into an Eagon 2 furnace fusion system from PANalytical wearing heat resistant gloves. A 27 mm diameter platinum mold was also inserted. Before initiating the instrument, one tablet of Ammonium Iodide (NH4I) was inserted to avoid the glass bead from sticking to the mold. Sample mix were heated to a liquid phase as the instrument held a temperature of 1200˚C after fusion cycle was initiated. After the fusion process, the instrument automatically poured the liquid sample into the platinum mold to produce the bead. After the instrument cooled down, the bead was removed from the instrument, and placed on a sheet of paper for further cooling. A total of eight glass beads were produced and stored in labelled Ziploc bags. Crucibles were rinsed in a high frequency ultrasonic instrument for 5 minutes in 5% citric acid at a temperature of 70˚C.
between every sample to avoid cross contamination. Platinum crucibles were left in an oven at 110˚ to dry after being rinsed.

![Figure 4-19: Principle steps for sample preparation in creating fusion beads for XRF (Willis, 2010).](image)

4.6.6 Filter-peel method (Clay)

The filter peel method is useful in that clay particles are caught on a filter by a vacuum pump, and transferred onto a glass slide where it is peeled off of the filter leaving the clay residue on the glass slide (Hillier, 2003). The thickness of the sample is essential for accurately absorbing the x-ray beam to capture the angular range and the identifiable peaks that are detected (Hillier, 2003).

An agate mortar was used to crush samples into fine powder. Two samples of ~10 g was inserted into individual glass beakers of 600 mL. The beakers were then filled with 200 mL of Na2CO3. To initiate sediments in suspension, the sample was stirred well, and placed into a VWR ultrasonic cleaner USC-T for 10 min. The beakers containing the sample was then left to rest for exactly 24 hours under a lid. After 24 hours, 400 mL of water was added to the beakers. Again, sample was stirred before placed into the ultrasonic cleaner for another 10 min.

![Figure 4-20: Filtrating the clay sample onto a filter using a filtrating device.](image)
After this procedure the samples were left to rest for exactly 6 hours under a lid. Sediment must remain in suspension in order to filtrate the sample onto a filter.

The filtrating of the samples constituted a filter-peel method: Oriented Aggregate mount for clay mineral identification (Poppe et al.). A 45µm pore size filter of 47mm diameter was placed on the filtering device. Approximately 175 mL sample was poured into a glass funnel that was clamped to the filtrating device (Figure 4-20). A vacuum pump was initiated to draw the liquid sample onto the filter. When ample amount of sample was drawn onto the filter creating a clay film, the filter was removed from the filtrating device. The clay film was then lain on top, face down, on a glass slide. A small steel cylinder was used to gently roll over the clay film to transfer the clay over to the glass slide. After mounting sample on a glass slide, the sample needed to go through subsequent treatments and measurements.

**Bulk mineral identification**

Bulk sample preparation is a good method for the identification and quantification of the bulk minerals in a sample (Shen et al., 2012). After micro milling samples to fine material, samples were lightly powdered into even finer materials of 10 ~ 20 µm using an agate mortar. Samples were loaded onto individual plastic sample holders with thickness of 5 mm and a shallow mold space of ~2 cm diameter. The powder was gently pressed into the mold using a glass slide. To minimize preferential orientation of crystal structure, surface of the sample was made slightly rough, however, at approximately the same level as the mold.

### 4.7 X-ray fluorescence (XRF) and X-ray diffraction (XRD)

#### 4.7.1 X-Ray fluorescence (XRF)

XRF is a fast and accurate analytical method to determine the chemical composition of different material. The detection limit of the elements is based on the type of spectrometer system used. Wavelength dispersive system (WDXRF) and was chosen for this study because of its wide elemental range (Beryllium to Uranium) compared to its counter system: the energy dispersive system (EDXRF) with an elemental range from Sodium to Uranium. Notably, elements possessing a high atomic number will generally have better detection limit.

Elements have a characteristic fluorescent x-ray radiation which is emitted once the sample is irradiated by an x-ray source. Elements are color coded, and each discrete energy has a corresponding color. Measuring the intensity of energy (color) one may determine the quantity of an element is found in a sample, this process is denoted the quantitative analysis.
X-rays are electromagnetic waves ranging between 0.01-10nm on the electromagnetic spectrum. Its energy equivalent ranges 0.125-125 keV. The wavelength of x-rays is inversely proportional to its energy following equation:

\[ E \lambda = h c \]  

Eq 4-15

Where, \( E \) is the energy (keV), \( \lambda \) is the wavelength (nm), \( hc \) is the product of Planck’s constant and the velocity of light.

In this study, a PANalytical Axios Max Minerals with 4kW Rh-tube was used for XRF analysis. Ten samples were processed as recognized by Table 9, and handled by Thanusha Naidoo at the University of Oslo. X-Ray Fluorescence (XRF) spectrometer utilizes several crystals to detect the incidence of a series of elements by scanning the sample multiple times (Bouwer, 2010). Fusion beads was used to determine XRF outcome: Elemental composition and the chemistry of the sample (Naidoo, 2018). A Super Q Software was used to interpreting the raw data.

4.7.2 X-Ray powder diffraction (XRD)

Samples processed by the Mccrone Micromill was used to prepare samples for XRD measurements. Two Different measurements was done: Clay and bulk analysis.

XRD is a semi quantitative method of measuring mineral abundance and structure in a sediment sample (Naidoo, 2018). Results are obtained by analyzing the diffractive patterns and comparing these with patterns from a reference database (Ermrich and Opper, 2011). Characteristic x-rays are produced once the electrons receive enough energy to release the inner shell from the anode. Here, the resulting spectra K\( \alpha \) and K\( \beta \) are the most common different peaks of intensity obtained (Dutrow and Clark).

X-ray diffractometers comprise of three essential elements: x-ray tube, sample holder and an x-ray detector. X rays are emitted through the x ray tube by the heating of a filament (cathode) to produce electrons. Electrons are then generated between the cathode and a metal anode by introducing high voltage (Ermrich and Opper, 2011, Dutrow and Clark). Diffracted beams are related to the planar spacing in the crystalline powder. Briggs Law explains the relationship between diffraction and the x-ray beam on a crystal:
\[ n\lambda = 2d \sin \theta \]  

Eq 4-16

Where \( n \) is an integer, \( \lambda \) is x ray wavelengths, \( d \) is the planar spacing and \( \theta \) is the diffractive angle. The different reflective intesities are determined by the symmetry of the crystal, positions of atom and number of electrons of the atoms (Ermrich and Opper, 2011)

In total, twelve samples were processed: Ten for the bulk analysis, and two for clay analysis (Table 9). In between measuring’s, several heating processes were performed. Firstly, sample were placed onto a sample tower and inserted into the XRD Instrument, Bruker D8 Advance with Lynxeye detector and CuKa radiation (\( \lambda = 0.154 \) nm) operated at 40mA and 40kV. After measuring the samples, they were taken out of the instrument and treated with Ethylene Glycol (EG) and left in the oven for 60˚C overnight. Samples were later placed in the XRD instrument after glycolation and measured again. Samples were again placed back into the Lynxeye detector for another measuring. Next step constituted the insertion of the samples into a furnace at 350˚C for 1 hour. Sample was measured again, before finally inserting it back into the furnace at 500˚C for 1 hour. Sample was measured again after final step.

Profex and Diffrac.Eva were software programs utilized to aid the identifying of minerals. In Profex, a convergence progress and goodness of fit is produced to refine the analysis to an acceptable limit. Refinement process is conveyed as Rietveld weighted profile (Rwp), expected R value (Rexp), chi squared (\( x^2 \)) and goodness of fit (GOF) (Toby, 2006).

4.8 Torshovdalen water budget estimation

4.8.1 Modified Thornthwaite and Mather monthly water balance method

Water balance (or budget) requires an initial assessment of available water resources for any type of land use: Development, agriculture etc. (Pascual-Ferrer and Candela, 2015). The Thornthwaite and Mather Water balance method have been adopted, modified and further developed by Scientists for use of estimating monthly and annual water budget (XU and Singh, 1998). Available climatological data and average daylight hours was used to compute a monthly and annual water balance (Pascual-Ferrer and Candela, 2015). Data has been collected from the Norwegian meteorological Institute (MET), Blindern observation site (59.94295°N 10.72065°E), and calculation assumes that data values are uniform for the region. This model can be estimated using steady state values or continuous data such as water input like
precipitation, and output data like evapotranspiration, runoff etc. (Dingman, 2015). In this study, a modified Thornthwaite water balance will be estimated.

**Field capacity**

The field capacity (FC) is the amount of water that is held by suction in the pore space between sand grains, after the water has been drained from the soil by gravitational forces. FC is a dynamic value and as it encompasses several processes such as evaporation, transpiration, rainfall etc. (Kirkham, 2014). For steady state applications, a single parameter for soil-water storage capacity ($\text{Soil}_{\text{max}}$) of the area is necessary and is defined as:

$$\text{SOIL}_{\text{max}} = FC \times Z_{rz}$$  \hspace{1cm} \text{Eq 4-17}

Where; FC is the field capacity and $Z_{rz}$ is the vertical range of the root zone. However, this value is commonly between 100-150 mm. (Dingman, 2015). In this study, a $\text{Soil}_{\text{max}}$ value of 100 mm is used to accompany the water balance estimation.
**Potential evapotranspiration (PET) and reference evapotranspiration (RET).**

Calculating PET requires information about the average monthly temperatures and sunlight hours according to latitude location (Xu and Singh, 2001, Pascual-Ferrer and Candela, 2015). Firstly, the monthly heat index \( i \) needs to calculate to establish the annual heat index \( I \). The annual heat index is estimated based on the sum of each monthly heat index \( i \):

\[
i = \left( \frac{T}{5} \right)^{1.514} \quad \text{Eq 4-18}
\]

And

\[
I = \sum_{i=1}^{12} i \quad \text{Eq 4-19}
\]

Where \( T \) is average monthly temperature and \( I = 0 \) when \( T < 0 \)°C. Calculating initial PET, using previous parameters in following equation:

\[
PET_{\text{notcorrected}} = 16 \times (\frac{10^T}{I})^a \text{ (mm)} \quad \text{Eq 4-20}
\]

Where, \( a = 6,756 \times 10^{-7} \times I^3 - 7,71 \times 10^{-5} \times I^2 + 1,79 \times 10^{-2} \times I + 0,49239 \)

Calculating initial PET will yield a PET value that is not corrected for hours of sunlight \( d \) and number of days \( N \) in a month. Following equation will correct for these factors:

\[
PET_{\text{corrected}} = PET_{\text{notcorrected}} \times \frac{d}{12} \times \frac{N}{30} \text{ (mm)} \quad \text{Eq 4-21}
\]

Where potential evapotranspiration (PET) is in (mm), \( d \) is the average monthly sunlight in (hrs), and \( N \) is the number of days in a specific month (1-31) (Pascual-Ferrer and Candela, 2015, Xu and Singh, 2001).

Reference evapotranspiration is calculated based on PET values and is calculating for specific condition in contrast to PET that assumes that all available water will evaporate. RET occurs under favorable conditions and may also be denoted actual evapotranspiration. (Pascual-Ferrer and Candela, 2015, Fish, 2011, Shonsey, 2009).

\[
\text{if } P \geq PET, \text{ then } RET = PET, \quad \text{Eq 4-22}
\]
if $P \leq PET$, then $RET = P + \text{part of soil storage} \quad Eq \ 4-23$

Water input, snow accumulation and melt factor

Precipitation can occur in two forms of a liquid and a solid phase: Snow and Rain. As snow is critical in many regions, the amount of snow and snowmelt will greatly affect the water balance. According to the Thornthwaite water balance method cited in Dingman (2015), following equations determine the type of precipitation:

$$Rain = F \cdot P \quad Eq \ 4-24$$

$$Snow = (1-F) \cdot P \quad Eq \ 4-25$$

Where $F$ is the melt factor in (mm) and $P$ is precipitation in (mm). $F$ is based on following conditions:

$$F=0, \ if \ T \leq 0^\circ C; \ F=0.167*T, \ if \ 0^\circ C \leq T \leq 6^\circ C; \ F=1, \ if \ T \geq 6^\circ C \quad Eq \ 4-26$$

Snowmelt provide a substantial source of water into the system. Snow becomes storage of water at certain temperature until melting of the snowpack occur (Healy et al., 2007). The melt factor (Melt) is calculated as:

$$Melt = F \cdot (\text{Snow Pack}_{m-1} + \text{Snow Pack}_m) \quad Eq \ 4-27$$

Where $\text{Snow Pack}_{m-1}$ is the monthly snow pack from the previous month, and $\text{Snow Pack}_m$ is the sum of snowpack at the end of each month.

$$\text{Snow Pack}_m = (1-F)^2 \cdot P + (1-F) \cdot \text{Snow Pack}_{m-1} \quad Eq \ 4-28$$
Snowpack equivalent (SWE) was estimated by using a formula developed by (Jonas et al., 2009). Where SWE is the snowpack water equivalent in kg/m², Hs is the snow depth in m and Pb is the bulk snow density in kg/m³.

\[ SWE = Hs \times Pb \quad Eq \ 4-29 \]

\[ Pb = 60.1 \times Hs^{0.89} + 237 \quad Eq \ 4-30 \]

\[ SWE = (60.1 \times Hs^{0.89} + 237) \times Hs \quad Eq \ 4-31 \]

**Soil moisture storage and soil storage**

Soil moisture storage is the amount of soil moisture in a system at a given time. Soil moisture storage (SM) is calculated based on several parameters such as P, PET, Soil\(_{\text{max}}\) and water input. Firstly, water input needs to be determined:

\[ W_{in} = P + \text{Melt} \quad Eq \ 4-32 \]

Where \( W_{in} \) is the water input (mm), P is the precipitation (mm), and Melt is the snowmelt (mm). If there is an increase or decrease in soil moisture storage, following conditions will determine the SM:

\[ \text{If } W_{in} \geq PET_{\text{corrected}}, \text{ then } SM = W_{in} - PET + \text{Soil Moisture}_{m-1} \quad Eq \ 4-33 \]

when \( \text{Soil}_{\text{max}} \) is reached.

\[ \text{If } W_{in} \leq PET_{\text{corrected}}, \text{ then } SM = \text{Soil Moisture}_{m-1} \times e^{(\text{-PET}-W_{in})/\text{Soil}_{\text{max}}} \quad Eq \ 4-34 \]

Soil storage (SS) occurs once the PET and RET are equal and the field capacity is not exceeded (P-PET<FC), when soil storage exceeds the field capacity, surplus of water will occur (P>PET+FC). Soil storage accumulates from month to month (Pascual-Ferrer and Candela, 2015).

\[ \text{If } PET=RET, \text{ then } SS = P-RET \quad Eq \ 4-35 \]
**Water deficit and surplus**

Water deficit occurs when the amount of evapotranspiration exceeds the value of precipitation, and as a result the difference between potential evapotranspiration and actual evapotranspiration equals the water deficit of the soil (Pascual-Ferrer and Candela, 2015) (McCabe and Markstrom, 2007).

Water surplus occurs when precipitation is much larger than potential evapotranspiration, and the water starts to accumulate in the soil. If water storage exceeds the value of Soil\textsubscript{max}, and P > PET, then water surplus = P - (PET+SM) (Pascual-Ferrer and Candela, 2015) (McCabe and Markstrom, 2007). Runoff can be further calculated from the availability of water surplus (Dingman, 2015).

