Soil physical characteristics, hydrological flow paths and the water balance

An assessment for a small agricultural catchment in Norway, as simulated by DRAINMOD

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Master thesis Irrigation and Water Engineering submitted in partial fulfillment of the degree of Master of Science in International Land and Water Management at Wageningen University, the Netherlands

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Abstract

Predicting the hydrology of a catchment is challenging. Direct measurements of soil physical parameters are time consuming and assumptions are made due to the spatial variation. Other parameters such as the evapotranspiration and (sub-)surface runoff are also difficult to measure. This thesis focused on a small agricultural dominated catchment in southern Norway, containing 3 main soil types ranging from heavy clay to sand and 2 land uses, namely agriculture and forest. The existing combinations of land use and soil types each have different effects on the hydrology. Currently it is unknown what the distribution of surface and subsurface runoff is, and how much the evapotranspiration is. In trying to understand these parameters and water distributions the hydrological model DRAINMOD has been used in this thesis. Simulations have been performed with different values for the surface storage, drainage intensity, drain spacing, saturated lateral hydraulic conductivity, various soil temperature thresholds and evapotranspiration parameters. Of these parameters, the drain spacing and the lateral hydraulic conductivity for the layer containing the drainage pipe, in this case the fourth layer, influence the water balance the most. For most of the simulations the overall evapotranspiration was 40% of the total precipitation. The main goal of this thesis was to assess if DRAINMOD is suitable for Norwegian catchments. Based on the results of simulations on Skuterud, DRAINMOD is not the best option. This is because the model does not simulate enough surface runoff, considering that the catchment is hilly. The surface storage was adjusted to compensate for this, but with no success. Further research with separate surface runoff and drainage observations will give better insight in the hydrology of Norwegian catchments.

Keywords: Norway, Skuterud, DRAINMOD, drainage, hydrology, soil physical parameters, evapotranspiration, surface runoff, JOVA, IRIDA

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1 | Introduction & background

Climate change scenarios predict an increase in both temperature and precipitation and more frequent rainfall episodes with an increased intensity for Norway. Hanssen-Bauer et al. (2009) mentions winters will become less severe and the number of freezing and thawing cycles might increase, resulting in less stable soil aggregates (groups of soil particles) and an increase in erosion (Kværnø and Øygarden, 2006). However, an increase in temperature can also mean a longer growing season (Skaugen and Tveito, 2004). Analyses have been done about the potential effects of changing climate conditions in Norway. The predicted increase in temperature in Southeast Norway is 3.4 °C for the period 2071-2100, compared to the years 1961- 1990, with 4.5 °C in the winter season and 2.5°C in the summer season. The predicted increase in precipitation is 12.2 %, compared to the same period of 1961-1990. However, it is not evenly spread throughout the year. The summer precipitation will decrease slightly (-4.4%), whilst increasing in autumn (15.1%), winter (28.9%) and spring (14.0%) (Kelman et al., 2011; Deelstra, 2015). The use of models is important as our understanding and knowledge of the hydrological processes can be improved, which are driven by these climatological changes. This thesis uses the hydrology model DRAINMOD to quantify drainage and other water management systems (Skaggs, 1980). It is a widely used model, mostly in northern America and China, and less in Europe and other parts of the world. However some work has been done in Finland, Sweden and Norway.

Norwegian agriculture is highly influenced by the winter season. If temperatures remain below zero and snow accumulation occurs longer than usual, the sowing of crops is delayed, reducing the overall growing period and thus the yield. Drainage during spring can drain the excess soil water and increase the growing period length. More intense drainage have been found to lead to earlier sowing times. Due to the drainage system, less water is present in the soil which means less ice formation during the winter period, which increases the infiltration capacity of the soil. Also, when ice is present in the soil it needs to be melted by increasing the soil temperature, which takes time and thus decreases the length of the growing season. This fast drawdown of groundwater is needed in spring and autumn for farmers to be able manage their fields, but also during periods of excess rainfall. To be able to deal with the effects of the changing climate a good understanding of the behavior of the drainage system is needed.

The runoff during winter periods in cold climates has been studied by Deelstra et al. (2009) for different catchments, including Skuterud in Norway. This study found that the runoff and infiltration was influenced by soil frost. Skuterud has been monitored for a long time, and data has been analyzed in various studies. Two previous studies have been done using DRAINMOD to simulate the hydrology of the Skuterud catchment. Deelstra et al. (2010a) simulated the discharge of the Skuterud catchment using DRAINMOD with satisfying results, however mentioned that further analysis needs to be carried out to improve the simulation results, which were both under and overestimated for individual years and seasons. The effects of winter on surface and subsurface runoff needs more attention as well. In their study, the DRAINMOD freezing inputs for the forest remains on the default value, whilst the agricultural simulations are only slightly adjusted from the default. Furthermore, the drainage parameters for the forest need adjustment, especially the drain depth and drainage pipe diameter, which might have to high values. There are no physical drains in the forest, however to simulate the forest a drainage system is assumed to be place. Another point of interest is that only 2 soil types were used, one for agriculture and one for forest. Using more soil types in the simulations might result in more accurate results.

Farkas et al. (2016) simulated the hydrology of Skuterud as well, using DRAINMOD and 4 other models, and had varying results concerning the evapotranspiration (ET) (232 - 390 mm y^{-1} for arable land and 358 - 517 mm y^{-1} for forest). One of the conclusions is that the effect of forest on the water balance of the catchment is important, which in turn means that obtaining information on the hydrological components is important for calibration. They also mention that the period of snow melt is the biggest challenge for simulations, due to the complexity of all the processes. Further more, the period of snow melt is the biggest challenge for simulations, due to the complexity of all the processes. A final conclusion is that the ET most likely approaches the 400 mm per year.

Studies that incorporate forested areas in their simulation usually use the modified DRAINMOD-FOREST, for better PET calculations, using other factors such as leaf area index to calculate the PET, but it remains a difficult parameter to predict (Amatya and Skaggs, 2001; Amatya et al., 2016).

The upscaling from field to catchment scale adds additional issues. A field rarely has a homogeneous distributed soil, a whole catchment even less so. There are often more land uses in a catchment scale than in a field scale, for instance forestry or urban area. Field data collection is one method of gaining information on soil physical parameters, evapotranspiration and surface runoff, but is often costly and time consuming. Models can help with this, by gaining additional insights in the local hydrology, but should not be a substitute for field work. Thus the issue arises of how to incorporate suitable soil physical parameters, from the field, in a model which represents a whole catchment in order to understand the hydrological processes.

1.1 Research questions

Previous modeling with DRAINMOD has been done on Skuterud (Deelstra et al., 2010b; Farkas et al., 2016), however more information about the water balance of different land use and soil types is needed, and how they are influenced by different model parameters. This research will try to fill the knowledge gap of DRAINMOD's suitability by *evaluating temporal and spatial variability in water runoff, from different soil type and vegetation combinations and watershed scale, by modelling the hydrological processes and as such to get detailed knowledge about the water balance components.* There are three major soil types in the agricultural area, and one soil type in the forested area, as will be shown in the following subchapters. Also, only the discharge measurements at the catchment outlet are available for calibration and validation of the model. The main and sub research questions are formulated as:

Is DRAINMOD suitable for predicting the water balance of small, agricultural dominated catchments in Norway?

1. How do changes in DRAINMOD parameters influence the hydrological processes and water balance for the prevailing combinations of soil type and land use combinations at field scale?
2. How can these combinations at field scale be combined to accurately predict the outflow at catchment scale?
3. Can DRAINMOD simulate the water balance over a longer period of time?

1.2 Theoretical background

This chapter will explain certain concepts to help better understand the water balance and its components.

1.2.1 Subsurface drainage

Drainage has different functions for different places on earth. In humid regions, drainage is needed to remove the excess water from the fields, soil or rootzone which creates better aeriation of the soil, higher temperatures and easier workability. In (semi-) arid regions drainage is rather needed for preventing water logging and salinization of the soil, by irrigation (Ritzema, 2016).

Drainage occurs via the surface or subsurface. Subsurface drainage systems are installed when the natural drainage to the underground is too slow to prevent water logging and too high water tables. These systems usually lower the average water table and discharge water from the soil above it, making it drier. The soil acts as a temporary buffer, in which high recharge is transformed into relative slow discharge. Surface runoff would have occurred faster if the soil remained undrained (Oosterbaan, 1994). However, Sloan et al. (2016) found that the occurrence of surface runoff depends on the hydraulic conductivity and surface storage, rather than the presence of drains itself. The same effect can be achieved with a surface drainage system. Surface drainage is not the same as surface runoff. Surface drainage is used when the topsoil has drainage problems, rather than the subsoil, and can be ditches for example. After installation, the waterlogged surface is drained and the infiltration capacity of the soil increases. In both cases drainage flow occurs by gravitational force (Oosterbaan, 1994), rather than with pumps for example.