**Surface water runoff (RO)**

Former soil conservation system (SCS), now natural Resources Conservation Service have developed procedures for estimating runoff and peak discharge. To estimate runoff, this method apply the runoff curve number (CN) approach (NRCS, 1986).

\[
Q = \frac{(P - I_a)^2}{(P - I_a) + S} \quad \text{Eq 4-36}
\]

Where Q is runoff (mm), P is precipitation (mm), S is the potential maximum retention after runoff begins (mm), and I\textsubscript{a} is the total losses before water surplus (mm). I\textsubscript{a} correlates to soil and cover parameter and is therefore very variable. According to (NRCS, 1986), I\textsubscript{a} can be approximated by the following equation:

\[
I_a = 0.2S \quad \text{Eq 4-37}
\]

Replacing Eq 4-37 in Eq 4-36, this will yield an estimate that will produce a specific runoff amount, by combining S and P parameters:

\[
Q = \frac{(P - 0.2S)^2}{(P + 0.85)} \quad \text{Eq 4-38}
\]
The curve number (CN) is a unique table of soil and cover conditions which S is related. Constants for this equation varies from use of different units, here S in mm (Dingman, 2015). S can be found by using the following equation:

\[ S = \frac{25400}{CN} - 254 \]  \hspace{1cm} Eq 4-39

NRCS method uses a table of land cover and hydrological conditions of soil types to represent a curve number. The Soil is divided into four different hydrologic soil groups (HSG). Here, several diagrams have been developed for different land use and infiltration rate (NRCS, 1986). However, urbanization may have altered the soil profile substantially and normal determination of the CN number do not apply. See appendix G for further information about the HSG. This report will focus on the use of the runoff curve for urban areas

**Groundwater flow**

Groundwater flow in the unconsolidated aquifer was estimated using a simple modifications of darcy's law for calculations through a cross-sectional area:

\[ Q = -KiA \]  \hspace{1cm} Eq 4-40

Where Q is the flow rate (L\(^3\)/t), K is the hydraulic conductivity (L/t), A is the cross-sectional Area (L\(^2\)). The cubic law was formulated by modifying Darcy’s law and make it applicable to calculate flow rate through fracture rock (Hiscock and Bense, 2014). In certain fractured rock aquifers assumptions on flow and aperture under laminar flow conditions have been made to simplify the use of Darcy’s law. Flow and hydraulic gradient is governed by the cubic law as follows (Witherspoon et al., 1980):

\[ Q_f = -2bwK \frac{dh}{dl} \]  \hspace{1cm} Eq 4-41

Where b is the aperture of the fracture (L), w is the fracture spacing (L), K\(_f\) is the hydraulic conductivity of the fracture (L/t) and dh/dl is the hydraulic gradient (-). Hydraulic conductivity can be calculated as follows:
\[ K_f = \frac{\rho g (2b)^2}{12\mu} \]

Eq 4-42

Where \( \rho \) is fluid density (1000 kg/m\(^3\)), \( \mu \) is fluid viscosity (8.9 kg/m\(^*\)s), and \( g \) is the gravitational acceleration (9.8 m/s\(^2\)).
5 Results

Results comprise findings from collected data as well as interpretation of data provided by the Norwegian Geotechnical Institute (NGI).

5.1 Lithology

Three wells well 1 (North), well 2 (East) and well 3 (South) was constructed for groundwater monitoring, in addition, two boreholes well location 1 (North) and well location 3 (South) was constructed for geophysical logging. Lithology logs, sample location and well configuration is graphically presented in Figure 5-1. Well 1 (North) (6651807N, 264294E) was drilled with a Sonic rig using a 102 mm drill bit. The requested drilling depth was ~15 m. As illustrated by Figure 5-1 well 1 (North) constitutes ~3 m topsoil containing organic matter, dominant primary minerals such as quartz and feldspar and clay minerals such as illite and chlorite. Evidence of shale and calcareous layers were found between 3 - 12 m. The well was configured with 3 m solid casing in the unconsolidated layer and 10 m perforated casing in the fractured shale. The well was finalized by NGI. Cuttings from drilling was collected from well 1 (North) (B1_F4) and well 2 (East) (B2_F1 and B2_F4). The most surprising finding was sample B1_F4 as illustrated in Figure 5-1. Sample contained various angular clasts of granitic origin among dark shale debris.
Well 2 (East) (6651611N, 264025E) and Well 3 (South) (6651428N, 263936E), were constructed using a 76 mm drill bit. However, at Well 2 (East), a larger casing of 115 mm diameter was used to finalize the borehole. As seen in Figure 5-1, depth to bedrock at well 2 (South) was ~7.46 m and ~17.57 m for well 3 (South). Borehole stratigraphy of well 2 (East) constitutes of a ~2 m soil layer of organic matter and silty loam with inclusions of anthropogenic waste such as building bricks and a ~1 m layer of silty clay loam. Quartz and feldspar are dominant primary minerals found in the soil. Shale was found from 3 m. Well was configured with 3.11 m solid casing in the unconsolidated layer and 3 m perforated casing in the fracture shale. Borehole stratigraphy for well 3 (South) constitutes ~12 m of alternating thick layers of silty clay and thinner layers of silty loam. Shale with inclusions of calcareous layers were found between 12.3 - 17.57 m. The well was constructed with a ~12.75 m solid casing in the unconsolidated layer and 2 m of open casing in the fracture shale.
### Figure 5-1: Lithology logs, sample location and well configuration from well 1 (North), well 2 (East) and well 3 (South)

<table>
<thead>
<tr>
<th>Depth (m)</th>
<th>Stratigraphy</th>
<th>Well Configuration</th>
<th>Sample:</th>
<th>Depth (m)</th>
<th>Stratigraphy</th>
<th>Well Configuration</th>
<th>Sample:</th>
<th>Depth (m)</th>
<th>Stratigraphy</th>
<th>Well Configuration</th>
<th>Sample:</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>Soil</td>
<td>Solid Casing</td>
<td>B1-L1.5</td>
<td>Soil</td>
<td>Solid Casing</td>
<td>B2-L1</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>0-3 m</td>
<td>B1-L2.5</td>
<td>Silty loam</td>
<td>0-3.11 m</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>B1-L3.5</td>
<td>Silty clay</td>
<td>6.11-7.46 m</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>2</td>
<td>Shale</td>
<td></td>
<td>B1-F1</td>
<td>Shale</td>
<td>Open Casing</td>
<td>B2-F1</td>
<td></td>
<td></td>
<td></td>
<td></td>
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</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>B1-F1.5</td>
<td></td>
<td>3.11-6.11 m</td>
<td>B2-F4</td>
<td></td>
<td></td>
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<tr>
<td>3</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>Plugged</td>
<td></td>
<td></td>
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<td>4</td>
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<tr>
<td>12</td>
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<td>18</td>
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<td></td>
<td></td>
<td></td>
<td></td>
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</tr>
</tbody>
</table>

**Well 1 (North)**
- Elevation: 102.22 m
  - (6651807N, 264294E)

**Well 2 (East)**
- Elevation: 73.4 m
  - (6651611N, 264025E)

**Well 3 (South)**
- Elevation: 60.95 m
  - (6651428N, 263936E)
5.2 Geophysical properties

Geophysical logging was carried out by NGI at well location 1 (North) and 3 (South) by means of: Natural gamma (GR), resistivity, and optical and acoustic tele viewer (NGI, 2019) and logs are represented from well 1 (North) in Figure 5-2 and from well 3 (South) in Figure 5-3. As seen in Figure 5-2, here, GR show a significant jump from 40 - 80 CPS marking the shale and unconsolidated sediment boundary. Bedrock is thus confirmed at ~13 m depth. Resistivity represent the same jump from 24 - 213 ohm. Figure also illustrates characteristic caliper and amplitude log results in form of a blue sinusoidal wave. Here, the results demonstrate a fracture at 18.5 m with a strike and dip of 260/62 SW with an aperture of ~0.12 m. Surprisingly, neither GR nor resistivity implies an obvious change in the lithology where fracture occurs, however, an apparent sinusoidal wave is detected in the acoustic amplitude and caliper log, illustrating the change in density in the shale. Well 3 (South) show one clear fracture compared to well 1 (North). Concerning well 1 (North), the acoustic and amplitude log show an anomaly in the shale layering between 7-9 m in well 1 (Figure 5-3) indicating alternating shale (120 CPS) and possibly sandy shale (19-54 CPS). Peaks seen in the resistivity log strengthen these evidences.
Figure 5-2: Geophysical logs from well 3 (South). From left: Optical tele viewer, acoustic-caliper and amplitude, resistivity and natural gamma (GR).

Figure 5-3: Geophysical logs from well 1 (North). From left: Optical tele viewer, acoustic-caliper and amplitude, resistivity and natural gamma (GR).
5.3 Geomechanical properties of soil and sediment

Several laboratory tests were performed on three undisturbed cylinder samples extracted from well 3 (South). Table 10 gives an overview of samples collected and initial properties of the samples. The following experiments were performed by NGI.

Table 10: Initial properties: volume (V). Density (ρ). Length (L) and specific weight (γ) from samples 1, 2 and 3 near well 3 (South).

<table>
<thead>
<tr>
<th>Location</th>
<th>Cylinder</th>
<th>Depth [m]</th>
<th>Length [cm]</th>
<th>Volume [cm³]</th>
<th>Density [Mg/m³]</th>
<th>Specific Weight [kN/m³]</th>
</tr>
</thead>
<tbody>
<tr>
<td>Well 3 (South)</td>
<td>1</td>
<td>4.10</td>
<td>21.0</td>
<td>855.1</td>
<td>3.34</td>
<td>32.8</td>
</tr>
<tr>
<td>Well 3 (South)</td>
<td>2</td>
<td>5.35</td>
<td>60.0</td>
<td>2443.2</td>
<td>1.90</td>
<td>18.6</td>
</tr>
<tr>
<td>Well 3 (South)</td>
<td>3</td>
<td>6.40</td>
<td>73.5</td>
<td>2992.9</td>
<td>1.91</td>
<td>18.7</td>
</tr>
</tbody>
</table>

5.3.1 Shear strength

Cylinder 2 (5.37 m depth) and cylinder 3 (6.33 m depth) was tested for shear. Comparing both cylinders, cumulative results show that the hardness of clay and ability to withstand pressure is increasing with depth. Shear strength is measured in kilo-Pascal (kPa) and is highlighted in grey in Table 11 below. As can be seen in Table 11, some of the main characteristics from cylinder 2 (5.37 m depth) was the low shear strength (C_u) of 23 kPa and axial pressure (qu) of 46 kPa. Cylinder 2 classifies as a soft clay. Cylinder 3 (6.33 m depth) display a much higher shear strength (C_u) value of 36 kPa and axial pressure (qu) of 72 kPa. Cylinder 3 classifies as a medium hard clay. Cylinder 2 and cylinder 3 is graphically presented in Figure 5-4 where failure point occur after pressure is applied. Red arrows indicate where the axial clay sample collapses with respect to axial strain (kPa) and axial stress (%). Axial strain from cylinder 2 collapses at 8.9 %, whereas cylinder 3 collapses at 3.9 %. See appendix A for classification and standards.
Table 11: Results from NGI illustrating compiled results from the axial compression test. Shear strength (Cu), axial strain (ε), axial pressure strength (qu), water content (w), specific weight (γ), density (ρ), dry density (ρd), axial strength rate (ε-rate). Modified from (NGI, 2019).

<table>
<thead>
<tr>
<th>#</th>
<th>D</th>
<th>Cu</th>
<th>ε</th>
<th>qu</th>
<th>w</th>
<th>γ</th>
<th>ρ</th>
<th>ρd</th>
<th>ε-rate</th>
<th>L</th>
<th>A</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>m</td>
<td>kPa</td>
<td>%</td>
<td>kPa</td>
<td>%</td>
<td>kN/m3</td>
<td>Mg/m3</td>
<td>Mg/m3</td>
<td>%/mm</td>
<td>mm</td>
<td>cm2</td>
</tr>
<tr>
<td>2</td>
<td>5.37</td>
<td>23</td>
<td>8.9</td>
<td>46</td>
<td>35.2</td>
<td>18.9</td>
<td>1.93</td>
<td>1.41</td>
<td>2.7</td>
<td>138</td>
<td>40.7</td>
</tr>
<tr>
<td>3</td>
<td>6.33</td>
<td>36</td>
<td>3.9</td>
<td>72</td>
<td>35.4</td>
<td>18.7</td>
<td>1.91</td>
<td>1.41</td>
<td>2.7</td>
<td>138</td>
<td>40.7</td>
</tr>
</tbody>
</table>

Figure 5-4: Cylinder 2 and 3 are shown as shear strength (cu) axial strain (ε) and is graphically illustrates when failure point occurs. Modified from NGI (2019)

5.3.2 Undisturbed and remolded sample

Evaluation of the fall cone undrained shear strength was performed on undisturbed soil (Cufc) and remolded soil (Curfc) from cylinder 1, 2 and 3 from well location 3 (South). Figure 5-5 exemplifies the difference between undisturbed and remolded sample behavior. Cone penetration test show soil strength of 19-24 kPa for undisturbed, and 2.9-4.3 kPa for remolded sample. As seen in Figure 5-5, remolded sample demonstrate a slight decrease in stability after liquefication. Here, cone penetrates further (7.4-9 mm) than that of the undisturbed sample (6.2-8 mm). Interestingly, the most striking result was seen in Figure 5-5 at depth of 5.12 m. Contrary to other samples, for undisturbed soil sample, cone penetrates further (8 mm) compared remolded sample (7.5 mm). Sensitivity ranges from 4–8 kPa. Classification of sensitivity can be seen in the appendix A which is modified from (Vegdirektoratet, 2016).
5.4 Soil humus- and water content

Qualitative evaluation of humus in the soil was determined by the loss on ignition (LOI) on three samples from well location 3 (3A, 3B and B3-L4), nine samples from well location 1 (1A, 1B, B1-Ln and B1-Fn) and four from well location 2 (2A-C, B2-Ln and B2-Fn). In addition, each test has been evaluated for its water content in weight percent water per sample and presented in Table 12 and Table 13.

Humic substance from undisturbed cylinder samples are shown as an average of three trials in Table 12, while disturbed sediment samples are shown in Table 13. The latter contains calculation from three heating treatment steps on sediment samples. The table includes following parameters: Loss of water (105°C), loss of organic material (550°C) and loss of volatiles (1100°C), all values given as percentage. Percentages are calculated with respect to sample weight of ~2 g each. Loss of Organic content varies between 1-3 %, not including the outlier. Loss of volatiles ranges from 1-17 % between the different sample locations and sample depth. As seen in Table 13, sample B2-F4 at depth 4-5 m, represents a significant outlier compared to the other samples.

Humic levels and weight percent water is decreasing with depth demonstrated in all cylinder samples. It can be seen from the data in Table 12: Humic soil show 5.2 % in cylinder 1 A-B between 4.09- 4.16 m depth which is much higher than cylinder 2A-C and 3A-B with 2.4 % humic content at depth 5.09-6.42 m. Table also present an overview of weight percent water found in sample. Weight percent water is greater in cylinder 1B with 51% at depth 4.16 m.
Sample 2A, 2C and 3A have a unison water content of 33, 35, 35, 34% at depth 5.09-6.67 m. See appendix B for detailed calculations.

Table 12: Descriptions of each undisturbed sample as extracted from cylinder from well 2 (South). Humic content is displayed in percent average from organic content. Water content measured in percent with respect to weight difference between wet to dry sample. Modified from (NGI, 2019).

<table>
<thead>
<tr>
<th>Sample nr:</th>
<th>Depth m</th>
<th>Water %</th>
<th>M500 %</th>
<th>Average Organic content %</th>
</tr>
</thead>
<tbody>
<tr>
<td>1A</td>
<td>4.09</td>
<td>25.7</td>
<td></td>
<td>5.2</td>
</tr>
<tr>
<td>1B</td>
<td>4.16</td>
<td>51.0</td>
<td></td>
<td></td>
</tr>
<tr>
<td>2A</td>
<td>5.09</td>
<td>33.3</td>
<td></td>
<td></td>
</tr>
<tr>
<td>2B</td>
<td>5.22</td>
<td>100.0</td>
<td>2.4</td>
<td></td>
</tr>
<tr>
<td>2C</td>
<td>5.66</td>
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<td></td>
</tr>
<tr>
<td>3A</td>
<td>6.11</td>
<td>35.7</td>
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</tr>
<tr>
<td>3B</td>
<td>6.42</td>
<td>100.0</td>
<td>2.4</td>
<td></td>
</tr>
</tbody>
</table>

Table 13: Table contains calculation from heat treatment on disturbed sediment samples from well 1, 2 and 3. Table incudes following parameters: Loss of water (105°C), loss of organic material (550°C), and loss of volatiles (1100°C). all values given in percentage.

<table>
<thead>
<tr>
<th>Sample nr:</th>
<th>Depth m</th>
<th>Water %</th>
<th>550°C %</th>
<th>1100°C %</th>
</tr>
</thead>
<tbody>
<tr>
<td>B1-L1.5</td>
<td>0-1.5</td>
<td>1</td>
<td>2</td>
<td>6</td>
</tr>
<tr>
<td>B1-L2.5</td>
<td>1.5-2.5</td>
<td>1</td>
<td>1</td>
<td>14</td>
</tr>
<tr>
<td>B1-L3.5</td>
<td>2.5-3.5</td>
<td>1</td>
<td>1</td>
<td>2</td>
</tr>
<tr>
<td>B1-F1</td>
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<td>2</td>
<td>9</td>
</tr>
<tr>
<td>B1-F1.5</td>
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<tr>
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<td>0.5</td>
<td>11</td>
<td>2</td>
</tr>
<tr>
<td>B3-L4</td>
<td>3-5</td>
<td>1.5</td>
<td>3</td>
<td>1</td>
</tr>
</tbody>
</table>

5.5 NaCl-content in sediment cores

NaCl content was analyzed from well 3 (South) cylinder samples 1, 2, and 3 at depth 4.16-6.43 m. Table 14 show NaCl equivalents in sample (g/kg), NaCl equivalents in
porewater (g/l) and conductivity in (µS/cm). Conductivity calculations is achieved with a sample-water relationship of 1:1 (NGI, 2019).