Whether drainage is required depends on the highest permissible watertable level in the soil before it affects agricultural benefits, the *agricultural* drainage criteria, which depends on region, crop, land use and type and period. On the other hand is the *technical* drainage criteria, which looks at design discharge and drainage coefficient, among other things. Environmental and economic drainage criteria also play a role, but will not be considered in this thesis (Oosterbaan, 1994).

Drainage has been part of agriculture since its beginning. The first subsurface drainage through pipe systems were installed around 1800, made from clay (Watt, 2002). Steam engines excelled the mechanization of this process, and the fuel engine made installation of pipes even quicker. From 1940 onwards plastic pipes were used (Bos and Boers, 1994). Drainage theories were developed, making drainage an engineering science (Hooghoudt (1936, 1940); Dumm (1954); Ernst (1956, 1962); De Zeeuw and Hellinga (1958)). The introduction of the computer made the use of models popular, which can evaluate the theoretical performance of different drainage designs (Bos and Boers, 1994). DRAINMOD uses the theories of Kirkam and Hooghoudt.

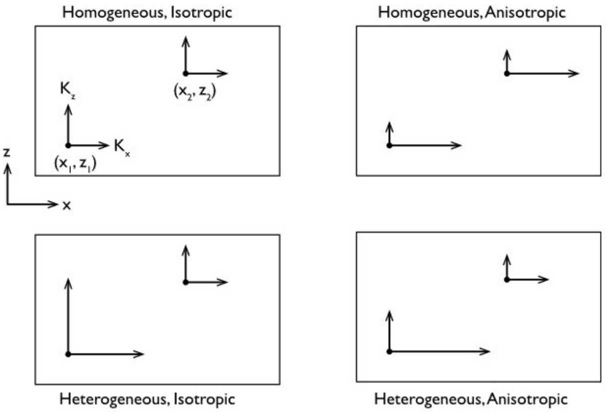

Figure 1.1: Graphic illustration of possible combinations of heterogeneity and anisotropy (Freeze and Cherry, 2017 (Accessed: 13-3-2017).

1.2.2 Hydraulic conductivity and soil anisotropy

The hydraulic conductivity (*k value*) is the velocity of water through the soil pore space (m s^{-1}). There can be considerable spatial variability in the *k value*, thus large scale field experiments work better than small scale experiments, as they can better incorporate the heterogeneity of the soil than small experiments (Braun and Kruijne, 1994). However time and budget are limiting factors. The *k value* depends on the soil type (which includes size, shape and distribution of pores), as well as soil temperature. In some cases the *k value* is the same in all *directions* (direction meaning width, height and depth; direction independent), called *isotropic* soil (horizontal conductivity = vertical conductivity) but in most cases this is not so. Due to layering and biological activity the texture, structure and porosity, the vertical permeability of the soil is often different than the horizontal permeability (direction dependent), called *anisotropic* soil (horizontal conductivity ≠ vertical conductivity). Next to this a soil can be homogeneous (same properties at every point) or heterogeneous (different properties at every point) as well (Figure 1.1).

In a soil profile, the topsoil is more variable in time than the subsoil, due to biological processes and drying and wetting. In clay soils this is more pronounced with swelling and shrinking properties (Oosterbaan, 1994).

Salazar et al. (2008) evaluated the feasibility of running DRAINMOD with k values determined using pedotransfer functions from particle size distribution and bulk density, which are easier to determine. TAlthough this does not incorporate anisotropy, their results suggest that k values estimated by pedotransfer functions can simulate drainage outflow as accurately as laboratory determined k values, however only for course textured soils. Jacobsen et al. (2011) mentions that anisotropy can also be dealt with by using a weighted k value, combining all layer k values and thicknesses :

$$K = \frac{k_1 D_1 + k_2 D_2 + ...k_n D_n}{D_1 + D_2 + ...D_n} \tag{1.1}$$

where D is the layer depth (cm). DRAINMOD requires the lateral conductivity as input but this is not measured in many cases (as for Norway). Soil samples for analyses are usually taken in the vertical direction, assuming isotropy.

1.2.3 Soil water characteristic

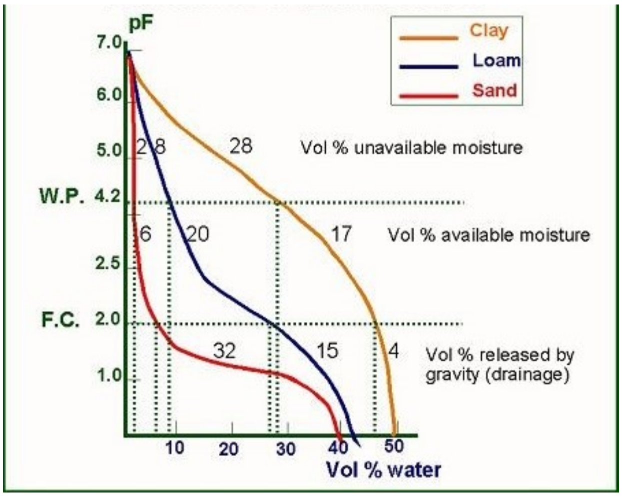

Figure 1.2: Soil moisture retention curve example (FAO, 2017 (Accessed on 23-2-2017).

The soil water characteristic, also called soil water retention curve or pF curve, is a measure of how strong water is held in the soil matrix in the unsaturated state. The pF curve is a way to express the metric head of water in the soil, expressed by pF = log (h), with h being the metric head (in cm) (Braun and Kruijne, 1994). It is the second most important soil property, after the hydraulic conductivity, in modeling soil water movement (Haan and Skaggs, 2003). Size and connectedness of pore spaces play an important role, which are influenced by soil texture, structure and soil organic matter (Tuller and Or, 2004). After a rainfall event the soil is complete filled with water, after which free drainage of the soils starts. Large pores in the soil matrix have small matric forces, meaning water will be released by gravity. When all this water is removed from the soil, the soil is at field capacity (F.C. in Figure 1.2) (Braun and Kruijne, 1994). This happens in the order of a few days, depending on the drainage capacity of the soil . It is considered the upper limit of water available for plants (Kabat and Beekma, 1994). Evapotranspiration will cause the water content in the soil to decrease, until the wilting point (W.P. in Figure 1.2), after which plants cannot extract any more water from the soil (Braun and Kruijne, 1994). When this permanent wilting point is reached depends on the crop and meteorological conditions (Kabat and Beekma, 1994). The water holding capacity or available soil moisture of the soil is the water stored between field capacity and wilting point. This is between pF 2.0 and 4.2.

1.2.4 Evapotranspiration

Evapotranspiration (ET) is a combination of two processes, evaporation from the soil and transpiration from the crop. It is an important part of the water balance, however it is difficult to measure or calculate, and empirical equations have been developed, but are only applicable for local conditions. That is why physically-based equations are used more often.

Evaporation is the conversion of liquid water into water vapor, which requires energy, such as solar energy. As water turns into vapor, the atmosphere becomes saturated, slowing down the evaporation process or even stopping, if the wet air is not replaced by drier air. The solar radiation, air temperature, air humidity and wind speed drive this process. The amount of water available at the surface affect this process. Water availability depends on rains, irrigation and the upwards water flux from the groundwater table. When enough water is available at the soil surface to evaporate, climatological conditions limit the process. In drier periods between rain events for example, limited water supply slows down the evaporation.

Transpiration is the conversion of liquid water into vapor, however this time within the plant tissue (within the leaf). Plants lose water mainly through small openings in the leaf (stomata). The vapor exchange with the atmosphere is controlled by the plant, and most of the water taken up by the plant leaves as transpiration, with only a little being used by the plant. The same climatological factors that influence the evaporation influence transpiration. However the transpiration is also influence by crop characteristics, the environment, cultivation methods, and soil moisture availability.

Although these two processes occur simultaneously, distinguishing them is difficult. It is also different during the growing season, with predominantly evaporation at the start and transpiration at the end being the main contributors (Allen et al., 1998).

Although it is difficult to measure or calculate, there are still quite a few methods of calculating the PET, which can be divided into three main categories (Zhao et al., 2013): energy based, which applies the energy balance concept; temperature based; and mass-transfer based, which estimates free water surface potential evaporation, taking air pressure deficit and wind speed into account. Examples of these categories are:

- Energy: Penman-Monteith (Allen et al., 1998), Makkink (1957), Priestley and Taylor (1972) and Abtew (1996);
- Temperature: Blaney and Criddle (1952), Hargreaves (1974) and Thornthwaite (1948);
- Mass-transfer: Dalton (1802), Penman (1948) and Rowher (1931)

Of these three, the mass-transfer based methods are one of the oldest, for example Dalton's law of Evaporation (Dalton 1802). Blaney-Criddle was recommended by the FAO if only air temperature was known (Doorenbos and Pruit, 1977) and Droogers and Allen (2002) suggest using the Hargreaves over the Penman-Monteith method if no accurate weather data is present.

For evapotranspiration to occur, three requirements must be met: enough water; energy to transfer liquid into vapor; and a vapor gradient which takes the vapor away from the soil surface into the atmosphere. Most of the PET calculation methods are based on one or more of these requirements (Feddes and Lenselink, 1994).

FAO Penman-Monteith Penman developed an equation, a combination of the energy balance and the mass-transfer method, to calculate the evapotranspiration from an open water surface, from sunshine, temperature, humidity and wind speed. Cropped surfaces and resistance factors were added to the equation later on. The resistance factor includes aerodynamic resistance and surface resistance factors, often bulked into one parameter. These parameters combined is the Penman-Monteith equation. The FAO Penman-Monteith method was developed by defining a reference crop with the following equation inputs: *"A hypothetical reference crop with an assumed crop height of 0.12 m, a fixed surface resistance of 70 s/m and an albedo of 0.23."* The resulting evapotranspiration resembles that of well-watered green grass, and forms the reference evapotranspiration ET_0 (or Potential Evapotranspiration: PET). This creates a standard to which evapotranspiration of different periods or regions and crops can be compared and related. The equation requires location (altitude and latitude), air temperature, humidity, radiation and wind speed as inputs. The next step is the addition of the crop coefficient K_c. The K_c incorporates the differences in crop canopy and aerodynamic resistance, which serves as a collection of the differences between crops. It can be calculated by relating the measured crop evapotranspiration ET_c to the ET_0, by $K_c = ET_c/ET_0$. The FAO has many Kc values for different crops, which can be used to calculate the ET_c from the ET_0 (Allen et al., 1998).

Hargreaves Unlike the previous method, this method is simple and needs minimal input data, namely maximum and minimum temperature. Where solar radiation, humidity and wind data are missing this method can still give reliable ET_0 data. If equipment cost are high or data quality is unreliable, the Hargreaves method is recommended. However time steps of 5 days should be taken to decrease variance in temperature ranges due to wind or cloud cover (Hargreaves and Allen, 2003).

Thornthwaite The third method requires mean monthly temperature and the latitude of the location to calculate the monthly potential evapotranspiration. The first step is to compute the heat index, which is the summation of each monthy heat index value. The second step is the determination of the potential evapotranspiration, still unadjusted. The last step is to adjust the potential evapotranspiration for hours of sunlight (Cruff and Thompson, 1967).

Comparable PET from other studies Literature values for wheat, pine and spruce can be used as a guideline. The FAO provides a guideline for wheat, which needs approximately 450-650 mm of water for high yields, depending on climate and length of growing season (Brouwer and Heibloem, 1986).

A study in Finland by Tao et al. (2015) in which wheat yield is simulated shows that their ET is lower than 450 mm for the growing season, namely 130-200 mm. Another study in Finland simulated groundwater recharge and found ET results for the Pinus sylvestris at around 240 mm per year (Ala-Aho et al., 2015). Regarding the Picea abis (Norway spruce), Ge et al. (2010) found that the ET was 88-93% of the annual precipitation in Finland (wich equals 474 mm per year). Lagergren et al. (2008) measured ET for the months June, July and August, with results varying from 144-186 mm and 136-155 mm, for thinned and reference plots, in central Sweden.

1.2.5 Considering slope

The slope angle has a great influence on occurrence of surface runoff, as do local depressions on the soil surface. For the simulations these inputs are needed, as they determine the ration between subsurface and surface runoff. Incorporating slope into DRAINMOD is possible by using the method described by Onstad (1984):

$$D_{sc} = 0.11R + 0.031R^2 - 0.012RS \qquad (1.2)$$

Where D_{sc} is the maximum depressional storage, R is the random roughness (cm) and S is the slope steepness (%). Figure 1.3 shows the equation . Higher slopes result in lower depressional storage with the same random roughness, with this effect being stronger for higher random roughness values. However, no field measurements have been taken of this random roughness, making estimation of the maximum depressional storage difficult, and in turn the ratio between subsurface and surface runoff. Available data on runoff only includes these two flows together, so no separated discharge values are known.

This issue will be overcome by comparing Skuterud with similar catchments of the JOVA program where the ratio between subsurface and surface runoff is known. Kværnø and Bechmann (2010) published an overview of 10 catchments in Norway,

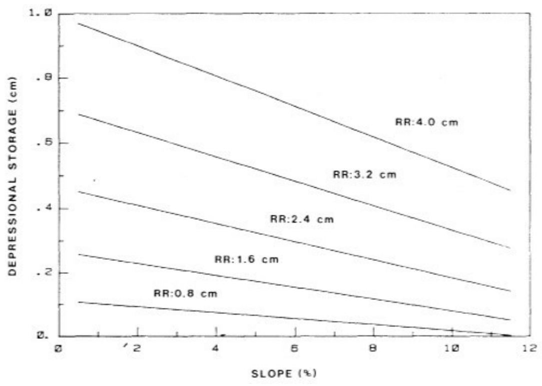

Figure 1.3: *Example of maximum depressional storage, random roughness and slope steepness (Onstad, 1984).*

although all of them had smaller surface areas. Figure 1.4 shows the division between surface and subsurface runoff. Most of the catchments have discharge their water through subsurface runoff, between 9 and 25%, whilst two catchments show equal preferences. Total runoff values range between 169 mm per year for Bye to 607 mm per year for Kvi (Figure 1.5). The catchments with higher yearly precipitation also show a higher percentage of runoff compared to precipitation (around 60% of precipitation leads to runoff). To compare, Skuterud on average has 519 mm of runoff and 785 mm precipitation per year, which is comparable to Syv and Ene. These catchments have slopes of 13 and 5%, respectively, which also occurs in Skuterud. Thus, around 20 % of surface runoff will be simulated.

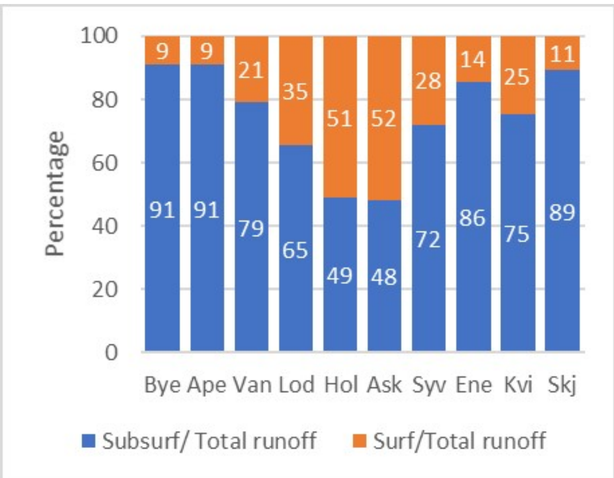

Figure 1.4: *Division of surface and subsurface runoff for 10 catchments in Norway (Kværnø and Bechmann, 2010).*

Within a year the amount of surface runoff varies, due to processes such as spring snowmelt and frozen soils. Bechmann (2014) analyzed nitro-

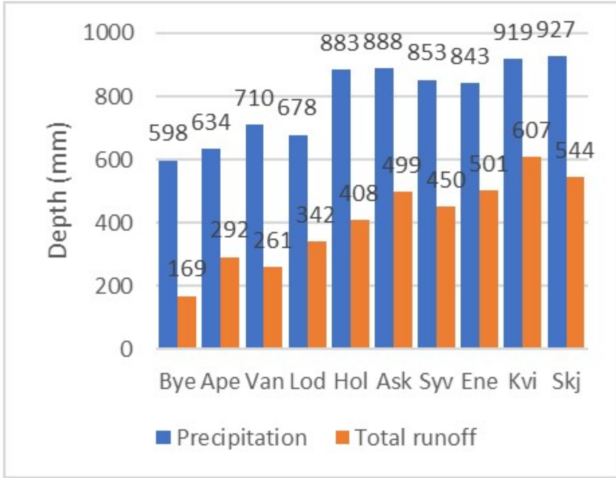

Figure 1.5: *Precipitation and total runoff for the 10 catchments in Norway (Kværnø and Bechmann, 2010).*

gen content in surface and subsurface runoff of 10 different catchments, which also include Bye and Van from figures . The results show that in Bye, surface runoff occurs from January to April, whilst in Van nearly all months show surface runoff, however most also occurs from January to April, which is caused by snowmelt (55% and 35% of total surface runoff, respectively).

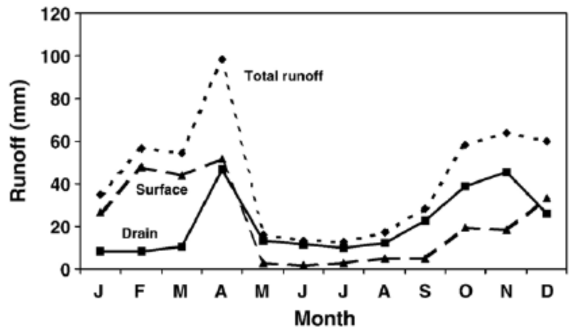

Figure 1.6: *Monthly division of runoff for a comparable catchment to Skuterud.*

Finally, Lundekvam (2007) also performed some research on hydrology and erosion in Norway, of which the result of Askim (near the research area) is relevant, as it has a similar climate. Results show that during December-April, up to 50 mm per month of surface runoff occurs and during autumn there is also runoff, however from May to August there is little surface runoff (see Figure 1.6).