Results found clear support that conductivity and salinity is decreasing proportionally with depth from all cylinder samples. As can be seen in Table 14, upper cylinder (1) at 4.16 m depth show a conductivity of 684 µS/cm, and NaCl content in porewater 0.35 g/l and 3.45 g/kg NaCl in sample. Subsequently, the lower cylinder (2) and (3) show a decrease in NaCl content in porewater showed 0.11g/l and NaCl in sample was 1.60-1.73 g/kg and conductivity with 219 µS/cm. Figure 5-6 illustrates the linear relationship between Salinity concentration and conductivity. Conductivity is increasing as NaCl content increases (y=0.0038x+0.8243).

<table>
<thead>
<tr>
<th>Cylider</th>
<th>Weight (wet)</th>
<th>Volume Water</th>
<th>Conductivity</th>
<th>Conductivity as NaCl-equivalents</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>15.1</td>
<td>44.9</td>
<td>684</td>
<td>0.35 g NaCl/l porewater</td>
</tr>
<tr>
<td>(4.16m)</td>
<td></td>
<td></td>
<td></td>
<td>3.45 g NaCl/kg</td>
</tr>
<tr>
<td>2</td>
<td>13.5</td>
<td>46.28</td>
<td>219</td>
<td>0.11 g NaCl/l porewater</td>
</tr>
<tr>
<td>(5.26m)</td>
<td></td>
<td></td>
<td></td>
<td>1.60 g NaCl/kg</td>
</tr>
<tr>
<td>3</td>
<td>13.3</td>
<td>46.93</td>
<td>219</td>
<td>0.11 g NaCl/l porewater</td>
</tr>
<tr>
<td>(6.43m)</td>
<td></td>
<td></td>
<td></td>
<td>1.73 g NaCl/kg</td>
</tr>
</tbody>
</table>

** Volume water represents amount of water added to the sample to perform the experiment.

![Graph](image)

*Figure 5-6: NaCl concentration is evaluated against conductivity from well 3 (South) samples.*
5.6 Grain size analysis and soil texture classification

Grain size analysis was performed on three undisturbed samples from cylinders extracted from well 3 (South) at 4.22 (1A-B), 5.29 (2-A-C), and 6.47 (3A-B) meters depth. Results are graphically presented of Figure 5-7, Figure 5-8 and Figure 5-9. In addition, ten samples from well 2 (East) from 0.1 - 2.2 m depth (W2-n), and ten samples collected from various locations along three transect 1, 2 and 3 (Ln-T1, Ln-T2, and Ln-T4) are presented graphically in Figure 5-11, Figure 5-12, Figure 5-13, Figure 5-14. The organization of all samples are shown in Table 15. See appendix G (a-g) for supplementary information.

Table 15: Grainsize distribution sample directive.

<table>
<thead>
<tr>
<th></th>
<th>Well 2</th>
<th>Well 3</th>
<th>Transect 1</th>
<th>Transect 2</th>
<th>Transect 3</th>
</tr>
</thead>
<tbody>
<tr>
<td>B2-0.1 m</td>
<td>B3-0.5 m</td>
<td>L1-T1</td>
<td>L1-T2</td>
<td>L1-T3</td>
<td></td>
</tr>
<tr>
<td>B2-0.3 m</td>
<td>B3-1.0 m</td>
<td>L2-T1</td>
<td>L2-T2</td>
<td>L2-T3</td>
<td></td>
</tr>
<tr>
<td>B2-0.5 m</td>
<td>B3-2.5 m</td>
<td>L4-T1</td>
<td>L4-T2</td>
<td>L3-T3</td>
<td></td>
</tr>
<tr>
<td>B2-1.3 m</td>
<td>B3-3.0 m</td>
<td></td>
<td></td>
<td>L4-T3</td>
<td></td>
</tr>
<tr>
<td>B2-1.5 m</td>
<td>B3-3.5 m</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>B2-2.0 m</td>
<td>B3-4.0 m</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>B2-2.2 m</td>
<td>1A-B</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>2A-C</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>3A-B</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>
Figure 5-7 presents an overview of the grainsize distribution from cylinder 1 at 4.22 m depth which is dispersed as follows: 22% gravel, 20% sand, 50% silt and 8% clay. Subsequently, Figure 5-8 show cylinder 2 at 5.29 m depth and Figure 5-9 show cylinder 3 at 6.47 m depth. Both samples display a homogenous distribution of finer materials: Clay content of 44-47% and silt content of 52-56%. Grainsize distribution of finer content is increasing with depth. See appendix G. a-f for grainsize distribution data.

Planned comparison from sediment samples collected from field campaigns reveal variable percent content of gravel, sand, silt and clay in the upper 0.3–4 m. However, silt is the dominant grain size ranging from 28–68% among twenty-two samples that encompasses transect- and well samples. Combined, the fine-grained sediments are dominating with a percent content of 34-100% between respective samples.

**Cylinder 1**

<table>
<thead>
<tr>
<th>Material</th>
<th>%</th>
</tr>
</thead>
<tbody>
<tr>
<td>Boulders</td>
<td>0</td>
</tr>
<tr>
<td>Gravel</td>
<td>22</td>
</tr>
<tr>
<td>Sand</td>
<td>20</td>
</tr>
<tr>
<td>Silt</td>
<td>50</td>
</tr>
<tr>
<td>Clay</td>
<td>8</td>
</tr>
</tbody>
</table>

Figure 5-7: Grainsize distribution performed by NGI. Sample is collected from a cylinder from 4.22 m depth. Modified from (NGI, 2019).

**Cylinder 2**

<table>
<thead>
<tr>
<th>Material</th>
<th>%</th>
</tr>
</thead>
<tbody>
<tr>
<td>Boulders</td>
<td>0</td>
</tr>
<tr>
<td>Gravel</td>
<td>0</td>
</tr>
<tr>
<td>Sand</td>
<td>0.1</td>
</tr>
<tr>
<td>Silt</td>
<td>53</td>
</tr>
<tr>
<td>Clay</td>
<td>47</td>
</tr>
</tbody>
</table>

Figure 5-8: Grainsize distribution performed by NGI. Sample is collected from a cylinder from 5.29 m depth. Modified from (NGI, 2019).
Figure 5-9: Grainsize distribution performed by NGI. Sample are collected from a cylinder extracted from 6.47 m depth. Modified from (NGI, 2019).

Additional Grainsize distribution from well 2 (East) and well 3 (South) is presented graphically in Figure 5-10 and Figure 5-11. Cumulative volume is plotted against percent volume. Samples were extracted from 0.5-4 m, and 0.1-2.2 m from well 3 (South) and Well 2 (East) respectively, see Table 15. Mean percent material of samples shows a dispersion of grainsize: 30 % Gravel, 36 % Sand, 28 % Silt and 6 % Clay. Results from well 3 are preceding samples from cylinder 1 (Figure 5-7). Samples represented in the Figure 5-11 and Figure 5-11 show the grain size distribution in soil samples collected at well 2 (East). Average soil content display: 2 % gravel, 6 % sand, 68 % silt and 24 % clay. In general, results obtained from well 2 (East) display a noticeably higher clay and silt content of 24 % and 68 % in the soil compared to well 3 (South) with a fine particle content of 6 % and 28 %.

Grainsize distribution from three transects are presented graphically in Figure 5-12, Figure 5-13 and Figure 5-14 below. Collectively, the mean value between sediment textures are presented in conjunction with the grainsize distribution plot. Samples from each transect show a varying content of gravel, sand, silt and clay shown below. An interesting observation is the great deviations between samples along transect 3. Results from transect 3 demonstrate a variable dispersion in grainsizes. In general, silt is the dominant particle size (62-76 %), however, one sample differ (L3-T3), were sand is the dominant particle size (44%). This may also be seen in Figure 5-14.
Well 3 (South)

<table>
<thead>
<tr>
<th>Material</th>
<th>Avg. %</th>
</tr>
</thead>
<tbody>
<tr>
<td>Boulders</td>
<td>0</td>
</tr>
<tr>
<td>Gravel</td>
<td>30</td>
</tr>
<tr>
<td>Sand</td>
<td>36</td>
</tr>
<tr>
<td>Silt</td>
<td>28</td>
</tr>
<tr>
<td>Clay</td>
<td>6</td>
</tr>
</tbody>
</table>

Figure 5-10: Grain size distribution from well 3. Samples are preceding samples presented in.
Figure 5-11: Grain size distribution from Well 2 (East). Sample constitutes the upper two meters from ground level.

<table>
<thead>
<tr>
<th>Material</th>
<th>Avg.%</th>
</tr>
</thead>
<tbody>
<tr>
<td>Boulders</td>
<td>0</td>
</tr>
<tr>
<td>Gravel</td>
<td>2</td>
</tr>
<tr>
<td>Sand</td>
<td>6</td>
</tr>
<tr>
<td>Silt</td>
<td>68</td>
</tr>
<tr>
<td>Clay</td>
<td>24</td>
</tr>
</tbody>
</table>

Transect 1

<table>
<thead>
<tr>
<th>Material</th>
<th>Mean</th>
<th>St.dev</th>
</tr>
</thead>
<tbody>
<tr>
<td>Boulders</td>
<td>0</td>
<td>0</td>
</tr>
<tr>
<td>Gravel</td>
<td>7.3</td>
<td>4.9</td>
</tr>
<tr>
<td>Sand</td>
<td>9.6</td>
<td>3.6</td>
</tr>
<tr>
<td>Silt</td>
<td>67.1</td>
<td>5.4</td>
</tr>
<tr>
<td>Clay</td>
<td>16.0</td>
<td>2.5</td>
</tr>
</tbody>
</table>

Figure 5-12: Grain size distribution from Transect 1. Graph illustrates the analysis from three sample locations.

Transect 2

<table>
<thead>
<tr>
<th>Material</th>
<th>Mean</th>
<th>St.dev</th>
</tr>
</thead>
<tbody>
<tr>
<td>Boulders</td>
<td>0</td>
<td>0</td>
</tr>
<tr>
<td>Gravel</td>
<td>9.8</td>
<td>5.8</td>
</tr>
<tr>
<td>Sand</td>
<td>14.9</td>
<td>1.64</td>
</tr>
<tr>
<td>Silt</td>
<td>64.5</td>
<td>6.0</td>
</tr>
<tr>
<td>Clay</td>
<td>10.7</td>
<td>1.7</td>
</tr>
</tbody>
</table>

Figure 5-13: Grain size distribution from Transect 2. Graph illustrates the analysis from three sample locations.
Figure 5-14: Grain size distribution from Transect 3. Graph illustrates the analysis from four sample locations.

Table 16: Percent Material per location within one transect. Sediment samples are deviating greatly between each sample.

<table>
<thead>
<tr>
<th></th>
<th>L1-T3 (%)</th>
<th>L2-T3 (%)</th>
<th>L3-T3 (%)</th>
<th>L4-T3 (%)</th>
<th>Mean</th>
<th>St. Deviation</th>
</tr>
</thead>
<tbody>
<tr>
<td>Boulder</td>
<td>0</td>
<td>0</td>
<td>0</td>
<td>0</td>
<td>0.0</td>
<td>-</td>
</tr>
<tr>
<td>Gravel</td>
<td>8</td>
<td>3</td>
<td>21</td>
<td>3</td>
<td>9.0</td>
<td>8.6</td>
</tr>
<tr>
<td>Sand</td>
<td>16</td>
<td>7</td>
<td>44</td>
<td>8</td>
<td>18.7</td>
<td>17.2</td>
</tr>
<tr>
<td>Silt</td>
<td>62</td>
<td>71</td>
<td>29</td>
<td>76</td>
<td>59.5</td>
<td>21.0</td>
</tr>
<tr>
<td>Clay</td>
<td>14</td>
<td>19</td>
<td>6</td>
<td>13</td>
<td>12.8</td>
<td>5.5</td>
</tr>
</tbody>
</table>
5.6.1 Saturated hydraulic conductivity

Saturated hydraulic conductivity (k\text{\text{sat}}) was calculated based on grain size distribution data between transects 1, 2 and 3, well 2 (East) and well 3 (South). Comparison of k\text{\text{sat}} is revealed in Figure 5-16. Results from transect 1, 2 and 3 are presented with respect to location of sample, while well 2 and 3 are presented with respect to depth. Ranges of different hydraulic conductivity according to its material classification for unconsolidated sediments may be found in appendix G.e (NRCS, n.d-b). Collectively, values from all locations along transect 1, 2, 3 and 4 ranges from 0.007-0.02 cm/h using Hazen’s method which correspond to a clayey soil with slow to very slow hydraulic conductivity. Whilst Puckett 1985 displayed rates of 0.4-5.1 cm/h, which correspond to a loamy soil with a moderately slow hydraulic conductivity (NRCS,
n.d-b). The graphs presented in the figure is quite revealing as it demonstrates the great difference between the methods with increasing depth. Hydraulic conductivity calculations from samples collected with depth from well 2 display ranges from 0.001-0.002 cm/h using Hazen method, and 0.1-0.5 cm/h using Puckett 1985. Well 3 displayed ranges of \( k_{\text{sat}} \) values of 0.01-0.1 cm/h using Hazen and 3-7 cm/h using Puckett 1985. See appendix G (a-f) for supplementary information.

Figure 5-16: \( k_{\text{sat}} \) was calculated based on grainsize distribution and compared between empirical methods. Hazen’s method is marked with a red line, while Puckett 1985 is marked with a green line.
Sediment samples from three transects 1, 2 and 3, well 2 and well 3 were evaluated for risk of frozen ground. As seen from Figure 5-17, Samples are grouped based on the percentage of grain size <0.02 mm, <0.2 mm, and >40% 0.002 mm diameter. Frost Group was categorized from a range from T1-T4, whereas T1 has the least risk of frost and T4 has the highest risk. Risk of frozen soil is decreasing with depth. Well location 2 (East) and 3 (South) has 71-91% of grainsize <0.02 mm, and 17-38% of grainsize <0.02 mm respectively. it is shown that cylinder 2 and 3 show a very high clay content of >40% thus highlighting a greater risk of frost developing (T4). Classification of frost groups are presented in appendix G.e.
5.7 Subsurface interface interpretations

Interpolations from well data is graphically presented in Figure 5-18. In general, interpretations of the sediment bedrock interface demonstrate a shallow sediment layer (~1-4 m) in the northern half of Torshovdalen. It is known from geophysical interpretations that the sediment/bedrock transitional layer in southernmost part decreases with depth to ~15-17 m. GPR interpretation of the subsurface is shown in the following diagrams accompanied by the raw GPR data (Figure 5-19, Figure 5-20, Figure 5-21, Figure 5-22, Figure 5-23 and Figure 5-24). Interpolations of GPR profiles was used to create a 2D representation of depth to bedrock and bedrock topography is presented in Figure 5-18. Interpretation of sediment/bedrock and or bedrock/bedrock transitions was done in collaboration with Trond Eiken from the University of Oslo. As shown in the raw GPR data, vertical axis represent depth in Nano seconds (ns) and the horizontal axis represent time in seconds (s). Interpretive graphs illustrate the digitized boundaries of high reflective surfaces estimated from the raw data profiles. Lines were partially digitized automatically by phase, and partly manually. GPR Profiles was performed at two different field campaigns. Profile P004 was recorded in early fall of 2018. Profiles P50, P53, P54 and P55 were recorded in January of 2019. See appendix C for supplementary information.

Data from field campaigns were treated using Reflexw – GPR and seismic processing software. It is important to note that, velocity was not estimated using typical survey methods. However, a velocity of 0.09 m/ns was chosen based on table values corresponding to silty soil texture (GPRRental, 2019, Eiken, 2019).
Figure 5-18: Depth to bedrock interpretation estimated from GPR profiles.
Figure 5.19: Profile P53, crossing the southernmost part of the valley. Above: Graph illustrate the depth and distance interpreted from the raw GPR data. The location of the profile is presented graphically adjacent to the graph. Below: Raw data from GPR recordings.

Figure 5.20: Interpretive plot from Profile P54, profile is crossing the southernmost point by well 3 location. Above: Graph illustrate the depth and distance interpreted from the raw GPR data. The location of the profile is presented graphically adjacent to the graph. Below: Raw data from GPR recordings.
Figure 5-21: Interpretive plot from Profile P50. Above: Graph illustrate the depth and distance interpreted from the raw GPR data. The location of the profile is presented graphically adjacent to the graph. Below: Raw data from GPR recordings.
Figure 5-22: Profile P4, recorded along the entire park. Above: Graph illustrate the depth and distance interpreted from the raw GPR data. The location of the profile is presented graphically adjacent to the graph. Below: Raw data from GPR recordings.