To conclude, the slope will not be used directly in simulations, instead the maximum surface storage will be adjusted to get runoff results that are similar to the above measurements.

2 | Catchment description

The Skuterud catchment (outlet coordinates: $59°41'4.92''N, 10°49'53.02''E$, Figure 2.1) lies 30 kilometers south of Oslo, Norway, neas Ås. The total area of the catchment is approximately 440 ha. Of this, 272 hectares is used as agriculture (62%), 129 hectares is forest (30%) and the remaining 38 hectares is urban area (8%) (Figure 2.2). Main cultivated crops include barley, oats, wheat from April to August, and rye and wheat from autumn onwards. The forest consists of Nordic spruce and Pine tree (Deelstra, 2005).

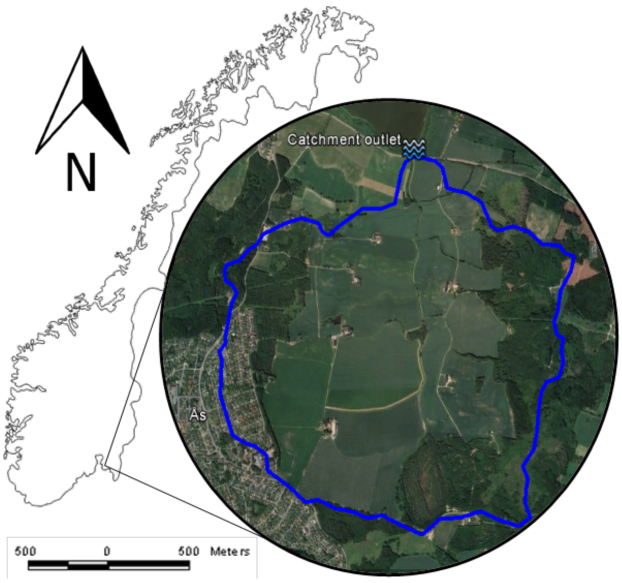

Figure 2.1: *Skuterud catchment in southern Norway.*

2.1 Climate

Climate has been observed in Ås since 1874 (Institute, 2017 (Accessed on 14-3-2017), which includes air and soil temperature, precipitation, snow days and depths, sunshine hours air pressure and humidity. Yearly data summaries are published as a report series "Meteorologiske data for Ås" from the Norwegian University of Life Sciences (NMBU, 23-01-2017), which also has includes the normal temperature and precipitation calculated by Gjelten et al. (2014), which is a 30 year period of averaged climate data, in this case from 1961 to 1990, called *'normal'* hereafter.

The average normal temperature for Ås is 5.3 °C, with temperatures lowest in January and February (-4.8 °C) and highest in July (16.1 °C) (Figure 2.3, crossbar indicates range between minimum and

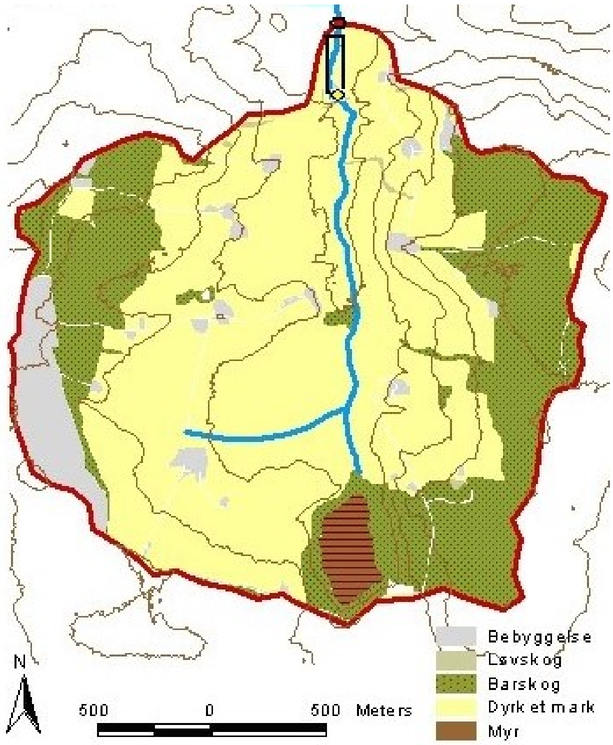

Figure 2.2: *Land use map of Skuterud. Green = forest, beige = agriculture, brown = bog and grey = urban.*

maximum values). The average temperature (1993 to 2015) is 6.2 °C, with the lowest temperature in January (-3 °C) and the highest in July (16.7 °C). In 1996 and 2010 the average yearly temperature was below the normal temperature. For the other 21 years the average temperature was higher than normal, with a maximum of 7.7 °C in 2014.

The annual normal precipitation in Ås is 785 mm, with the lowest amount in February (35 mm) and the highest in October (100 mm). The average precipitation from 1993-2015 is 909 mm, with the lowest amount of precipitation in March (44 mm) and the highest in October (111 mm). In the years 1993, 1996, 1997, 2003 and 2005 the amount of precipitation was lower than the normal precipitation. In the other 18 years the precipitation was higher than normal, with a maximum of 1200 mm in 2000 (Figures 2.4 and 2.5).

2.2 Geology, topography and landforms

Most of the catchment area is covered by fine marine deposits, with occasional gravel and stones. Moraine ridges transect the catchment at several

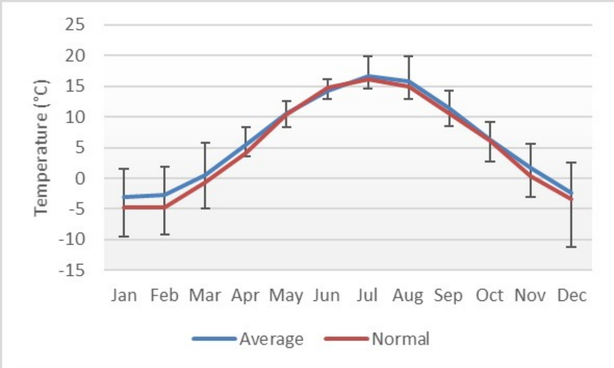

Figure 2.3: *Monthly average (1993-2015) and normal (1961-1993) temperature.*

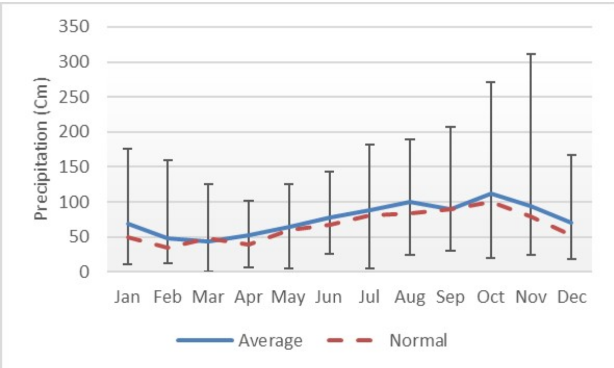

Figure 2.4: *Monthly average (1993-2015) and normal (1961-1993) precipitation.*

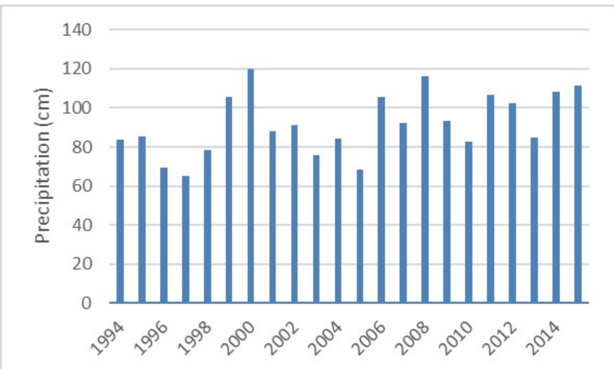

Figure 2.5: *Annual precipitation for Ås (1994-2015).*

places, as a result of the melting ice cap from the last glaciation. There is not a lot of exposed bedrock in the catchment, and there is a drained bog in the south-western part. The underlying bedrock is from the geological period Pre-Cambrium, such as gneiss. The highest point in the catchment is approximately 150 m.a.s.l., whilst the lowest point is 85 m.a.s.l. Slopes are long and gentle on the western side of the stream (Figure 2.6), whilst the slopes are shorter and steeper on the eastern side (Deelstra, 2005).

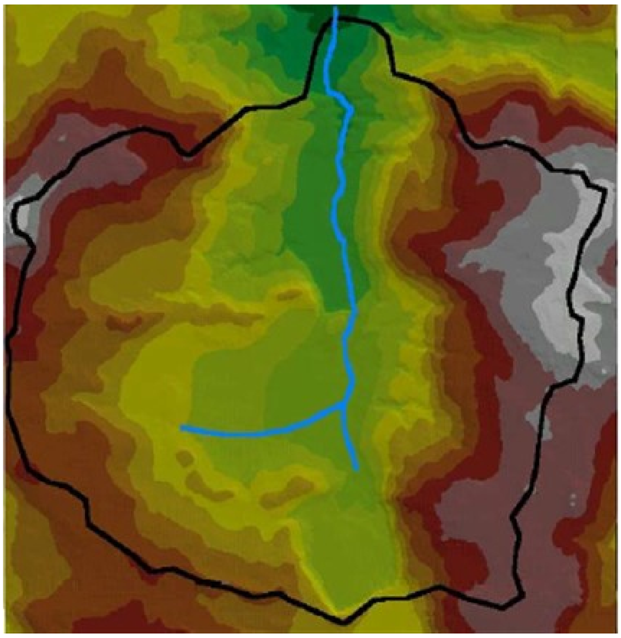

Figure 2.6: *Elevation map of Skuterud. Lighter colours have lower elevation levels, and darker have higher elevation levels.*

Table 2.1: *Soil layer depth (cm) and saturated hydraulic conductivity (k, cm h^{-1}) of different soils present in Skuterud.*

Rk		Je		He	
Depth	k	Depth	k	Depth	k
0-26	6.13	0-33	17.64	0-23	4.8
26-34	2.38	33-50	19.58	23-33	37.0
34-71	12.63	50-85	47.07	33-54	49.3
71 +	0.01	85 +	115.05	54+	49.8

Thirty-one main soils have been identified and classified, which were aggregated into three dominating soil types, being marine sand and morenic deposits (*Je*, Endostagnic Albeluvisols) and mainly marine silty clay loam (*Rk*, Luvic Stagnosol and *He*, Endostagnic Cambisol). The soils in the catchment have been classified according to the World Reference Base for Soil Resources. Crops are grown on all three soils, whilst the forest only grows on the silt loam (Je) (Deelstra, 2005). Table 2.1 shows the saturated hydraulic conductivities for the different soils and layers of those soils. There is a lot of variety for the different layers within the soils, as well as when comparing the soils. Of the total arable area in Skuterud (272 ha), 205 ha is on Rk soil (75%), 37 ha is on Je soil (14%) and 30 ha on He soil (11%). The 129 ha of forest is on Je soil.

2.3 Farming system

The catchment has nine farms, with a total of 51 individual fields, which have provided detailed information about farming practices since 1993, including tillage operations, crops types and sowing/harvesting dates. Most of the fields are systematically drained, with drain spacing between 8-10 meters and at a depth of between 0.8 and 1 meter (Deelstra, 2005).

2.4 Catchment runoff

Yearly catchment runoff varies from a low runoff of 256 mm in 1996 to a high runoff of 783 mm in 2000 (Figure 2.7). The average annual runoff of the period 1994-2015 was 519 mm. Runoff includes surface runoff as well as subsurface runoff, in addition to baseflow. Discharge is measured since 1994, using a Crump-weir with an automatic data-logger in combination with a pressure transducer and calculated on the basis of the existing head-discharge relation (Deelstra, 2005). This data will be used for the calibration and validation in chapter 5, as it is the only measured part of the water balance, besides precipitation.

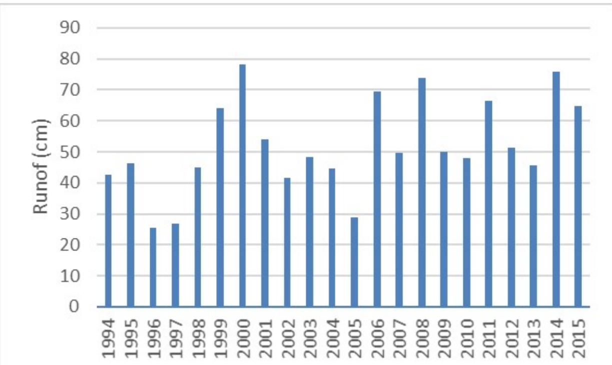

Figure 2.7: *Annual catchment runoff (1993-2015).*

Figure 2.8 shows the monthly average runoff of the period 1994-2015, with two peaks, one in April, the second around November. The peak in April can be explained by snowmelt, the second one at the end of the year by more precipitation. The period between November and April has freezing temperatures, thus precipitation is stored on the surface as snow. Runoff is mostly the aquifer being drained in this period. The seasonality gives more insight in this, which shows how much of the annual precipitaiton or runoff occurs in a season. It is calculated by taking the monthly measured precipitation or runoff as a percentage of the annual values (Table 2.2) (Øygarden et al., 2014). In the winter 26% of the annual runoff occurs, in spring 30%, summer 9% and autumn 34% (Table 2.3). During the summer there is quite some precipitation (29% of annual precipitation), however only 9% of the annual discharge occurs. The spring snowmelt can be seen in April again, as 15% of annual runoff occurs, whilst only 6% of annual precipitation falls.

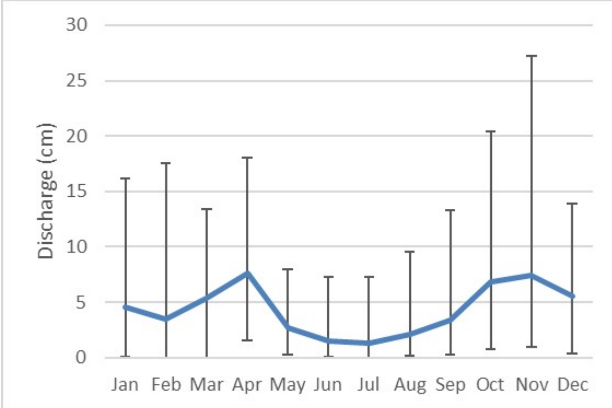

Figure 2.8: *Monthly average catchment discharge.*

Table 2.2: *Seasonality: average monthly value compared to average annual value, for precipitation and runoff (%).*

Season	Month	Precipitation	Runoff
	Dec	8	11
Winter	Jan	8	9
	Feb	5	7
	Mar	5	10
Spring	Apr	6	15
	May	7	5
	Jun	9	3
Summer	Jul	10	2
	Aug	11	4
	Sep	10	7
Autumn	Oct	12	13
	Nov	10	14
	Total	100	100

Figure 2.9 shows the cumulative runoff over the period 1994-2015. Each peak is the total amount of water that the catchment discharged for that spe-

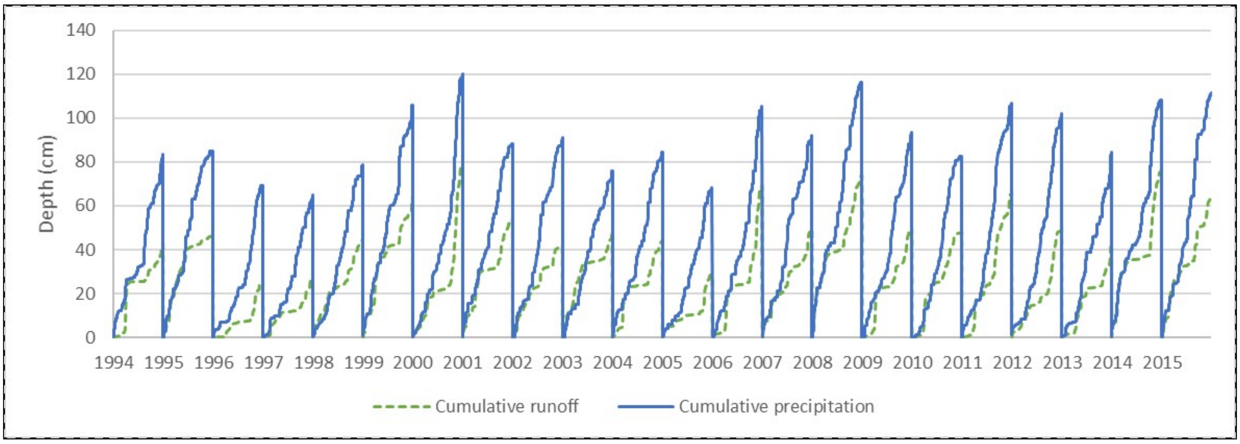

Figure 2.9: Daily and yearly cumulative discharge (1994-2015).

Table 2.3: Seasonal percipitation and runoff values.

Season	Precipitation (%)	Runoff (%)
Winter	21	26
Spring	18	30
Summer	29	9
Autumn	32	34
Total	100	100

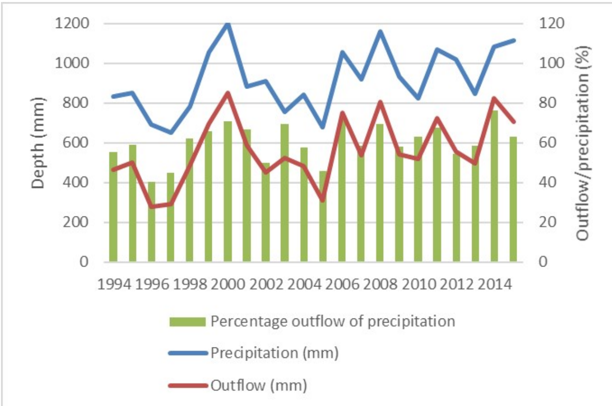

Figure 2.10: Monthly average precipitation and catchment runoff, and percentage discharge of precipitation.

cific year, which differ per year. The year 2000 has the highest peak (~80 cm), whilst 1997 has the lowest (~25 cm). A common pattern is a steep increase in spring (snowmelt), followed by a plateau in summer (increased evapotranspiration) and another steep increase in autumn (saturated soils).