Figure 5-23: Profile P55. Profile starts at well 2 and continues to the northern most point in the park. Above: Graph illustrate the depth and distance interpreted from the raw GPR data. The location of the profile is presented graphically adjacent to the graph. Below: Raw data from GPR recordings.
Figure 5-24: Profile 51 is stretching across the entire park. Above: Graph illustrate the depth and distance interpreted from the raw GPR data. The location of the profile is presented graphically adjacent to the graph. Below: Raw data from GPR recordings.
5.8 Mineralogy

5.8.1 Major element geochemistry

Elemental composition was examined from ten samples collected from well 1 (North), well 2 (East) and well 3 (South). Figure 5-25, Figure 5-26 and Figure 5-27, illustrates the evaluation of the mineralogy between sediment (B1-n, B2-n, B3-n) and bedrock samples (F1-n, F2-n and F3-n). Each graph is showing elemental composition for each test in weight percent (wt%).

In general, correlation between all sediment samples from well 1, 2 and 3 display a high SiO$_2$ content of 53-69 (wt%), this also is true for bedrock samples, however, with a distinctively less amount of 39-48 (wt%). In bedrock, strong evidence was found that calcium oxide (CaO) content is significantly higher in bedrock samples of 7-21 (wt%) compared to sediment samples (1-3 wt%). Aluminum oxide (Al$_2$O$_3$) is the second most abundant element composition within the samples and it is apparent in both sediment and bedrock samples (11-18 wt%). Results show that less dominant miscellaneous elements represent < 5%. Levels of Iron oxide (Fe$_2$O$_3$), potassium oxide (K$_2$O) Magnesium oxide (MgO) and Sodium oxide (Na$_2$O$_3$) have a noticeably low presence, however similar between bedrock and sediment samples. Values ranges from 4-9 (wt%), 3-4 (wt%), 1-4 (wt%) and 1-4 (wt %) respectively. See appendix D (a-c) for supplementary information.
Figure 5-25: Elemental Analysis from Well 1 (North). Samples are organized in ascending order with depth. “L” denotes sediment sample and “F” denotes bedrock. Elements are presented in weight percent (wt%).

Figure 5-26: Elemental Analysis from Well 2 (East). Samples are organized in ascending order with depth. “L” denotes sediment sample and “F” denotes bedrock. Elements are presented in weight percent (wt%).

Figure 5-27: Elemental Analysis from Well 3 (South). Sample “F” denotes bedrock. Elements are presented in weight percent (wt%).
5.8.2 Clay classification and mineral identification

Ten samples were evaluated for bulk and clay classification: Six samples from well 1, three samples from well 2 and one sample from well 3. The refinement process is reported in each figure caption.

**Bulk composition**

In general, samples taken from the sediment layer from well 1 (Figure 5-28), well 2 (Figure 5-29), and well 3 (Figure 5-30) show a high content of quartz and plagioclase that makes up ~60% in each sample. Comparatively, in bedrock, quartz and plagioclase makes up ~35%. Potassium feldspar (K-spar) show a higher content in sediment samples. Interestingly, there is a significant difference where well 1 display a higher content of K-spar (13-19%) in sediment samples compared to well 2 and well 3 samples (~6%). In bedrock, K-spar is ranging from ~2-7% for all samples. Mica is found of substantial levels in all samples, considering well 1 and well 2 that are the most similar in composition. It is apparent from that Calcite content is significantly increasing in bedrock samples (13-36%) compared with sediment samples (0.1-5%). Well 1 and 2 is different from well 3, in that calcite content is ~10% in sediment sample. Chlorite is present in every sample at varying content from 3-19%. Pyrite is mostly found in bedrock samples and constitutes ~1-2%.

![XRD Well 1 (North)](image)

*Figure 5-28: Six samples from Well 1 (North) was evaluated. Samples are represented with depth. Convergence progress data. Rwp 10.17; Rexp 8.41. X² 1.46; GOF 1.2. “Other” comprises of various minerals: Magnesite and Graphite.*
Clay classification

Clay classification was performed on two sediment samples from different well locations: well 2 (East) and well 3 (South) samples B2-L1 and B3-L4 respectively. Figure 5-31 and Figure 5-32 below illustrate results in a diffractogram with four different treatment segments superimposed: 1) No treatment, 2) ethylene glycol (EG), 3) heating at 350°C and 4) heating at 550°C. Labels on both figures indicate location as d-spacing denoted (Å) of the different clay groups. Graphs show intensity energy plotted against 2θ diffractive angle.

Main clay groups found within each sample (B3_L4 and B2_L1) contain members of the chlorite, kaolinite and illite group. It can be seen from the data in Figure 5-31 from sample (B2_L1), the most pronounced clay group was Illite, which showed a typical (001) peak at 10.5 Å and (003) peak at 3.3 Å. After treatment of glycol ethylene (EG), all peaks were weakened, however, peaks were intensified following heating processes. Chlorite (001) peak occurs at 14 Å, here, the peak is destroyed after going through heat treatment. Another peak
indicating the chlorite group is a weak 7 Å (003) peak which in turn collapses after treatment of 550°C. There is also a very weak chlorite (060) peak at 1.54 Å and a much higher chlorite (002) at 4.70 Å. Kaolinite shows peaks at 7.18 Å (001) and 3.58 Å (002) (Poppe et al., n.d).

As seen in Figure 5-32, sample (B3-L4) have a characteristic illite group peak (001) at 10.5 Å and (003) peak at 3.3 Å. Peaks remain unaltered after EG, as can be seen from the changes in the superimposed diffractogram. However, heat treatment of 350°C has weakened the peaks. Chlorite have both (001, 060) peak at 14 Å and 1.54 Å respectively. Peaks remain unchanged after EG treatment, however, heat treatment of 550°C destroys the 14Å peak completely. Presence of Kaolinite is found at peaks located at 7.18Å and 3.58Å (Poppe et al., n.d).
Figure 5-31: Diffractogram of Sample B2_L1 showing four different treatment processes superimposed.

Figure 5-32: Diffractogram of Sample B3_L4 showing four different treatment processes superimposed.
5.9 Rate of infiltration

Data from infiltration tests are presented graphically Figure 5-33, here, infiltration rate (cm/hr) is plotted against duration in hours. Fourteen infiltration tests were performed: Three from transect 1 (Ln-T1), three from transect 2 (Ln-T2), four from transect 3 (Ln-T3) and four from transect 4 (Ln-T4). Each infiltration rate has been corrected according to a correction factor recommended by (Solheim, 2018): For this thesis, all infiltration rates have been corrected with 0.6. Overall Ksat estimations show varying rates between each transect ranging from 3-30 cm/h. Common experience for all testing locations was the fast rate obtained after 1 cm, which was ranging from 64-540 cm/h. See appendix E for supplementary information.
Figure 5-33: Infiltration rates from fourteen MPD measurements divided by four transects in Torshovdalen

5.10 Diver data and groundwater flow

5.10.1 Rate of in-flux

Purging well 3 (South) was performed to evaluate the rate of in-flux of water from the fracture rock aquifer to the well. The mean rate established from purging the well four times show that water is flowing in to the well at a rate of 0.6 m/h. See appendix F for supplementary information.

Table 17: Purging well 3 (South) to establish rate of water in-flux to the well.

<table>
<thead>
<tr>
<th></th>
<th>Δhead (-)</th>
<th>Time (hr)</th>
<th>(m/h)</th>
<th>(m/s)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Pumping 1</td>
<td>0.3</td>
<td>4.0</td>
<td>0.1</td>
<td>2E-05</td>
</tr>
<tr>
<td>Pumping 2</td>
<td>0.8</td>
<td>4.2</td>
<td>0.2</td>
<td>5E-05</td>
</tr>
<tr>
<td>Pumping 3</td>
<td>1.7</td>
<td>1.2</td>
<td>1.4</td>
<td>4E-04</td>
</tr>
<tr>
<td>Pumping 4</td>
<td>1.3</td>
<td>1.5</td>
<td>0.9</td>
<td>3E-04</td>
</tr>
<tr>
<td><strong>Average</strong></td>
<td><strong>1.0</strong></td>
<td><strong>2.7</strong></td>
<td><strong>0.6</strong></td>
<td><strong>2E-04</strong></td>
</tr>
</tbody>
</table>

5.10.2 CTD Diver data

Data from well logging including: Groundwater level (m), temperature (°C) and conductivity (mS/cm). Data was retrieved from a CTD diver placed in well 3 (South) from December 27th 2018 to April 1st 2018. As can be seen from Figure 5-34, Groundwater level is undulating between 0.84-2.6 m below ground surface. Groundwater level is trending downwards from -1.34 m in late December until reaching the lowest level at -2.63 m February 10th. Highest groundwater level was recorded on March 23rd at -0.84m.
Temperature and conductivity are graphically presented in Figure 5-35, Conductivity readings show a steady conductivity ranging from 1.1-1.2 mS/cm, where the lowest value was recorded on December 28th and the highest was recorded on February 10th. Temperature readings revealed steady values at ~8°C. See appendix F for supplementary information.

**Figure 5-34: Water level fluctuations recorded from December 27th until April 1st at well 3 (South) location.**

**Figure 5-35: Data from Conductivity and temperature readings.**
5.11 Field observation

Locations for field observation and sampling transects are shown graphically in Figure 5-36. The figure shows an overview of Torshovdalen including important observations such as water accumulation, how the snow was deteriorating, and the ephemeral spring seepage.

![Field Observation Diagram](image)

Figure 5-36: Gray area indicates the max extent of flooding when it occurs based on observation. White circles are areas with intermittent/ephemeral springs, red dotted lines mark the transect for soil property assessment, and red arrows point towards the respective profile. Profiles are not for scale.

The outcome of snow melting was observed in Torshovdalen from January 2018 to May 2018. The purpose of this experiment was to get familiarized with the study area and establish zones prone to ponding. Map of Torshovdalen Figure 5-36 show where ponding was dominant, namely at A1, A2 and A3. From current observations it was apparent that water was frequently accumulating at these locations during the spring melt of 2018. Pictures documented below (Figure 5-38 March 25th 2018 was Sunny. with temperatures of low -0.9 and high 10 degrees
°C. Pictures show Area 2 (A2) where ponding is eminent., Figure 5-37, Figure 5-39 and Figure 5-40) show the spring melt transformation.

March 25th 2018 (Figure 5-37, March 25th 2018 was Sunny, with temperatures of low -0.9 and high 10 degrees °C. Pictures show Area 2 (A2) where ponding is eminent.) experienced high temperature of 10°C and low of -0.9°C. Large amounts of melt water had accumulated throughout the park, especially at designated areas (A1, A2, and A3). Pools and streams of water were flowing through and over the snow cover. April 13th 2018 (Figure 5-37) revealed a notable difference in the snow cover and experienced high temperatures of 13.4°C and low of 2.8°C. Snow was scarce, however ponding was still occurring. As can be seen from Figure 5-37, area A3 was still affected by the snow.

April 20th 2018 (Figure 5-39) experienced high temperatures of 14.2°C and low of 6.2°C. However, ground was damp, and water was still collecting at designated areas. Water from seepage areas were experiencing heavy water outpour, as can be seen from Figure 5-39. Ephemeral outlets showed an organic matter/ bio-film covering the seeping water. Water was showing both a discoloring of brown/red, but also a look of iridescence. Snow was also discolored red/brown at- and around the lower springs. Water accumulations at more horizontal segments exhibited a steady water depth of ~0.5 cm. On May 10th 2018 ephemeral springs was experiencing a water outflow averaging between 0.8-1.8 L/min. As can be seen from Figure 5-40, fall precipitation outcomes provide evidence for the aforementioned investigation of water accumulating. Area A2 especially, signifies an area of great accumulation of water.
March 25th

Figure 5-38 March 25th 2018 was Sunny, with temperatures of low -0.9 and high 10 degrees °C. Pictures show Area 2 (A2) where ponding is eminent.

April 13th

Figure 5-37: April 13th 2018 was Sunny, with temperatures of low 2.8 and high of 13.4 degrees °C. Pictures show Area 3 (A3) where water is in the process of drying up from the ground. However, water has remained in this area although the park is considerably dry.
Figure 5.39: April 20th 2018. Overcast, humid and chilly. With low temperatures of 6.2, and high of 11.4 °C.

Figure 5.40: November 12th 2018. Heavy precipitation has created ponds of water accumulating. Pictures are taken at Area 2 (A2).
5.11.1 Soil moisture content

Soil Moisture was measured along four transects throughout the summer and early fall of 2018. Figure 5-41 display four graphs representing the outcome of soil moisture and variation regarding seasonal changes. The upper two graphs differentiate soil moisture fluctuations on June 8th and October 10th representing spot seasonal changes for all transects. The lower two graphs highlight only transect 3, representing summer and fall months. Soil moisture is given in percentage (%): Zero values indicate discrepancies while measuring.

The analysis found evidence of distinct differences between transect 1, 2 and 4 compared to transect 3. As shown in Figure 5-41 in the upper two graphs, June 8th 2018, soil moisture values display transect 1, 2 and 4 in which experienced a steady fluctuation of soil moisture ranging between 4-21 %, while transect three was ranging with markedly higher soil moisture values of 7-64 %. Measuring’s performed on October 10th 2018 show a similar trend: Transect 1, 2 and 4 show values ranging between 19-43 % while transect 3 is ranging between 20-63 %.

The bottom half of Figure 5-41 show the soil moisture result from transect 3 in which ephemeral springs is located. At approximately 15-20 m and 75-80 m in both figures, ephemeral springs are showing peaks of soil moisture ranging between 50.8-61.6 % and 40.7-63.8 % for the summer months, and 46.5-60.7 % and 43.5-63.2 % for the fall months. See appendix I (a-e) for supplementary information.
Figure 5-41: Upper two graphs show all of the transects superimposed in a single day to show variation, June 8th and October 10th 2018. Lower graphs show how transect three fluctuate seasonally. Graphs illustrate soil moisture content captured in field campaigns during the summer and fall.
5.12 Water balance

The water balance for Torshovdalen was estimated by using equations developed by Thornthwaite in 1944 (Thornthwaite and Mather, 1957). However, other equations have also been implemented to supplement calculations from Pascual-Ferrer and Candela (2015), Thornthwaite and Mather (1957), Mccabe and Markstrom (2007), NRCS (1986). The approach was to roughly estimate climatic variabilities using simple empirical calculations. Results from correlational analysis is summarized graphically in Figure 5-42, Figure 5-43 and Figure 5-44 and represent water balances for the two consecutive hydrologic year of 2016/17, 2017/18. In addition, a water balance representing predicted climate change are also presented.

Monthly precipitation and temperature averages were collected from Blindern weather monitoring station located close to the study area. The result of the water balance estimation from 2017/18 is summarized in Table 18. Table demonstrate the breakdown of the water balance and provide a rough estimation for the behavior of water in Torshovdalen. See appendix H (a-e) for supplementary information. Field capacity is assumed to be 100 mm.
Table 18: Estimated water balance of the soil for 2017/2018 for the area of Torshovdalen sub-catchment.

<table>
<thead>
<tr>
<th>2017/2018</th>
<th>Fall mm</th>
<th>Winter mm</th>
<th>Spring mm</th>
<th>Summer mm</th>
<th>mm/yr</th>
<th>m³/yr</th>
<th>l/s*km²</th>
</tr>
</thead>
<tbody>
<tr>
<td>P</td>
<td>307</td>
<td>171</td>
<td>59</td>
<td>104</td>
<td>642</td>
<td>89866</td>
<td>20</td>
</tr>
<tr>
<td>Snow</td>
<td>52</td>
<td>171</td>
<td>15</td>
<td>0</td>
<td>238</td>
<td>3335</td>
<td>1</td>
</tr>
<tr>
<td>Rain</td>
<td>255</td>
<td>0</td>
<td>44</td>
<td>104</td>
<td>404</td>
<td>5651</td>
<td>1</td>
</tr>
<tr>
<td>W-PET</td>
<td>207</td>
<td>171</td>
<td>223</td>
<td>0</td>
<td>600</td>
<td>84043</td>
<td>19</td>
</tr>
<tr>
<td>PET</td>
<td>115</td>
<td>0</td>
<td>176</td>
<td>466</td>
<td>757</td>
<td>105974</td>
<td>24</td>
</tr>
<tr>
<td>RET</td>
<td>115</td>
<td>0</td>
<td>44</td>
<td>104</td>
<td>263</td>
<td>36858</td>
<td>8</td>
</tr>
<tr>
<td>SS</td>
<td>192</td>
<td>171</td>
<td>15</td>
<td>0</td>
<td>379</td>
<td>53008</td>
<td>12</td>
</tr>
<tr>
<td>Deficit</td>
<td>0</td>
<td>0</td>
<td>-132</td>
<td>-362</td>
<td>-494</td>
<td>-69116</td>
<td>-16</td>
</tr>
<tr>
<td>Surplus</td>
<td>50</td>
<td>24</td>
<td>0</td>
<td>0</td>
<td>74</td>
<td>10362</td>
<td>2</td>
</tr>
<tr>
<td>Snowpack</td>
<td>38</td>
<td>404</td>
<td>224</td>
<td>0</td>
<td>666</td>
<td>93298</td>
<td>21</td>
</tr>
<tr>
<td>Snowmelt</td>
<td>14</td>
<td>0</td>
<td>224</td>
<td>0</td>
<td>238</td>
<td>33360</td>
<td>8</td>
</tr>
<tr>
<td>RO</td>
<td>9</td>
<td>0</td>
<td>69</td>
<td>24</td>
<td>103</td>
<td>14351</td>
<td>3</td>
</tr>
<tr>
<td><strong>Groundwater</strong></td>
<td><strong>Recharge</strong></td>
<td>41</td>
<td>0</td>
<td>45</td>
<td>0</td>
<td>86</td>
<td>12018</td>
</tr>
</tbody>
</table>

**Precipitation (P), water input (W), potential evapotranspiration (PET). Reference evapotranspiration (RET), soil storage (SS) and Runoff (RO).**

Figure 5-42 illustrate parameters from the water balance calculation from 2017/18. As seen in the figure, during the fall months 2017: September (2018), October and November, Precipitation (P) is greater than potential evapotranspiration (PET), thus soil storage accumulates (164 mm) exceeding the field capacity, and thus water surplus occurs. and reference evapotranspiration (REF) equals PET. Under these conditions, soil storage starts to collect water. Surplus water was detected (50 mm) during the fall season, demonstrating the “wettest” conditions of the hydrologic year of 2017/18.