On the annual scale runoff follows precipitation, so when there is more precipitation there is more discharge (Figure 2.10). However, the same figure shows that not all years have the same outflow/precipitation ratio. The highest ratio is in 2014 (precipitation 1082mm, discharge 825 mm) with 76%, the lowest is in 1996 (precipitation 692 mm, discharge 276 mm) with only 40%. In general, with higher precipitation, not only more discharge in total quantities occur, but also a higher percentage of the precipitation leaves the catchment as discharge. This might be because in drier years a higher percentage of precipitation leaves the catchment as ET, as basic vegetation requirements need to be met, whilst in wetter years there is an abundance of water for the vegetation and soil, thus water leaves the catchment as runoff.

The catchment also includes an urban area, which covers 8% of the area, of which the measured runoff is 7.7% of the total runoff. Assuming the urban runoff does not infiltrate in the agricultural area, but rather flows to the catchment outlet without hinder, the urban area contributes to 8% of the total runoff. This has been removed from the observed discharge when comparing it to the simulation results.

3 | DRAINMOD model description

DRAINMOD is a field scale simulation model that quantifies drainage and other related water management systems, such as evapotranspiration and surface runoff. It describes the hydrology of soils that have poor natural drainage and high water tables and therefore have an artificial drainage system. . They hydrology is based on simple water balances in the soil profile, on the soil surface, and of the drainage system (figure 11). Nitrogen and carbon cycles in subsurface drainage water and salinity effects on yield and irrigation management can also be simulated (DRAINMOD-NII, not used in this thesis (Youssef, 2003)), next to prediction of soil temperatures and the effects of freezing, thawing and snowmelt (Luo et al., 2000). The model predicts hydrologic variables, such as infiltration, subsurface drainage, surface runoff, evapotranspiration, vertical and lateral seepage and water table depth, as well as other outputs such as crop yield (Skaggs et al., 2012)

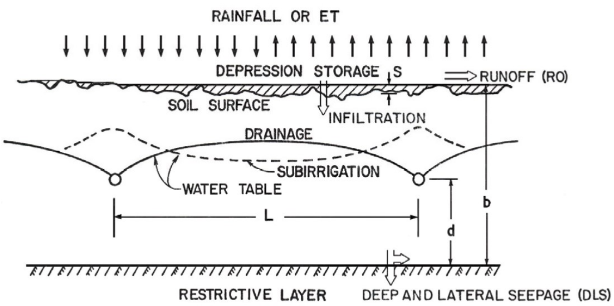

Figure 3.1: *Hydrologic components of a drainage management systems (Skaggs et al., 2012).*

The model uses two main water balance equations for the surface and soil, with the climate acting as main driver for the processes (Equations 3.1 and 3.2) :

$$P = F + \Delta S + RO \tag{3.1}$$

$$\Delta V = D + ET + DLS - F \tag{3.2}$$

Where P is precipitation, F is infiltration, ΔS is the change in surface water storage, RO is surface runoff, ΔV is the change in soil water, D is drainage, ET is evapotranspiration and DLS is deep and lateral seepage. The two processes are linked by the infiltration of water from the surface to the subsurface (the soil).

The following subchapters will briefly explain the processes and inputs needed for DRAINMOD.

3.1 Weather data

Weather data includes precipitation, daily air temperatures (maximum and minimum) and potential evapotranspiration (either as an input file or calculated by the model using Thornthwaite). Precipitation can be either on the hourly basis or on a user-specified time daily basis input, and does not use any equations. User-specified means that the same daily precipitation falls, starting form a chosen time and over a chosen period of time. This is used if hourly data is not available.

3.2 Infiltration

Information about the infiltration capacity of the soil are not available. If no infiltration data is available, DRAINMOD will calculate the infiltration using the hydraulic conductivity and the soil pF curve, the Green and Ampt equation (Equation 3.3):

$$f = K + \frac{KM_d S_f}{F} \tag{3.3}$$

where f is the infiltration rate, (cm h^{-1}), F is cumulative infiltration (cm), K is the vertical hydraulic conductivity of the transmission zone (cm h^{-1}), M_d is the difference between final and initial volumetric water contents (cm^3 cm^{-3}), and S_f is the effective suction at the wetting front (cm)(Skaggs, 1980). See chapter 2.1.9 Soil physical parameters for more information about how DRAINMOD calculates this .

3.3 Surface drainage

Surface drainage, or runoff amounts, depend on the surface storage, which needs to be filled before runoff starts. The storage is divided into two parts, the first being water that can freely flow over the surface and the second being local depressions where water flow is blocked (S1 and S2 in figure 3.2). When the storage is full, ΔS becomes 0, and runoff occurs, calculated by $RO = P - F$, from equation 3.1.

3.4 Subsurface drainage

Water that has infiltrated into the soil can leave through three ways, being either drainage, ET or seepage. Based on the depth of the water table from the soil surface (positions 1 to 6 in Figure 3.2), three equations are available to calculate the subsurface drainage, namely the steady-state Hooghoudt equation, Kirkham's equation and Darcy's law. 3.2a shows the different water table elevations, 3.2b shows the drainage rate plotted as a function of the midway elevation point of the water table m. The figure shows that different water table positions (1 to 5) have different drainage rates.

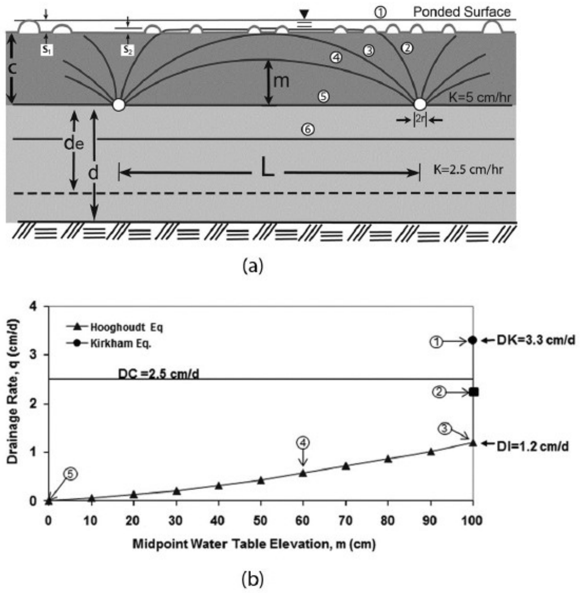

Figure 3.2: *Relationship between water table elevation and drainage rate (Skaggs et al., 2012).*

The calculation of subsurface drainage is based on the idea that water movement in the saturated soil is in the lateral direction. When the soil surface storage ΔS is full, the drainage rate q (cm h^{-1}) is calculated using equations developed by Kirkham (position 1):

$$q = \frac{4\pi K_e(t+b-r)}{Lg} \tag{3.4}$$

where t is ponded depth (cm), b is distance from surface to drain (cm), r is the radius of drain (cm), K_e is the equivalent lateral hydraulic conductivity of the profile (cm h^{-1}), L is the drain spacing (cm) and g is a constant and calculated by DRAINMOD for a given drain size, depth, spacing, and depth of profile (Skaggs, 1980). At position 1 the subsurface drainage is at its maximum, which is also the design discharge for the drain pipe. Due to drainage and ET the water table continues to drop (position 2) and Kirkham's equation is no longer valid. Drainage rates continue to decline until the ponded water is removed and the midpoint water table elevation is at the same height as the surface, at which point Hooghoudt's steady state equation is used to determine the drainage (position 3):

$$q = \frac{4K_e m(2d_e+m)}{L^2} \tag{3.5}$$

where q is the drainage rate (cm h^{-1}), m is the midpoint water table elevation above the drain, K_e is the equivalent lateral hydraulic conductivity of the profile (cm h^{-1}), d_e is the equivalent depth from the drain to the restrictive layer (cm), and L is the drain spacing (cm). The equivalent depth takes convergence of flows below the drain into account, by reducing the layer between the drain and restrictive layer as follows (Oosterbaan and Ritzema, 1993):

$$\text{If } d < R \text{ then } d_e = d \tag{3.6}$$

$$\text{If } R < d < \frac{L}{4} \text{ then } d_e = \frac{dL}{(L-d^2)+8dLln(\frac{d}{R})} \tag{3.7}$$

$$\text{If } d > \frac{L}{4} \text{ then } d_e = \frac{L}{8ln(\frac{L}{R})} \tag{3.8}$$

where d is layer between drain and restrictive layer and R is the drain radius. The equivalent depth is used because the drains are located above the restrictive layer, which means water flows will be partly horizontal and partly radial when nearer to the drains. This restricts the flow, because the cross section through which the water flows decreases (in other words, resistance increases)(Van Beers, 1976). The equivalent K_e belongs to d_e.

Although the drawdown process (positions 3 and 4) is not a steady-state process, for which Hooghoudt works, the equation estimates relatively well (Skaggs and Tang, 1976). When the water table falls below the drains (position 6), the drainage rate becomes zero. Limitations for the drainag rate ()Equations 3.4 and 3.4) are the hydraulic conductivity of the soil and the hydraulic capacity of the pipes, which is called the drainage coefficient (DC) (Skaggs et al., 2012).

3.5 Soil water distribution

The soil above the water table is the unsaturated zone. The soil moisture in this zone is used by the plants during the growing season and it slowly dries up. When drying up there will be an upward movement of water from the water table to supply water to the roots (capillary rise). At a certain moment the soil moisture content becomes low and the capillary rise cannot supply the needed water, and the moisture content moves into the direction of wilting point, and plants have problems getting hold of water. DRAINMOD needs the soil water characteristic (*pF curve*) for each soil layer and the volumetric water content at the permanent wiling point to calculate the soil water distribution and volume of water free pore space (Skaggs et al., 2012).

3.6 Evapotranspiration

Evapotranspiration determination is a twostep process in DRAINMOD. First, the daily potential evapotranspiration (PET) is determined, which represents the amount of water that will be removed from the soil-plant system by evapotranspiration (ET) when soil moisture content is not a limiting factor, as explained in the soil water distribution section and further explained in this sub chapter. If it is a limiting factor the ET is set equal to the smaller amount that can be supplied from the soil system.

Two methods of input are possible for the PET. The first is using local climate data to calculate the PET using any of the available formulas, such as the ones mentioned before. The second method may be calculated by DRAINMOD with the temperature-based Thornthwaite method, which typically under-predicts PET during fall, winter, and spring months and over-predicts during the summer (Skaggs et al., 2012). However, the method has been found to give reliable estimates when used with monthly correction factors (Amatya et al., 1995), which can be determined by using the Penman-Monteith method for a relatively short record (2 to 5 years) for a weather station, and then setting the correction factors for each month equal to the ratio of the monthly PET values to the respective values calculated with the Thornthwaite method. Monthly adjustment factors should only be minimally adjusted (<15%)(Skaggs et al., 2012).

Once PET is established, the model determines if soil water conditions are limiting. Inputs used to determine whether soil water conditions limit ET are (1) the soil water characteristic, (2) the relationship between maximum steady upward flux and water table depth, (3) the effective depth of the root zone, and (4) soil water content at the lower limit (permanent wilting point). Section 3.9 will go into more detail about these four inputs.

ET is calculated using the following equation (Norero, 1969), by

$$ET = \frac{PET}{1 + (\frac{\psi}{\psi^*})^2} \tag{3.9}$$

where k is a constant, ψ is the soil water potential in the root zone (obtained from soil water characteristics) using the average root zone content, and ψ^* is the value of ψ when ET = 0.5 PET (Skaggs, 1980).

3.7 Soil temperature

If freezing and thawing are present in the study area, this can be included into DRAINMOD. The daily maximum and minimum temperatures are needed for this. When freezing conditions are indicated by below zero temperatures, the model calculates soil temperature, by solving the heat flow equation, and ice content in the soil profile and modifies the soil hydraulic conductivity and infiltration rate. Precipitation is separated into snow or rain, depending on the rain/snow dividing temperature and whether the daily average temperature is above/below this threshold. If the temperature is below the snow melting threshold temperature, snow will accumulate on the soil surface. Soil surface temperature is recalculated when snow cover exists. If the snow melts this amount is added to the rainfall, which can then infiltrate or run off depending on freezing conditions (Luo et al., 2000).

Feezing and thawing Luo et al. (2000, 2001) introduced the processes of freezing, thawing and snowmelt into DRAINMOD, by solving the heat flow equation to predict soil temperature. The thermal conductivity of the soil depends on the composition of soil particles and the soil water content, and is calculated by (Karvonen, 1988):

$$\lambda = a + b\sqrt{X_w} + 4b\sqrt{X_i} \tag{3.10}$$

where λ is the thermal conductivity, X_w is the volume fraction of water, X_i is the volume fraction of ice, and a and b empirical are coefficients. When predicted soil temperature is below the threshold temperature for freezing conditions, the unfrozen water content in soil can be obtained from a soil freezing characteristic curve, which relates unfrozen water content to below-zero temperature.

Snow accumulation/melt A rain-snow dividing temperature determines if precipitation is rain or snow. Precipitation is considered as snowfall if the daily average air temperature is below the rain-snow dividing temperature. This snow accumulates on the soil surface until the average air temperature rises above the rain-snow dividing temperature. The total snow depth is calculated by tracking the amount of snow water equivalent and considering the daily densification effect. When the soil is covered by snow, the soil surface temperature is modified assuming an equilibrium condition between snow and the first soil increment. Only the air temperature is considered as the driving force for snow melt. Above the snowmelt base temperature, snowmelt is calculated using the degree-day method, based on equivalent depth of water of the daily snowmelt, mean daily temperature, snowmelt base temperature, and the degree day coefficient (Davar, 1970):

$$M = C(T_a - T_b) \tag{3.11}$$

where M is the equivalent depth of water of the daily snowmelt (cm), C is the degree-day coefficient (cm °C^{-1} day^{-1}), T_a is mean daily temperature (°C) and T_b is snowmelt base temperature (°C).

The amount of infiltration water from snowmelt is limited by the soil ice content. When the ice content exceeds a critical value infiltration ceases and snowmelt leaves as surface runoff. The value of the critical ice content can be estimated as about 60% of field capacity (θ at -300 cm) (Luo et al., 2000, 2001). When ice is formed, soil hydraulic conductivities are modified using the equation by Motovilov (1978):

$$K(\theta_i) = \frac{K(\theta_w)}{(1 + 8\theta_i)^2} \tag{3.12}$$

where $K(\theta_i)$ and $K(\theta_w)$ are hydraulic conductivities with and without ice respectively, and Îÿ$_i$ and Îÿ$_w$ are volumetric contents of ice and unfrozen water, respectively.

3.8 Seepage

DRAINMOD has the possibility to incorporate lateral and vertical seepage into the simulations, however this function has not been used, because of the assumption that the clay layer becomes an impermeable layer after 2 m depth (see *k values*, table 2.1 in chapter 2.2)

3.9 Soil physical parameters

The two most important soil physical parameters for DRAINMOD are the soil hydraulic conductivity and the soil water characteristic curve (of pF curve). These are both used to determine other inputs for DRAINMOD, namely the relationship between water table depth and the drainage volume or upward flux, when data is not available from the field. These are explained in more detail below (Skaggs, 1980):

- Hydraulic conductivity: artificial drainage usually involve lateral flow to and from drains, thus effective horizontal conductivity is used in the model. Soil data which includes the hydraulic conductivity usually represent the vertical k of the layer, instead of the horizontal k, which may differ by a factor 10.
- Soil water characteristic: the soil water characteristic h(θ) is a measure of how tight water is held in the soil matrix in unsaturated state. It is used in DRAINMOD to determine other input factors, such as the relationship between water table depth and drainage volume, upward flux and infiltration parameters. This, together with the hydraulic conductivity, is the most important in soil water modeling.
- Water table depth vs drainage volume: DRAINMOD determines how far the water table rises or falls if a certain amount of water is removed or added. The water yield (volume of water drained) can be calculated with the soil water characteristic. For this the assumption is used that the unsaturated zone is drained to equilibrium, and that the water table recedes in such a way that the vertical hydraulic gradient above this water table is zero. The volume drained per unit area is the difference between θ_o(y) prior to drainage (assumed to be the saturated value) and θ(y) at water table depth y.
- Water table depth vs upward flux (capillary rise): the upward flux is calculated mathematically using the unsaturated hydraulic conductivity for the different layers and the pressure head.
- Green-Ampt equation parameters: these can be determined from field measurements. However, these are not available for this thesis. DRAINMOD uses the *k value* and h(θ) to calculate the parameters.

3.10 Farming system

Most of the model inputs have been discusses, however a few remain. These are crop input factors such as rooting depth and water stress, and management factors such as weir and drainage settings which include the drain distance and depth.

4 | Methodology

4.1 Calibration & validation

Calibration will be done for the years 1994-2000, with 1993 as a warmup period. Values for the different parameters will be determined from literature as much as possible. Validation will be done for the years 2001-2015, with 2000 as a warmup period. DRAINMOD will be calibrated by comparing the simulated daily, monthly and yearly runoff with the observed, using the Nash-Sutcliffe modelling efficiency (NSE) for evaluation (Moriasi et al., 2007; Skaggs et al., 2012). Simulated drainage and surface runoff will be treated as one, as drainage and surface runoff have not been measured seperately. Two methods for the NSE will be used:

$$NSE(abs) = 1 - \frac{\sum_{i=1}^{n} |O_i - P_i|}{\sum_{i=1}^{n} |O_i - \bar{O}|} \tag{4.1}$$

$$NSE(sq) = 1 - \frac{\sum_{i=1}^{n} (O_i - P_i)^2}{\sum_{i=1}^{n} O_i - \bar{O}^2} \tag{4.2}$$

Where O is observed, P predicted and n is the total number of observations. For the NSE, ranges vary between $-\infty$ and 1, with 1 showing a perfect fit. Two methods for the NSE are used as the squared method is more sensitive to higher flows or peaks whilst the absolute method is less sensitive to these extremes (Krause et al., 2005), which may give a more realistic result as low flows occur more often than high flows. Table 4.1 shows acceptable values for NSE, which serve as a guideline.