In winter, during December, January and February, W (P + snowmelt) was 171 mm, indicating an addition to soil storage. Also, the snowpack was accumulating (404 mm) holding water in its solid phase anticipating a cumulative snow water equivalent of 166 kg/m². Notably, during spring months significant snowmelt occurred in March of 2018 (224 mm) and P was greater than PET contributing to runoff values (69 mm). A surplus of water was demonstrated in both fall and winter, subsequently, in April, May and the summer months, PET was greater
than P and the storage experienced a deficit of water (-362 mm), this would be continuing over
the summer with increasing water deficit (-494 mm).

Torshovdalen and Torshovbekken catchments was divided into separate land cover
groups in order to specify the recharge and runoff percentage. Table 19. The land cover was then
divided into soil groups based on different infiltration capacities (NRCS, 1986), values are
generated from an estimation over Torshovbekken catchment area using Nevina (NVE, 2019).
Considering two Aquifers: unconsolidated sediment aquifer and a fractured bedrock aquifer.

Table 19: Different estimation of water balance parameters with respect to different soil hydrologic groups.

<table>
<thead>
<tr>
<th>Land cover</th>
<th>Area (km²)</th>
<th>P (m³/yr)</th>
<th>RO (m³/yr)</th>
<th>Infiltration (m³/yr)</th>
<th>PET (m³/yr)</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Torshovbekken:</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Forested (6.9%)</td>
<td>5.15</td>
<td>0.35</td>
<td>228099</td>
<td>214780</td>
<td>42664</td>
</tr>
<tr>
<td>Other (30.9%)</td>
<td>1.60</td>
<td>1.60</td>
<td>1028099</td>
<td>968068</td>
<td>192296</td>
</tr>
<tr>
<td>Urban (62.2%)</td>
<td>3.19</td>
<td>3.19</td>
<td>2049587</td>
<td>1929911</td>
<td>383356</td>
</tr>
<tr>
<td><strong>Torshovdalen:</strong></td>
<td>0.14</td>
<td>0.14</td>
<td>89866</td>
<td>102472</td>
<td>16809</td>
</tr>
<tr>
<td>Other (100%)</td>
<td>0.14</td>
<td>0.14</td>
<td>89866</td>
<td>102472</td>
<td>16809</td>
</tr>
</tbody>
</table>
Figure 5-42: Figure show the estimated average monthly soil water balance for Torshovdalen area in the hydrological year of 2017/2018.

Figure 5-43: Figure show the estimated average monthly soil water balance for Torshovdalen area in the hydrological year of 2016/2017.

Figure 5-44: Soil water balance with increased climatic factors, based on the hydrologic year of 2016/2017. Precipitation and Temperature is increased monthly by 10% and 0.5°C respectively.
Considering Torshovdalen sub-catchment, estimations of the annual change of storage was made. Parameters used to estimate the change in storage were taken from the annual soil water balance. Notably, $G_{\text{win}}$ is the amount of water that is made available for infiltration after subtracting the PET and not considering the available water for runoff. $G_{\text{out}}$ is the sum of the water infiltrating from the larger catchment of Torshovbekken (5.15 km²) and the water infiltrating from Torshovdalen (0.14 km²). Table 20 demonstrate the difference between the aforementioned hydrological years.

<table>
<thead>
<tr>
<th>Year</th>
<th>Change in storage</th>
<th>Year</th>
<th>Change in storage</th>
</tr>
</thead>
<tbody>
<tr>
<td>2017/2018</td>
<td>(P + $G_{\text{win}}$) - (Q + RET + $G_{\text{out}}$)</td>
<td>2016/2017</td>
<td>(P + $G_{\text{win}}$) - (Q + RET + $G_{\text{out}}$)</td>
</tr>
<tr>
<td>149</td>
<td>20886794</td>
<td>138</td>
<td>19371072</td>
</tr>
<tr>
<td>20887</td>
<td>1</td>
<td>19371</td>
<td>0.61</td>
</tr>
<tr>
<td>5</td>
<td></td>
<td>4</td>
<td>28</td>
</tr>
<tr>
<td>Climate change</td>
<td>895</td>
<td>Climate change</td>
<td>125322112</td>
</tr>
<tr>
<td></td>
<td>125322</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

Groundwater flow in the area of Torshovdalen was calculated based on numbers from the previously estimated budget for the soil. Bedrock topography and surface topography of Torshovdalen is graphically illustrated in Figure 5-45. The figure show the difference in depth to bedrock, indicating the different cross-sectional area the water may be entering. Groundwater is entering the Torshovdalen area from the greater Torshovbekken catchment at 0.0001 m³/s through a cross-sectional area of ~240 m² (A1) and is exiting Torshovdalen at 0.0003 m³/s through a cross-sectional area of ~985 m² (A2). Watershed gradients are calculated based on topography from both sub catchments (0.06-0.09). Available water is calculated from the water balance parameters (P+MELT-PET) and represent the potential water available in the unconsolidated aquifer. Interestingly, as shown in Table 21 the volumetric flow calculated from the fracture found at well location 3 (South) had a surprisingly high hydraulic conductivity of ~6 m/s, and a volumetric flow of ~2 m³/s. Permeability for the surrounding Shale and nodular limestone bedrock have a permeability ranging between 3x10⁻⁷-2x10⁻⁶ m/s (Norconsult et al., 2000), that yields a volumetric flow ranging between 2x10⁻⁸-1x10⁻⁷ m³/s·m⁻².
Figure 5.45: 3-Dimensional representation of the relationship between surface and bedrock. Depth to bedrock is denoted in the northern part and the southern part of Torshovdalen catchment showing the great difference in cross-sectional area of which water is entering and exiting.

Table 21: Numbers are calculated with data found in Morgan (2004) Makurat et al. (1990), Norconsult (2018) and Hiscock and Bense (2014).

<table>
<thead>
<tr>
<th>In Torshovbekken (5.15 km²):</th>
<th>Out Torshovdalen (0.14 km²):</th>
</tr>
</thead>
<tbody>
<tr>
<td>( W_{\text{available}} = 0.02 \text{ m}^3/\text{s} )</td>
<td>( W_{\text{available}} = 0.0005 \text{ m}^3/\text{s} )</td>
</tr>
<tr>
<td>( Q_{\text{in}} = 0.0001 \text{ m}^3/\text{s} )</td>
<td>( Q_{\text{out}} = 0.0003 \text{ m}^3/\text{s} )</td>
</tr>
<tr>
<td>( \Delta h = 0.09 ) ( )</td>
<td>( \Delta h = 0.06 ) ( )</td>
</tr>
<tr>
<td>( A_{1\text{cross-section}} = \sim 240 \text{ m}^2 )</td>
<td>( A_{2\text{cross-section}} = \sim 985 \text{ m}^2 )</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>Bedrock ( b )</th>
<th>Fracture in bedrock ( f )</th>
</tr>
</thead>
<tbody>
<tr>
<td>Aperture = 0.000039 m</td>
<td>Aperture = 0.123 m</td>
</tr>
<tr>
<td>Spacing = 0.05-3 m</td>
<td>( K_f = 6 \text{ m/s} )</td>
</tr>
<tr>
<td>Permeability = 0.0000003 m²/s</td>
<td>( Q_f = 2 \text{ m}^3/\text{s} )</td>
</tr>
<tr>
<td>( Q_b = 2 \times 10^{-13} - 1 \times 10^{-11} )</td>
<td>( T_f = 1 \text{ m}^2/\text{s} )</td>
</tr>
<tr>
<td>( \Delta h = 0.06-0.09 ) ( )</td>
<td>( \Delta h = 0.06 ) ( )</td>
</tr>
</tbody>
</table>
6 Discussion

In this chapter, interpretation of results from chapter five are evaluated and discussed. Understanding the applicability and importance of hydrogeological and physical characterization of the subsurface in urban development planning is an important aspect. In addition, recognize how climate change affects the groundwater recharge and flow into- and out of the catchments as well as how the soil textural properties affect the conduct of the water.

Thorough characterization and site assessment is imperative in urban development planning (Craul, 1999). Site assessment of Torshovdalen catchment was initiated by the water and sewerage works of Oslo municipality as a part of a larger interdisciplinary study. The study was led by PHD candidate Julia Kvitsjøen: The purpose of the project was to implement methods that will offer cost effective solutions for storm water management best practices.

6.1 Geological heterogeneities and aquifer properties

Installation of monitoring wells for hydrogeological assessment in Torshovdalen was done in order to produce detailed status on the sediment- and bedrock properties and groundwater conditions. A major consideration for well implementation was the possibility of geophysical wireline logging. In addition, by establishing a monitoring scheme by utilizing a datalogger (CDT-Diver) in well 1 (North), it was also possible to evaluate environmental changes and corresponding actions. Location of the wells was proposed based on prior geological knowledge of the area. Several parameters were discussed: 1) Possible triangular correlation across the entire park, 2) surface topography and 3) depth to bedrock. In addition, prior geological knowledge of the area suggested that the shale and nodular limestone layering may be fractured to various degrees and may be considerably water-bearing (Morgan, 2004, Løset, 2014, Norconsult, 2018). Three wells were constructed representing the extent of the park: North, East and South locations (Figure 6-1).

However, the desired well zones were decided in cooperation with Norwegian Geotechnical Institute (NGI), Agency for Urban Environment (BYM), and the Agency for Water and Sewerage Works in Oslo (VAV). Therefore, considerations were made according to the different agency’s prerequisite for well implementation. Initially, wells were planned to be located in a triangle fashion positioned in the center of Torshovdalen. A golden middle way was reached in order to satisfy both the initial objectives and the agencies of the municipality.
In the end, well locations were satisfactory according to the planned investigative extent that satisfied the sediment/bedrock interface assessment. Although well construction was successfully implemented with various results, one might argue if other types of wells should also have been considered. E.g. Monitoring wells in the unconsolidated aquifer. Its important to note that that the location of well 1 (North), could implicate the accuracy of water levels and the quality of the water as it was placed close to a main road (Aller et al., 1991).

Findings provided a better knowledge of the depth to bedrock and unconsolidated sediment thickness. Showing that the bedrock topography is controlling the surface topography, and that the lower half of Torshovdalen show a thicker and more heterogeneous stratigraphy. In addition, findings made it possible to interpolate the collected stratigraphy and lithology to estimate the unconsolidated porous aquifer. These findings may support the hypothesis that there are two aquifers; one fractured rock and one unconsolidated porous aquifer. The sudden increase in sediment thickness may be explained by the fault zone crossing diagonally though Torshovdalen (Figure 3-2). This is discussed further in the next sections.

The initial drilling rig (total sounding, Gm100GT) used for the project is produced for the rough determination of the stratigraphy, and for shallow penetration of bedrock to no more than ~3 m depth. However, for the purpose of the investigation one would argue that using a stronger rig (Vibration, Eijkelkamp sonicsampdrill) and using a larger casing whilst drilling, would produce more stable boreholes for geophysical logging. Although a sonic rig was brought in when problems arose, it is important to notice the importance of evaluating the equipment needed. As a reoccurring problem among all well locations, cuttings clogged the boreholes and impeded the continuation of drilling and the casing used for drilling got stuck multiple times. Proper sealing of the well was not accomplished at the eastern nor the southern well. This may implicate the contamination of the groundwater from water seeping from the unconsolidated aquifer into the fracture rock aquifer.
The construction of well 3 (South) was done without using an outer casing when penetrating the ground. However, a polymer drilling fluid was implemented to ease the drilling with only the sounding rod. At the final stage of implementation, the well casing was forced into the borehole using the rig for support. It was thus not possible to place neither bentonite nor sand in between the borehole and well casing. It is important to bear in mind that there may be some groundwater leakage from the unconsolidated aquifer above and into the filter zone of the fractured aquifer, contaminating the deeper water in the well. Because of this potential limitation, water quality and true in-flux may be influenced. However, evidence provided by Kristiansen (2019) suggested that the high clay content of the unconsolidated sediments of Torshovdalen hindered surface water from leaking from the unconsolidated aquifer into the fracture aquifer. These issues highlight the importance of proper well construction, design and sealing (Aller et al., 1991). From the current observational study and drilling log, it was also
hypothesized that lenses of clay were likely to occur in the unconsolidated sediments. The comparison of findings is conceptualized in Figure 6-2 and Figure 6-3. Profiles represent findings on the subsurface stratigraphy and are heavily based on the integration of different sources of data acquired in the field including drilling logs, grainsize distribution and ground penetrating radar (GPR).
Figure 6-2: Map illustrates the transect in which stratigraphic profiles were made. Red lines indicate the transects, and red stars indicate the location of the wells.

Figure 6-3: Conceptual geological profiles based on results from subsurface investigation. Vertical scale of profiles are exaggerated.
6.1.1 Fractured rock aquifer

At well location 1 (North), geophysical logs did not indicate any fractures although there was evidence of density anomalies confirmed 4-6 m into the shale formation. The findings of shale are consistent with local stratigraphy where Elnes- and Vollen formation is expected according to Graversen et al. (2017), these formations display an alternating shale and nodular limestone which is typical for the Cambro Silurian sedimentary sequence found in the area (Bjørlykke, 2004). Furthermore, this was represented by the variation in sound amplitude signal in the borehole showing alternating layers of high density (Limestone) and low density (shale) (Athy, 1930). These findings are also illustrated in the acoustic caliper log that show a set of dark colored, semi sinusoidal waves that are indications of liquid filled holes in the shale (Crain, n.d).

Resistivity logs also showed undulating measurements suggesting a difference in permeability. This imply that there may be available groundwater in the low resistivity layers (Varhaug, 2016). According to Størmer (1953), Øset (2014), dikes and veins are commonly found intercepting the Cambro-Silurian bedrock of the Oslo region. Although many fractures may be water bearing, purging well 1 (North) and monitoring the in-flux of water did not demonstrate any evidence of water entering the well. The filter casing may also be congested due to drilling fluid and sediments in suspension. However, flushing the well did not change the in-flux of water. Alternatively, this well did simply not intersect the water table at this location.

Common faulting from the Permian igneous activity constituted normal faulting that coincides with the Oslo graben (Ramberg and Smithson, 1971). A fault is diagonally crossing the middle of Torshovdalen representing a Cambro-Silurian rock boundary between the heavily broken down ungrouped sedimentary rocks and the alternating shale and nodular limestone (Graversen et al., 2017) as seen in Figure 3-2. One interesting finding was the angular granitic/salmon colored clasts that appeared in the mass of cuttings while drilling (Figure 5-1). These clasts appeared after penetrating ~1.5-2 m in bedrock. It is possible to hypothesize that these fragments may be from either the granite basement or from the basal conglomerate that is underlying the alum shale formation (Gabrielsen et al., 2015). Findings was highly unexpected and thus suggests that we might be lower in the stratigraphic column than anticipated. On the other hand, these clasts may be fragments from the expected igneous intrusions (Morgan, 2004, Bruton et al., 2010). These findings are in line with studies done by
(Norconsult, 2018) that found evidence of syenite and rhomb porphyry as igneous intrusive dykes along the Sogn-Ulven stretch when planning an underground tunnel in Oslo. Syenite porphyry is a phaneritic plutonic rock which has normally high content of orthoclase (Winter, 2010) which may explain the occurrence of the salmon colored clasts. The existence of these types of dykes is recognized by (Løset, 2014). These igneous intrusions are anticipated to be more water bearing than the surrounding sedimentary rocks.

Typical bedding structure within the Cambro-Silurian sedimentary sequence have a simple NE-SW strike with a vertical joint set running perpendicular to the bedding planes (Makurat et al., 1990). Alternating shales and nodular limestone layers contain a series of compaction cleavages of plane joints that are parallel to the bedding surfaces and have a strike direction in the NE-SW and NW-SE direction. Contrary to expectations, as seen in Figure 6-4, a section of massive limestone was found outcropping in the middle of Torshovdalen. This formation displays joints represented striking in the dominant joint-set directions (NE-SW and NW-SE).

This massive limestone formation, according to the stratigraphic column (Figure 3-1), is very unexpected considering the formation expected to be found in. There are possible explanations for this assumption. One hypothesis is that this limestone may belong to the lower Huk- or Bjørkåsholmen formation. As mentioned in the previous paragraph, this hypothesis is built on the findings of granitic clasts encountered at well 1 (North), implying that there may be missing sedimentary sequences in which have been uplifted and eroded as a result of faulting.
Well 3 (South) intercepts a thick unconsolidated sediment layer of ~13 m before entering the fractured shale layer. The results from purging well 3 (South) showed a high influx of water to the well indicating that the water table was reached. This is also supported by the results of the geophysical logs that demonstrated a large joint aperture of 0.13 m at 18.5 m depth. This is likely an important groundwater bearing fracture. The fracture system is interpreted by findings and literature and presented in Figure 6-5. Previous excavation may have resulted in a change in the fracture aperture causing some fractures to become relatively large (Makurat et al., 1990). The surrounding matrix and the smaller joints in the adjoining fracture system is suggested to be of relatively low permeability (Morgan, 2004). Water in-flux to the well proved to be of high volumetric rate (2 m³/s), which may indicate that the large fracture may be of considerable length intercepting the well, therefore, a diver was installed to monitor the fluctuation in the water table, temperature and electrical conductivity.
Figure 6-5: Principle simulation of the Cambro Silurian joint and fracture system. Figure show how the wells theoretically intersect the fracture. Adapted and Modified from Makurat et al. (1990). Strike and dip illustration is made from visualgeology.

The strike and dip of the fracture measured 260°/62° SW, which is in the dominant fracture directions in the Cambro-Silurian (Norconsult, 2018, Makurat et al., 1990). Natural gamma (NG) and resistivity showed evidence of alternating radioactive layer of shales and less radioactive layer of limestone (Russell, 1944). The jump in the resistivity coincides with the increase in density, subsequently, NG coincides with findings of radioactive matrix which yields high NG with high clay content (Crain, n.d). Which can be seen from the local stratigraphy depicted in Figure 3-1.

Collectively, from well 1 (North), well 2 (East) and well 3 (South), the mineralogy results from X-ray diffraction experiments show that high levels of calcium (Ca) was found in bedrock samples (B1-F1, B1- F1.5, B1-F4, B2-F1, and B2-F4) which indicated a Ca-rich sedimentary rock such as limestone (Boggs, 2011). This observation supports the hypothesis that limestone may be more dominating in the area than anticipated. In general, presence of calcite and mica was of substantial quantity in all bedrock samples. In addition, there was distinctively less
plagioclase and K-spar in these samples. According to Bjørlykke (1974) the major constituents in shales of the Oslo region are Illite and Chlorite which correspond to findings from the clay classification in this study. Also, mica and quartz are common in both shale and calcareous layers of various extent (Bruton et al., 2010, Bjørlykke, 1974). Mineralogy from soil and sediment samples demonstrated that the general mineral composition where quartz, and feldspar (plagioclase and potassium) followed by mica and chlorite. The findings may imply that some of the constituents of the sediment sample may be from chemical- and physical weathering of the sedimentary rocks. In addition, quaternary deposits are of glacio-fluvial and marine deposits that may have impacted the mineral materials in the unconsolidated sediments (Rosenqvist, 1972).

Studies done by Kristiansen (2019) found evidence of connection between the dissolution of minerals from the Cambro-Silurian limestone and the groundwater quality. The weathering and dissolution of minerals are clearly affecting the distribution of major elements. Silicate weathering also effect the groundwater quality, by the dissolution of minerals such as sodium, potassium and magnesium (Kristiansen, 2019). Chloride was a major contributor to the groundwater composition, which may be derived from precipitation, weathering of halite or the leaching of porewater from the marine clays. Another source explaining the salinity in the sample may be due to the road salting that occurs during the winter. According to Rosenberry et al. (1999) as cited by Kristiansen (2019), suggests that road salt contamination may even penetrate the bedrock and contaminate the groundwater, which in turn may explain the high content of NaCl found in the groundwater samples. This may also hinder the formation of frozen ground in the unsaturated zone as the freezing point is increased by the presence of dissolved salts (Andersland and Ladanyi, 1994). Groundwater samples exhibited high pH levels compared to that of other water sources demonstrating a basic composition. This may be due to the carbonate system that may be controlling the pH in the groundwater.

Ground penetrating radar (GPR) was used to capture the subsurface interface reflections with a special interest in the loose sediment- bedrock contact. A 50 MHz antenna worked best when interpreting the shallow interfaces, and since the geophysical logs previously demonstrated a varying bedrock topography (shallow in the northern half, and deeper in the southern half) resulted in a change of antennas from 50-250 MHz to get a deeper visualization. The study confirms the findings from GPR which revealed a shallow unconsolidated sediment layer of 0-3 m above bedrock in the northern part, and a deeper 13 m layer in the southern part.
which was proven by the results attained in the drilled wells. Interpolating GPR data, a surface model of the bedrock topography was built. There is a step in the bedrock topography in the middle of the valley, coinciding with the location of the ephemeral springs. This may imply that the springs are due to this phenomenon. In addition, a fault zone is expected to cross approximately through this area (Figure 3-2). It is therefore likely that the observed drop in bedrock topography could be attributed to the uplift due to normalfaulting.

Moreover, graphics presented by Figure 5-18 to Figure 5-24 represent interpretation of distinct transitions between type of sediment/bedrock interfaces. Evaluating the aquifer boundaries, it is suggested that the springs are outcropping the water table, which in turn seem to be controlled by the bedrock topography. Alternatively, the springs may be a result from a perched water table. From an observational point, it is arguable if this spring might be intermittent and not ephemeral as flow occur both in response to seasonal variability and is event based (Dingman, 2015). As seen in Figure 5-21 and Figure 5-24, bedrock is approaching the land surface around the location of the springs, implying that the water is not penetrating the bedrock and is forced to seep out onto the surface and may be due to the low hydraulic conductivity through Cambro-Silurian rocks (Morgan, 2004). Another possibility is the outpour because of the raising of the water table (Fetter, 2001).

In some of the deeper profiles there are several clear reflective surfaces that may be a transition between different type of rock interfaces. Stratification appears to show diagonal discontinuities and dome-shaped reflections which may correspond to the high frequency folds striking in the SW-NE direction (Morley, 1987). The Cambro-Silurian sedimentary rocks are characterized by folding and thrusting (Bruton et al., 2010, Morley, 1987), it is thus arguable if this phenomenon may be the dome-shaped reflections seen in the GPR profiles. Alternatively, The diagonal reflections may also represent igneous activity such as dykes and veins (Morgan, 2004). It is important to note that geophysical logging was performed in a separate and deeper borehole. This borehole was produced next to well 1 (North) and well 3 (South). This may affect results as fracture was found at greater depths than the actual depth of the well. Another implication was that wireline logging was performed the day after drilling because the borehole was at risk of collapsing. The wells were drilled using drilling fluid and geophysical logging was performed before the fluids were optically clear and materials in suspension had not yet settled. The result of early wireline logging affected the optical tele viewer, and ultimately produced poor borehole images (Williams and Johnson, 2000).
Some of the GPR results were difficult to interpret due to poor resolution, and the wavy reflections obtained from radar facies. It is difficult to explain the results, but a possible reason may be related to the foreign objects found in the soil, and the presence of underground culverts and electrical lines buried in the sediment layer that may disturb reflections and also display hyperbolic reflections (Birkenfeld, 2010, Tong, 1993). Another apparent limitation in this study was the process of evaluating the precise depth from the two-way travel time in nanoseconds (ns). Converting to depth is dependent on the GPR-signals propagation velocity (0.9 m/ns) and time to reflective signals: e.g. 100 ns ~ 4.5 m depth. The propagation velocity of 0.9m/ns is a rough estimate only, and variations of + - 20-30 % is possible. The indicated depths have a corresponding uncertainty (Eiken, 2019). GPR profiles were recorded from different direction across the park, it can thus be suggested that GPR profiles should have been completed perpendicular to the strike of the bedding planes (NE-SW) in order to obtained stronger results (Sensors & Software, 1992-1999).

6.1.2 Hydraulic properties of unconsolidated sediment

Grain size distribution from current findings ranges from sandy, silty loam, silty clay loam to clayey loam. The unconsolidated sediment layer varies in thickness between the northern and southern part of Torshovdalen. Table 22 demonstrate the average soil texture found at depths in Torshovdalen sub-catchment. A general trend indicates that the upper 0.30 m in the northern half of the park display a silty loam texture. The lower half of Torshovdalen demonstrate a sandy loam in the upper soil horizon between 0-4 m, subsequently fining downwards from loam to silty clay. The bedrock boundary, as seen in figure Figure 3-2, demonstrate the separation between the upper and the lower parts of Torshovdalen.

Well 2 (East) exhibited a silty clay loam from 0.10-2.20 m depth. From these results it is clear that the expected permeability of the soil is generally low and decreasing with depth at well 2. The homogenous clayey content was an unexpected finding, but it is likely that similar circumstances exits at other places in the park as well. This observation may be supported by the compacted marine deposits (Olsen et al., 2013).

This study proposes the possibility of perched aquifers. In accordance with present results, the observed correlation between $K_{sat}$ values and soil texture profiles suggests that the silty clay layer, is creating a divide between the upper silty loam layers, and a semi-confined aquifer in the fractured bedrock. These impermeable lenses may contribute to a sudden increase in water content that will fully saturate the above soil and force water to move laterally. Thus making
water seep through the surface and ultimately creating a seepage which also may be a reason for the springs found in Torshovdalen (Fetter, 2001). The theory is supported by the subsurface investigative results and the presence of concentrated clay content creating a perched aquifer. These clay layer may also function as a barrier for water percolating downwards attenuating the process (Reddi and Inyang, 2000). This is true for the southern part and the area by the ephemeral springs.

Sodium chloride found in sediment samples collected at well 3 (South) displayed a small NaCl content, however, it is important to note that these samples were from 4-6 m depth, which is considerable far from the source. Limitations apply to the method of quantitatively measure NaCl content. The experiment is not suitable for real NaCl content as all ions that contributes to conductivity are accounted for. High concentrations total dissolved contents (TDS) will affect results if NaCl content is calculated from conductivity. However, Calculations of NaCl equivalents yield acceptable results for geotechnical evaluations of marine sediments. Another limitation is that samples are from disturbed and handled sediment sample and will not yield in-situ conditions.

Results from grain size distribution was used to estimate the saturated hydraulic conductivity (K_{sat}) by comparing empirical formulas Hazen’s method and Puckett 1985. It must be pointed out that the majority of the samples had an effective grainsize of 0.002 mm which complicated the use of Hazen’s method, however, effective grainsize of the 10^{th} percentile (D_{10}) used by Hazen was found by estimation (Eq 4-10 and Eq 4-11). Results show that K_{sat} values obtained with Puckett 1985 are 2/3 orders of magnitude higher than those obtained with Hazens formula. Subsequently, regarding K_{sat} below 4 m depth, findings clearly indicate a transition from moderate to low permeability as the silty clay layer is reached. The observed differences when using both empirical formulas suggests that Puckett 1985 is more suitable for samples dominated with finer particles. A limitations to these calculations is that different literature suggests different permeability coefficients when using Hazens formula, also literature suggests that this formula may not be used for anything other than sandy sediment samples (Fetter, 2001, Koenig, 1911). This may explain why Hazens formula was greatly deviating from Puckett 1985.
Table 22: Typical sediment facies found throughout Torshovdalen catchment and the corresponding saturated hydraulic conductivity class.

<table>
<thead>
<tr>
<th>Lithology</th>
<th>Zone</th>
<th>$K_{sat}$ class</th>
</tr>
</thead>
<tbody>
<tr>
<td>Anthropogenic Silty loam</td>
<td>Unsaturated zone</td>
<td>Moderate/Moderately rapid</td>
</tr>
<tr>
<td>Loamy/Sandy loam</td>
<td>Saturated zone</td>
<td>Moderately rapid</td>
</tr>
<tr>
<td>Silty loam</td>
<td>Saturated zone</td>
<td>Moderate</td>
</tr>
<tr>
<td>Silty clay, clay</td>
<td>Saturated zone</td>
<td>Very slow or semi impermeable</td>
</tr>
<tr>
<td>Shale/Limestone</td>
<td>Fractured Aquifer</td>
<td>Very slow or semi impermeable</td>
</tr>
</tbody>
</table>

It is important to note that there are several sewer lines extending through the Torshovdalen (Figure 3-8). Common construction methods of these sewer lines indicate that there may be alien soil filling in the trench where the pipes are placed (Ødegaard et al., 2014). This may affect the infiltration capacity and grain size distribution, thus representing a great uncertainty. Although these sewer lines were placed in the late 1800s and early 1900s, the soil profile is still affected by the reworking of the soil.

Collectively, geo-mechanical properties from cylinder samples 1, 2 and 3 from well 3 (South) show evidence of a soft to a medium hard clay with low to medium shear and generally low sensitivity to failure. Shear strength is the maximum value of stress the soil can withstand, implying that shear resistance remains high after failure (Duncan et al., 2014). The characteristics and strength of soil samples containing clay particles are important for stability analysis and is a fundamental requirement for considering slope failure. Low sensitivity indicate that the soil has a high resistance to interparticle movement (Duncan et al., 2014).

Findings may therefore imply that the clay found in the soil/sediment layer is showing ductile behavior and thus relative resistant to frictional displacement with respect to depth. Sensitive clays are common in Norway in which these results are likely to be related, as quick clay was of main concern (Thakur et al., 2006). According to (Nordahl-Olsen, 1993) avalanches of quick clay was found in the area, which may be a detrimental factor for urban development planning (Thakur et al., 2006). However, quick clay was not found in soil samples, and is proven by the remolded shear strength that exhibited values > 0.5 kPa. There are several processes that may lead to the reduction of shear strength such as increased pore space, cracks through the soil, and the swelling of clays.
The study confirms that humic substance content is associated with Environmental Protection Agency (2011) statement that low organic content is one of the features defined for urban soils. Cylinder samples from well 3 (South) location and results from testing the humic substance in soil samples from Torshovdalen sub-catchment indicate a low humic content of 2.4-5.3 %. Bjerketvedt et al. (2016) suggested that increase in humic substance will enhance the texture and thus increase the pore space and even though the soil has a fine texture, the amount of organic matter will increase the permeability (Bjerketvedt et al., 2016). The low content of organic matter found in the top soil was probably not enough to support this statement as infiltration rates were relatively slow. Organic matter was estimated by the loss on ignition method. Loss on ignition was set to equal the content of humic substance and is denoted in percent of dried sample of only grainsize <500 mm. Meaning, M500 represent % LOI of only material passing through a sieve of mesh 500 mm and is therefore not representative for the entire sample. Because of the limitations when calculating the loss on ignition, we must treat these results with caution

### 6.2 Ground-surface water interaction in Torshovdalen

The unsaturated zone encompasses the upper ~1.7 m above the presumed water level (0.8-2.6 m below surface) in the unconsolidated sediment layer. This value is the mean value from water level data. Water level data were established from interpreting drill logs and the mean water level from the southern well 3. The water table is assumed to be at equal depths throughout the valley. As mentioned previously, Torshovdalen encompasses a great variable depth to bedrock by exhibiting a shallow layer in the northern half and reveals a valley at the southern end by a sudden change in topography as seen in
Figure 6-6. Surface topography of Torshovdalen sub-catchment is also gently sloping on the eastern boundary and has sharply sloping on the western boundary, naturally draining water to the lowermost point.

![Interpretation of transitional interfaces - P51](image)

Figure 6-6: A: Depth to bedrock and depth to the water table were evaluated from GPR profiles. Water level was estimated from ground water level monitoring in well 3(South). B: Interpolated bedrock topography from point elevation. C: Map over location of profile. Figure A show how the bedrock varies throughout the park. Figure B show the 3 dimensional relationship between the bedrock and surface.

Set sampling transects for infiltration tests were established to ensure comparability (Figure 4-1). These transects were first and foremost determined based on observing the snowmelt: Identifying problem areas in the park that were prone to ponding and water accumulation. Sampling transects were planned to intersect these areas from four different locations and also oriented approximately perpendicular to the valley bottom to get a broader perspective. This
was done with the intention to investigate if there is a correlation between water assemblages and soil hydraulic properties, and if there is a difference along each transect. Interestingly, results from $K_{sat}$ estimation from infiltration rates and corresponding empirical formulas from grain size distribution did not validate the hypothesis of a much slower $K_{sat}$ at the ponding locations. This may indicate that not only the hydraulic conductivity of the loose sediments control water ponding and flooding. Other parameters, such as depth to bedrock, being that the vertical permeability of the bedrock is much lower than that of the overlying sediment, underground structures, alien finer sediments at depth, and surface topography leading to uneven surface runoff drainage.

An objective of this project was to identify infiltration rates in the study area, using the modified Philip Dunne (MPD) method. This method was used to determine the soil property which is crucial for a reliable urban water management. It is important to note that this method will not yield information about the deep subsurface infiltration rates and will only be valid for point surface infiltration measurements. Infiltration rates show variable but low $K_{sat}$ values at all locations ranging between 3-30 cm/h. Infiltration was very slow at almost all locations, and several trials was not necessary as saturated conductivity was presumed reached. The observed change between the correction factor of 0.8 and correction factor of 0.6 demonstrated a 14% difference. By compensating for sandy soils, infiltration rates would prove to be much faster.

The difference in the obtained infiltration rates may be due to several factors. Firstly, the location of where the test was performed. Locations with higher compaction tendencies will have a changed soil structure with increased bulk density and low pore space and thus infiltration rates will decrease (Shukla, 2014). Secondly, soil textures are spatially variable and may contain different amounts of silt, sand and clay particles from location to location. Urban soil is known for its great variability in soil texture and is therefore important to investigate (Craul, 1999). This accords with earlier observations between sampling locations. As mentioned earlier, well 2 (East) showed high clay content between top to bottom samples, whilst well 3 (South) exhibited an upper layer of sandy loam. This was also specifically observed across transect 3. Other factors influencing the infiltration rates include: The chemical and physical nature of the water, soil moisture conditions, biological activity, temperature, percent of air in pores, atmospheric pressure and type of equipment used (Johnson, 1963).

Common for all infiltration tests was the fast infiltration rate attained at the beginning of each test, indicating a relatively permeable and/or high suction pore space at the shallowest
layers of soil. The rates subsequently slowed down as soil became saturated, leading to a decreased suction, causing the pressure gradient to decrease (Dingman, 2015). Finer sediment samples have micropores in which normally are neglected, however, it may influence water retention properties. Soil sample including both macro- and micropores may have two different fluid pressure fields, indicating that these pores may reach equilibrium at different rates. Also, soil texture is complex to any variation in suction (Koliji et al., 2006). Findings demonstrated that infiltration rates after 1 cm was ranging from 64-540 cm/h between locations. In line with previous studies by Solheim (2017) also demonstrated that infiltration rates at the beginning were fast (0-200 cm/h) and later slowing down. In view of infiltration rates investigated by (Kristiansen, 2019, Storteig, 2019, Solheim, 2017), it is a clear similarity between the obtained infiltration rates: All rates display $K_{sat}$ rate varying between 2-96 cm/h. From this standpoint, the results provided by aforementioned literature can be considered true values for parks and recreational sites in Oslo containing the same anthropogenic soil in the upper soil horizon. Oslo is covered in marine deposits and is somewhat homogenous in content of silt, clay, sand, till and glaciofluvial sediments (Olsen et al., 2013).

Figure 6-7 illustrates the difference between empirically calculated and observed $K_{sat}$ values from infiltration testing. Interestingly, by comparing results from calculated and observed values, there is a significant difference in saturated hydraulic conductivities. Figure 6-7 illustrates how empirical methods and the observed infiltration rates differ. From the empirical formulas, Puckett 1985 produced a more realistic $k_{sat}$ value (0.8-5.1 cm/h) in that it was closer to the observed $K_{sat}$ values measured from MPD method. Hazen’s method demonstrated a great deviation from the other values with a very slow $K_{sat}$ range of 0.002-0.007 cm/h. It is important to note that soil samples were collected at depth and infiltrations tests were completed at the surface. This raises the assumptions that the silty loam in the subsurface is much more impervious than the anthropogenic soil in the upper 0.30 m as seen in Table 22. Alternatively, it can also mean that the top soil contains a great amount of humus, enough to increase the pore space to yield high infiltration rates.
Figure 6-7: Infiltration testing compared to empirical $K_{sat}$ estimations from Hazen and Puckett 1985.

Soil moisture was measured at 10 cm depth for every 2 m along the transects throughout the summer months and early fall of 2018 to capture the seasonal changes in the soil. The measurements demonstrated an expected lowered soil moisture value during the summer months, and high soil moisture during the rainy fall months. On the contrary, transect 3 demonstrated a constant high soil moisture value throughout summer between (40.7-63.8 %) and fall between (43.5-63.2 %) at the points where it intersected the three ephemeral springs.

Interestingly, the location of the ephemeral springs showed high values, even though no visible water was seen. Overall, these findings are in accordance with soil moisture field campaign days as seen in Figure 6-8 that illustrate soil moisture tests in conjunction with precipitation and temperature. Soil moisture tests were sometimes preformed on days with no precipitation, and soil moisture results would still demonstrate high moisture at spring locations. This may further the hypothesis that the springs are event-driven springs and abrupt changes in topography may initiate the hillside spring at high flow events, however, continuous groundwater flow will keep the area moist during low flow events.

Observation has demonstrated that antecedent moisture conditions and the underlying bedrock may be factors governing the mechanism for these springs. One limitation of the soil
moisture measurements was that the Delta-T moisture probe could not always penetrate through the dry and hard land surface which was a result of the intense dry summer. These points are represented as zero values in (Figure 5-41).

![Figure 6-8: Mean precipitation and temperature is presented from February 1st 2018 to November 28th 2018. Soil moisture field campaigne dates are shown in black.](image)

Another hypothesis may be that the springs located in the valley are due to leakage from a sewer line located within proximity (Figure 3-8). In accordance with results from groundwater quality in the same study area, that indicated high concentrations of Escherichia Coli (E.coli) during dry days (3705.2 E-coli/100ml) and low values after a rain event (Kristiansen, 2019). Stocker et al. (2015) showed that the population of E. coli generally decreased with the increase water input, Kristiansen (2019) suggested that because of this change in E-coli concentration, the dominant source of E-coli in the springs may be from animal manures and not sewer leakage.

### 6.2.1 Ponding and water accumulation

Torshovbekken catchment, will theoretically not infiltrate much water as ~70% of the catchment is covered by urban pavement and buildings (NVE, 2019). Groundwater recharge into Torshovdalen sub-catchment may therefore occur from water infiltrating in the upper part of the catchment, from the forested areas and scattered green parks. Results obtained from observation, grain size and infiltration tests suggest that there is a high runoff and low infiltration potential in Torshovbekken and Torshovdalen sub-catchment.
In continuation with GPR results, the possible correlation between depth to bedrock and ponding was investigated. This study found an interesting correlation at ponding location A2 where a shelf-like obstacle in the bedrock topography was assessed. This phenomenon might explain the water accumulations at A2 location (Figure 5-38 March 25th 2018 was Sunny, with temperatures of low -0.9 and high 10 degrees °C. Pictures show Area 2 (A2) where ponding is eminent.). According to this observation, we can infer that the reason why the water is accumulating here is because of the topographic depressions detected. The study was not successful as it was not able to determine if the bedrock topography was the reason for ponding, however, it is possible that the depression in the bedrock are retaining the water, and as precipitation continues, this will fill up and surface ponding will occur. Topographic depressions may hinder overland flow and result in ponding even though terrain is sloping (Craul, 1999). Infiltration rates were compared to literature to infer a correlation. At Area A2, infiltration rate from L1-T2 was 2 cm/h and 6 cm/h from L2-T2, comparatively, important differences was observed in infiltration testing done by Kristiansen (2019) which showed a higher $K_{sat}$ at A2 location of 77 cm/h. At location A1, infiltration rates were measured as 22.5 cm/h.

Findings suggests that area A2 may be exceeding the infiltration capacity during rain events and flood. The area experiences a drop in hydraulic head, attenuating the specific yield/unconfined Storativity, creating more water draining into the ground because of gravity rather than the horizontal movement of water. The volume of water released from storage per unit hydraulic head show an increase in the specific storage due to the low hydraulic gradient. (Cushman and Tartakovsky, 2017). With no overlying confining layer, porosity is thus important and storage is displaying higher specific retention and low specific yield (Freeze and Cherry, 1979, Heath, 1983). Several questions remain unanswered as to why the water accumulates at the aforementioned areas. However, high silt and clay content, decrease in topographic gradient and low aquifer thickness could be a major factor causing ponding.

Likelihood of frozen ground is determined based on soil textural properties, water content, weather, climate and the frost action process. The frost action process is the downward advance of the freezing front. This may occur as the heat extraction exceeds the heat supply (water flow). Active frost lens consists of three zones: Frozen soil, a freezing fringe and underlying unfrozen soil layer. Soil within the fringe will expand when the pressure in the ice exceeds the overburden.
pressure. With sufficient ice pressure the soil skeleton separates and a new ice lens is formed (Andersland and Ladanyi, 1994).

High clay content is proportional to the risk of frost developing. Predicting soil frost depth will determine the risk of frost reaching the lower layers of higher clay content which in turn will significantly affect the infiltration capacity (DeWalle and Rango, 2008). The present evidence of sediment facies, seen in Table 22, demonstrated an increase in clay and silt content with depth and thus increasing the risk of frozen ground in the area. This was especially seen in cylinder samples from 5-6 m depth where clay fraction was 47 and 44% respectively. Oztas and Fayetorbay (2003) suggests that during the colder seasons with risk of temperature below freezing, infiltration rates will decrease due to the reduction in permeability in an already low permeability layer. Thus, development of frost that may creep further into the ground may become an implication for meltwater infiltration and may cause ponding and surface runoff challenges.

The development of frozen ground is only possible if soil was frozen prior to snowfall as snow exhibit insulating properties. Water moves upwards due to pressure gradients and freezes once frost zone is reached. This process may create a thicker frost zone (DeWalle and Rango, 2008). The risk of frozen soil was evaluated from grain size distribution and grouped in different groups according to its probability following a classification system developed by Johnsrud et al. (2013)(Appendix G.e). Collectively, all samples are categorized to be of little to no risk (frost group T2/T3). On the contrary, in the lower layer in well 3 (South) between 4-6 m, the grain size distribution results suggest that there is a high frost susceptibility (T4). This theory further supports the idea that the lower clay layers may be affected if there is a downward advance of the freezing front. Lenses of clay and silt alternating downwards, could possibly enforce the downward movement of frost (DeWalle and Rango, 2008).

Climate causes ground surface temperature to vary cyclically from year to year. It is suggested by Andersland and Ladanyi (1994) that between 9-15 m depth the soil keeps a steady temperature throughout the year. Below this depth, temperature will increase linearly for a homogenous material. Temperature increase is due to heat generation in the central core of the earth (Andersland and Ladanyi, 1994). In northern countries, temperatures will increase at an interval of 20 °C per kilometer, while other part of the world might experience an interval of 40 °C per kilometer (Skoglund).
6.3 Weather, climate and the soil water balance of Torshovdalen.

The hydrogeological year of 2017/18 in Oslo demonstrated great seasonal contrasts compared to mean normal values for the thirty years period of 1931-1960 and 1961-1990 (Eklima). Heavy long-lasting winter, a late cold spring, and an abnormal temperature increase that lasted throughout the summer and affected the hydrological cycle of 2018 (Skaland et al., 2019). Considering the year of 2017/18, total precipitation was 642 mm/yr., which is significantly lower compared to the mean normal values over a thirty period (1961-1990) which showed 763mm/yr. Mean temperature for the period May through July was record high and demonstrated 3.1°C more than mean normal of 1961-1990. In which became the warmest period (may-July) since 1900 (Skaland et al., 2019). This Period was expressed by low precipitation rates and was estimated to be 26% less than mean normal values (1961-1990). East of Norway had ~200 % more precipitation than the normal during the winter season of 2018 (Grinde and Mamen, 2018). Figure 6-9 show the precipitation (% difference from the normal) and temperature (deviations from the mean) between the months of May-July.

![Figure 6-9: Precipitation (% Normal mean) and temperature (deviations from the mean) during the summer months between May and July as demonstrated from the eastern region of Norway. Modified from (Skaland et al., 2019).](image)

The months from March to July 2018 was characterized by high monthly temperatures and low monthly precipitation, ranging from 7-22°C and 15-27 mm/month respectively. On the contrary, in 2016/17, the temperature and precipitation was ranges from 3-17 °C and 40-100 mm/month. This observation indicates an average increase of temperature by 4°C/month and
an average decrease in precipitation by 53 mm/month for the hydrological year of 2017/18. The great difference between the two consecutive years suggests that 2016/17 was a much wetter year compared to 2017/18.

Comparisons have been made between water balances from consecutive hydrological years (2016/17 and 2017/18), and a fictive hydrological year affected by predicted climate change. Figure 6-10 illustrates important soil water balance parameters and how they differ between the aforementioned time periods. In the current study, a climate change water balance was implemented using the assumptions brought forth by National Research Council (2011) that in conjunction with a temperature increase of 1-4°C, precipitation will increase with 5-10% in high latitude areas. Therefore, a water balance was constructed with an annual increase of precipitation and temperature of 10% and 5°C respectively. Since 2017/18 was an abnormal year, 2016/17 was used as a template for the possible climatic changes. As seen in Figure 6-10, estimations on climate change resulted in a higher amount of precipitation, however, less snowmelt was observed implying less water available for runoff (Figure 6-12). The year of 2016/17 was used as a template for climate change

![Water Balance Parameters](image)

*Figure 6-10: Comparing water balances for two consecutive hydrological years, and one fictive hydrological year affected by predicted climate changes. Here, annual precipitation, potential evapotranspiration (PET), infiltration and surface runoff are represented.*

Observing snowmelt was the initial assessment of the behavior of water in Torshovdalen and seasonal changes are documented in Figure 6-12 and Figure 6-15. The long cold winter followed by a late spring with low humidity which demonstrated a dominant sublimation process as seen in Figure 6-11. In April, air pressure was greater than vapor pressure at surface and melting occurred. This is proven by the great volume of water as shown in Figure 6-12.
As observed in Figure 6-11, decrease in humidity and low vapor pressure explain the dominant evaporation processes, subsequently, increase in humidity and vapor pressure initiated condensation (Mockus et al., 2004).

![Vapor pressure and humidity](image)

*Figure 6-11: Vapor pressure (yellow) and air humidity (black). changes in air vapor pressure between a thirty-year period (1961-1990) and the hydrological year of this project (2017/2018).*

Vapor pressure is a function of temperature, and the snowpack would be expected to release water when the surface temperature of the snow reached 0°C and 6.11 hPa (DeWalle and Rango, 2008). Temperature data collected during field campaigns shows that temperatures changed from -2°C mean monthly temperature in March to, +7°C mean monthly temperature in April (Figure 6-13). The water balance suggests that potential evapotranspiration is increasing from late March implying that snowmelt water will be less affected by the increasing vapor pressure. The water table was also theoretically affected by the infiltrating water especially during the snowmelt season.

Water table is graphically presented in Figure 6-12. According to Price (1996) this difference may affect the amount of aquifer recharge to the area, and ultimately affect the flow in the unconsolidated aquifer. One might argue that year 2017/18 would experience a lowering of the water table because of the observed water deficit, however, considering the silty soil texture found locally in Torshovdalen, water will most likely fill the pore space in between grains, raising the water table especially after the snowmelt recharge episode.

According to de Beer (2018) groundwater level is significantly lowered throughout the country in the past 70 year, and is continuously lowered. NVE and NGU has received raports from more than 80 water wells from the tool “Landsomfattende mark-grunnvannsnettet” (LGN)
which is run by NVE in cooperation with NGU (de Beer, 2018). Base flow may be more than 80 percent of water in smaller sub-catchments in the winter periods (Wong and Colleuille, 2005). Thus knowledge of groundwater contribution to the total discharge is important for the water resource and to quantify the groundwater contribution (Wong and Colleuille, 2005).

Figure 6-12: Water table fluctuations from the hydrological year of 2017/18 (grey), 2016/17 (orange) and climate change (CH) (blue).

Figure 6-13: Temperature comparison between hydrological years: 2017/18, 2016/17 and mean normal values from 1961-1990.
March 25\textsuperscript{th} signified an output phase for the snow melt as the snowpack couldn’t retain more water and melting occurred Figure 6-14 which is supported by the observations on infiltration capacity, soil moisture and vaporization. Theoretically, the free water observed after snowmelt are more likely to contribute to surface runoff than being affected by evaporation and transpiration (Dingman, 2015). However, to which extent this applies to this area is arguable. Findings are supported by the evidence of water accumulating at designated areas (A1, A2, and A3).

April 14\textsuperscript{th} signified a day of decreasing humidity, however, increasing vapor pressure Figure 6-11. This imply that evaporation may have been the dominant process (Mockus et al., 2004). April 20\textsuperscript{th} marked the day where no snow was observed. Estimations for the water balance in Torshovdalen demonstrate the occurrence of the snow melting in the months between March and April. Even though snowmelt was apparent, more water was expected to be seen on the ground surface. As can be seen in Figure 6-8, great amounts of snow (max ~52 cm) was accumulated during the winter, which is equivalent to ~156 kg/m\textsuperscript{2} water. It is important to note that, estimating snow density and snow water equivalent is based on variable parameters both in space and time, thus signifying only a rough estimate.
Figure 6-15: Left: April 14th (2018). Middle: April 20th (2018). Images show spring melt progressively changing from snow cover to no snow cover over the course of 6 days. Map to the right exhibits the location where the images were taken.

Figure 6-16: Comparing estimated amounts of snow water equivalents (SWE) and snowmelt for three hydrologic scenarios.

Comparisons were made between different hydrological years and are presented graphically in Figure 6-17. Estimations provided by NVE (2019) demonstrated an estimated runoff in the study area to be 17.6 l/s*km². This estimation is based on an automatic generated value by using runoff indexes and field parameters and considering the size of the catchment. Current study confirms this assessment as own calculations reveal a mean runoff of 19 l/s*km². The present study was considered to determine the effect of urbanization and the behavior of water. Torshovbekken catchment consists of ~60 % impervious and ~40 % permeable surface areas. The great difference in land cover will thus yield different volumes of runoff compared to Torshovdalen sub-catchment in which is considered to be 100% permeable. Interestingly, even though 2017/18 was considered a dry year, it was estimated to generate ample amount of runoff in both sub-catchments.
However, as previously discussed, the amount of snowmelt generated high volumes of water over a short period which is reflected in current calculations Figure 6-16. These results corroborate the finding from water balance and demonstrate how runoff is reacting to climate change in urban environment (National Research Council, 2011, NOU, 2015). Amount of surface runoff will increase with the increase of impervious surfaces and climate change (NOU, 2015). Surface runoff is a function of precipitation but also a function of land use. As can be seen in Figure 6-17, general conclusion from sub catchment of Torshovbekken (~60% urban) generated more runoff compared to Torshovdalen (100% forested/other).

Precipitation (P) is the governing parameter when conducting water balance calculations, although the potential evapotranspiration (PET) plays an equal part (Pascual-Ferrer and Candela, 2015). Evapotranspiration may arguably be the most influential parameter in Thornthwaits’ water balance method (Dingman, 2015). Evidence provided by Jakimavicius et al. (2013) revealed that Thornthwaits’ method for estimating evapotranspiration gave the most correct predictions (R=0.78-0.96) with corresponding low uncertainties compared to similar temperature based methods. Which provides ample confidence that the Thornthwaite method will yield fairly accurate estimations.

6.4 Stormwater management and risk evaluations

The purpose of estimating a water balance was to evaluate the hydrological responses in Torshovdalen sub-catchment, and forward recommendations for stormwater management solutions. In general, in view of current findings which projected poor infiltration and high...
runoff rates, suggests that excess stormwater is a recurring problem, and will worsen not only with changing climate but also in conjunction with urbanization.

Infiltration rates have been discussed previously and overall findings demonstrated a slow infiltration rate (3-30 cm/h). This will affect the amount of runoff that is generated as most of the water will flow across the surface. This water will follow the natural drainage lines and eventually discharge into the sewer system. There are two significant sewer lines following the western edge of the park with adjoining manholes. Torshovbekken is running in the outer sewer of the park according to predicted historic watercourse (VAV, n.d) (Figure 3-8). Surface water is generally directed into the sewer and away from rooftops, streets and open spaces (Ødegaard et al., 2014). The capacity of a sewer pipe is thus imperative for calculating the amount of water entering and leaving the system. The sewer is leading into a larger tunnel that will dispose of the waste-waters.

Stormwater management actions that target a sustainable runoff and drainage systems such as rain gardens, retention pools and sedimentation basins may help stabilize the groundwater table which in turn promotes good conditions for the surrounding vegetative and biological environment (Wagner et al., 2002). One can consider taking advantage of areas that are noticeably accumulating water to implement aforementioned actions and even daylight the buried Torshovbekken through these points. It is important to bear in mind that these measures should not be treated as natural bodies of water but storm water management facilities, due to the poor quality of the surrounding groundwater and overland flow that carries pollutants as shown in results by (Kristiansen, 2019). Groundwater in direct contact with the surface water may be easily polluted due to the water surface interaction in the city (Wagner et al., 2002). Collecting stormwater should be a priority as runoff has proven to be highly available in the study area and infiltration capacity is low.

Rehabilitating Torshovbekken will manage runoff to protect infrastructure Bernhardt and Palmer (2007) and should be considered as an option as relatively clean stormwater will be transported away relatively quickly. Even though current findings imply a low infiltration rate, the volumetric flow produced by a stream will help carry water further until it discharges in the Oslo fjord. A stream will relieve combined sewer and treatment plants from storm flow and hydraulic loads and improve the water quality by exposing the water to air, daylight, vegetation and soil that will defuse pollutants (Pinkham, 2000). Implying that less polluted rainwater is not discharged into the sewer but redirected to the nearest stream. Also, reducing the risk of
infrastructural damages due to the attenuated flow from the constructed river (Wagner et al., 2002). Considering stream morphology, establishing a river may result in implications such as stream erosion and sediment transport as geomorphic processes are closely related to water movement (FISRWG, 2001). Establishing a flood plain when restoring stream corridors will help capture water that exceeds the stream's capacity (Wagner et al., 2002). In accordance with stormwater management and the implementations of variable stormwater solutions it was suggested by Kristiansen (2019) that because of the poor quality of water, consuming and bathing would not be advised if said practice was implemented in Torshovdalen.

Oslo Kommune (2015) offered in the stormwater management document “principles for reopening rivers” options for different types of streams to be rehabilitated. Two stream types were suggested: Nature like water course and channelized watercourse. Natural watercourse will communicate with the groundwater through the riverbed and will greatly affect the rise and lowering of the water table. Oslo municipality discourages the channelized watercourse, except when located closely to roads and areas with high urbanization degree. However, implementing a combined stream with both natural and channelized may be a good option to hinder erosion (Figure 6-18). Oslo Kommune (2015) also encourages communication between the stream and the groundwater to secure base flow. The observations made in this study suggests that stormwater should be retained in sedimentation basins or retentions pools especially at locations.
of ponding and accumulation. Implementing retention pools to collect stormwater, will result in less stormwater occupying the sewer. This in turn, will relieve the pipes from exceeding its capacity. The result will be less leakage and flooding of waste water from the sewer and into the surrounding soil. Implementing stormwater solutions in Torshovdalen, such as retention pools and natural streams, will then theoretically result in less polluted bodies of water due to the decrease in waste water entering the soil, if said stormwater is directed away from the sewer.

Road salts and anti-icing commonly used in the winter is a great contributor to salts dissolving into the ground (Li et al., 2014, Kristiansen, 2019, Amundsen et al., 2012). The degradation of the de-icing agents may produce an anoxic environment in the soil, which may increase the transport of heavy metals such as iron and manganese as these metals are mobile in reduced environment (Amundsen et al., 2012). The dissolution of calcium and magnesium may lead to a colloidal transport of fluid through the soil, which may lead to a decrease in pore space and thus decrease hydraulic conductivity. The pollution of road salt and the addition of Sodium Chloride (NaCl) may lead to decreased soil stability and fertility, however, the impact is decreasing with increased distance from the source (Hofman et al., 2012). Considering that Torshovdalen is mostly surrounded by roads one might argue that road salt will still have a considerable effect on the soil environment. On the other hand, anthropogenic sources such as animal manure and sewer leakage may also be an important source of contamination. Study executed by Kristiansen (2019), concluded that sources of contamination, mineral dissolution and weathering processes will affect the infiltrating water and ultimately affect the groundwater.
7 Conclusion

The site assessment has identified the soil textural properties of Torshovdalen. The study has shown that the soil contains an overall silty loam soil in 0-4 m, and a silty clay type soil in the lower soil horizon. Torshovdalen demonstrates a restricted drainage as infiltration measurements at all sample location showed a low infiltration rate of 3-30 cm/h. Hydraulic conductivity was estimated from soil textural properties comparing two empirical formulas from Hazen and Puckett 1985. Mean $K_{sat}$ values using Hazen method was 0.007-0.02 cm/h and 0.4-5.1 cm/h using Puckett 1985. Furthermore, infiltration testing showed much higher $K_{sat}$ of 3-30 cm/h. Estimation suggests a variable $K_{sat}$ rate implying a moderate to very slow $K_{sat}$ classification. Puckett 1985 proved to be a better choice when estimating $K_{sat}$.

The soil contains a higher percentage of humic substances which demonstrates a higher water retention capacity compared to the lower layers. Humic substance and soil moisture is decreasing with depth. The high clay content soil in the lower soil horizon, and the abrupt changes in soil texture determines the vertical permeability and may cause perched water above these interfaces. Soil moisture values indicated a constant high-water content in areas of seepage in the park and these locations are situated were bedrock topography changes due to possible uplift from normal faulting. The leakage of water through these points may be due to the abrupt change in bedrock topography.

Results from GPR profiles suggests a shallow depth to bedrock ~1-3 m in an undulating bedrock surface, however in the southern end of the park, depth to bedrock increased dramatically ~13 m. Depressions in the undulating bedrock topography may be a reason for water accumulation on the surface. Depressions in the bedrock hinder overland flow even though terrain is sloping. Capture and retain water and secure floodways rather than infiltrating stormwater is considered the best stormwater management solution for Torshovdalen.

The present findings confirm that a fracture in the Cambro-Silurian shale was detected at well 3 location with an aperture of 0.13 m and was highly water bearing according to influx estimations. These findings provide insight to the possible fracture network that exists in the bedrock under local over pressured conditions controlled by the low vertical permeability of the overlying sediment cover. Water level fluctuations showed an increase in March 2018 and reached its highest level at 0.84 m below surface, implying that the water from spring melting is affecting the groundwater in the fractured rock.

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The study contributes to the understanding of the water balance, based on Thornthwaite method. The study has estimated annual runoff values for 2018 for Torshovbekken catchment to be 19 l/s*km2 and 3 l/s*km2 for Torshovdalen sub-catchment. Estimation of change in water storage for Torshovdalen sub-catchment for was 4.4 l/s*km2 2016/17 and 5 l/s*km2 for 2017/2018. Including climate change factor, change in storage was 3 l/s*km2 of water available in the aquifer in Torshovdalen. It is suggested that with increased temperature and precipitation, more available water will affect the urban hydrological cycle Oslo. Despite its exploratory nature, this study offers some insight into spring snow melting season which was dominated by sublimation in the year of 2017/18, which caused the snow to ablate without much snowmelt as air water vapor pressure was lower than the constant vapor pressure at snow surface of 6.11 hPa. The snow cover was non-exiting by April 21st, and little to no surface water was accumulating in problem areas.

The empirical findings in this study provide an understanding of the groundwater flow in the unconsolidated aquifer from Torshovbekken into Torshovdalen. Flow in to the sub-catchment of Torshovdalen was calculated to be 0.0001 m³/s, and subsequently, flow out of was 0.0003 m³/s. Therefore, it seems that a greater volumetric flow is exiting Torshovdalen catchment compared to what is coming in. One of the more significant findings to emerge from this study was the fracture found in well 3 that showed proof of a groundwater bearing fracture at 18 m depth with an aperture of 0.13m. Water yield rate in the well was ~6 m/s, which constitutes a volumetric flow of ~2 m³/s.

The purpose of the current study was to determine the groundwater recharge and flow in Torshovdalen sub-catchment. Proper hydrogeological assessment and the estimation of the soil water balance of Torshovdalen sub/catchment were useful tools in determining the hydrogeology of the area. Geotechnical and geophysical investigation, infiltration testing, and periodic field observation were performed in Torshovdalen to determine fundamental characteristics of the underground heterogeneity and the hydraulic properties of the soil, and the seasonality of the groundwater table and the soil moisture content. These findings provide valuable insights for the stormwater management plan in Oslo.
7.1 Further work

- The study should be repeated using an E.R.T measurement to investigate the subsurface. The water flow in the unsaturated zone would be an interesting factor to implement.

- Quantitative evaluations are needed to correctly estimate how much water is actually entering the sub-catchment and consider to which extent the subsurface is affected by the sanitary sewer. Using urban water management tools such as SWMM and Mike Urban to model and monitor the area.

- Continued efforts are needed to make the water balance more accurate. Closer monitoring of water balance parameters such as evaporation and snowmelting. In addition, more information about the water flux in both unconsolidated sediment aquifer and the fractured bedrock aquifer. Investigate the fractured bedrock and the actual water yield by repairing well 2 (East) and well 1 (North) and deploy more divers to monitor the groundwater. In addition, implement monitoring wells in the unconsolidated aquifer.

- More information would help us to establish a greater degree of accuracy on infiltration capacity. Great variation in infiltration rates suggests that more tests are necessary to get an accurate rate. Correlating infiltration deeper in the ground with point infiltration at surface will give a contrast in how the water is infiltrating differently at different depths. Using the Modified Philip Dunne infiltrometer (MPD) deeper into the ground, will help determine different permeabilities with depth. Investigate the fractured bedrock and the actual water yield.
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9 Appendicies

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A. Geochemical properties of undisturbed sediment samples

Table 23: Results from Undisturbed and remolded shear strength test using cone penetrometer test. Average Cone penetration depth (i). Fall cone undrained shear strength of undisturbed soil (Cufc). Fall cone undrained shear strength of remolded soil (Curfc). Results are highlighted in grey for undisturbed and remolded samples denoted (U) and (R) respectively.

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Table 24: Sensitivity reference index. Sensitivity (St) is measured in kPa. (Modified from vegdirektoratet 2016).

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<th>Sensitivity (St) (kPa)</th>
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<td>Medium</td>
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<tr>
<td>Very Sensitive</td>
<td>High</td>
<td>&gt; 30</td>
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Table 1: Classification of hardness of clay. Modified from (NGI, 2019).

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<td>Low</td>
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B. Sample preparation

Table 25: Table contains calculation from heating treatment on sediment samples. Table includes following parameters: Loss of water (%), loss of organic material (550), and loss of volatiles (1100), all values given in percentage. Values are calculated from a small sample of weight between 1 to 2 g. Sample weight is given with 5 significant figures.

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<th>Sample</th>
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<th>Weight 110</th>
<th>%water</th>
<th>550 (kr+pr)</th>
<th>OM</th>
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<th>1100 (kr+pr)</th>
<th>Post-weight</th>
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### C. GPR profiles

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C.a. Manual interpretations of GPR profiles

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Figure 8-1: P53

Figure 8-2: P54

Figure 8-3: P50

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Figure 8-4: P55

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Figure 8-5: P004
D. Mineralogy

D.a. XRD raw data

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Other (Corundum, Sillimanite, Calcite, Ti, Fe)
### B1_F1.5

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### B1_F4

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### B1_F1

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### B1_L1.5

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**11.10.2018**

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*Figure 8-6: XRF raw data part 2*
D.c. Clay classification

B3_L4

*Figure 8-7* Clay classification of sample B3_L4, supplementary evaluation of sample.

B2_L1

*Figure 8-8* Clay classification of sample B2_L1, supplementary evaluation of sample.
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Figure 8-9: Purging of well 1, 2 and 3, summer 2018

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G.e.  Saturated hydraulic conductivity

Table 27: Soil textural classes and related saturated hydraulic conductivity. modified from (NRCS, n.d-b)

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<td>Very rapid</td>
<td>1x10^-2</td>
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<td>Coarse</td>
<td>Sandy</td>
<td>Rapid</td>
<td>1x10^-2 - 4x10^-3</td>
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<td>Moderately Coarse</td>
<td>Loamy</td>
<td>Moderately rapid</td>
<td>4x10^-3 - 1x10^-3</td>
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<tr>
<td>very fine sandy loam, loam, silt loam, silt</td>
<td>Medium</td>
<td>Loamy</td>
<td>Moderate</td>
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<td>Fine and very fine</td>
<td>Clayey</td>
<td>Slow</td>
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## G.f. Frozen soil

Table 28: Frost Group classification of sediment samples.

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Figure 8-10: Grainsize distribution compared to the risk of frozen soil. The y-axis show the cumulative grainsize distribution (%), and the x-axis show the grain size (mm) (Johnsrud et al., 2013).
Table 29: Frozen ground risk parameters. Modified from (Johnsrud et al., 2013).

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<td><strong>Little risk</strong></td>
<td>T2 Gravel, sand and marine deposits</td>
</tr>
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<td><strong>Medium risk</strong></td>
<td>T3 Gravel, sand and marine deposits *Clay</td>
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<td><strong>High risk</strong></td>
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G.g.  Ksat calculations

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H. Soil water balance calculations

H.a. Soil group

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**Infiltration bekken**

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I.a. May 23rd 2018

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