Table 4.1: *Indicators for NSE, source: Skaggs et al. (2012).*

Discharge	Statistic	Acceptable	Good	Excellent
Daily	NSE	>0.40	>0.60	>0.75
Monthly	NSE	>0.50	>0.70	>0.80
Annual	NSE	>0.60	>0.75	>0.85

4.2 Numerical experiments

Numerical experiments will be carried out to see which parameters have the greatest effect on distribution of water between surface and subsurface water, and how the ET is affected by changing the parameters. The effects of different parameter values for surface storage, drainage coefficient, drain spacing, lateral saturated hydraulic conductivity and threshold temperatures will be analyzed in this thesis. These parameters have been chosen based on previous studies done in Skuterud, using DRAINMOD (Deelstra et al., 2010a; Farkas et al., 2016) and the paper by Haan and Skaggs (2003). These experiments have only been done on the Rk soil, for illustrational purposes.

The surface storage, drainage coefficient, and drain spacing will be tested by changing the parameter with the following percentage of the original value: ±10%, ±25%, ±50%, -95%, +100% and +200% of the default value (based on the DRAINMOD reference report (Skaggs, 1980). The lateral saturated hydraulic conductivity will be tested by using the following values 0.01, 0.05, 0.1, 0.5, 1, 1.5, 2, 3, 5, 10 cm h^{-1}, per layer, with the other layers at their original value, whilst the threshold temperatures will be tested by using equal steps of 1 °C, from -1 °C to 3 °C. The critical ice content will be simulated with steps of 0.1 cm^3 cm^{-3}, from 0 to 0.9 cm^3 cm^{-3}.

For the simulations the assumption P = ET + DR + RO has been used, meaning lateral seepage is not considered. For the precipitation input the observed daily values are evenly distributed over a certain amount of hours, starting from a certain time, in this case 10 hours from 10 AM onwards.

Table 4.2: *Parameter settings for DRAINMOD calibration, period 1994-2000.*

Model parameters	Rk	He	Je	Fo
Drainage system				
Drain depth (cm)	100	100	100	100
Drain spacing (m)	10	8	10	100
Depth to impermeable layer (m)	2	2	2	2
Effective drain radius (cm)	1,5	1,5	1,5	1,5
Surface storage (cm)	0,3	0,3	0,5	0,5
Drainage coefficient	1	1	1	1
Freeze/thaw, snowmelt characteristics				
Rain-snow dividing temp. (°C)	0	0	0	0
Melt temp. (°C)	1,5	1,5	1,5	1,5
Melt coefficient (mm °C d^{-1})	3	3	3	3
Critical ice content(cm^3 cm^{-3})	0,2	0,2	0,2	0,2
Lateral Sat. Conductivity (cm d^{-1})				
1st layer	1	1	1	1
2nd layer	1	1	1	1
3rd layer	0,1	0,1	0,1	0,1
4th layer	0,1	0,1	0,1	0,1
Weather				
Monthly factor	2	0,8	0,8	1

4.3 Calibration result

DRAINMOD has been calibrated against the observed discharge for the period 1994 to 2000, taking 1993 as warmup period. The daily temperature, precipitation and potential evapotranspiration value have been taken as input, with Table 4.2 showing other DRAINMOD parameters. These are based on Deelstra et al. (2010a), however the lateral saturated conductivity and monthly evapotranspiration adjustment factor (here monthly factor) are different. Figure 4.1 shows the Nash-Sutcliffe results for daily and monthly values, per year. This has been acquired by calculating the daily or monthly Nash-Sutcliffe for only one specific year. The overall values, so not taking individual years into account, can be found in Figure 4.1. Overall there is a reasonable fit, although there are some underestimated years, namely 1995, 1997 and 1998. This is due to a wetter summer than average (1995: June & July precipitation was 292 mm, instead of 83 mm), a drier year than average (1997) and first a drier half year, then a wetter half year than average (1998). Also the method, either NSE (abs) or NSE (sq), shows some variation.

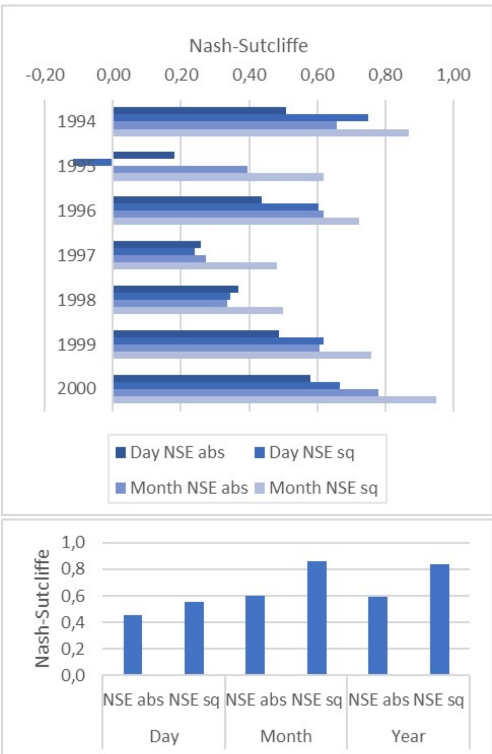

Figure 4.1: *Annual (top) and overall (bottom) Nash-Sutcliffe results of the calibration period.*

Figure 4.2 shows the annual cumulative simulated and observed runoff for the period 1994-2000, which shows that there is quite some overlap, especially in 1996, 1997 and 2000. However, DRAINMOD underestimates the runoff in summers in general, and overestimates the spring snowmelt.

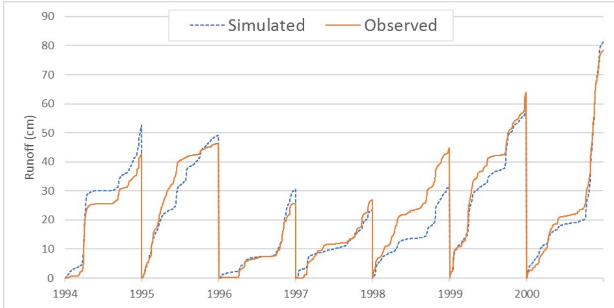

Figure 4.2: *Simulated and observed cumulative runoff over the calibration period.*

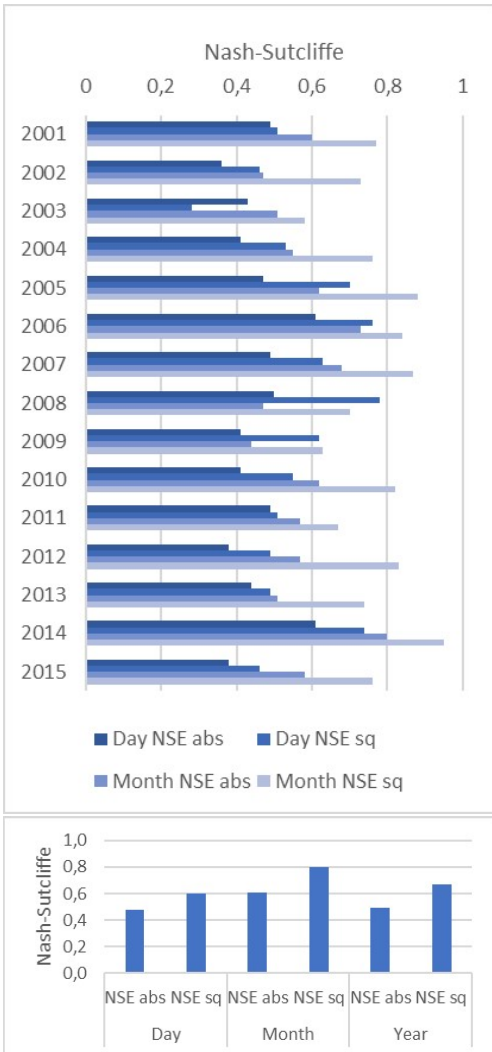

Figure 4.3: *Annual (top) and overall (bottom) Nash-Sutcliffe results of the validation period.*

4.4 Validation result

Validation has been done by taking 2000 as warmup year, and simulating up to 2015, keeping the parameter values the same as during calibration. Nash-Sutcliffe results are acceptable (Figure 4.3), although a few years have lower values, due to a wetter than average summer (2002, 2008 and 2009), a drier year than normal (2003) and wetter second half of the year (20012 and 2015). Figure 4.4 shows the simulated and observed cumulative runoff from 2001 to 2015. The model does not accurately predict the summer and autumn runoff, and either underestimates the spring runoff or is similar to the observed runoff.

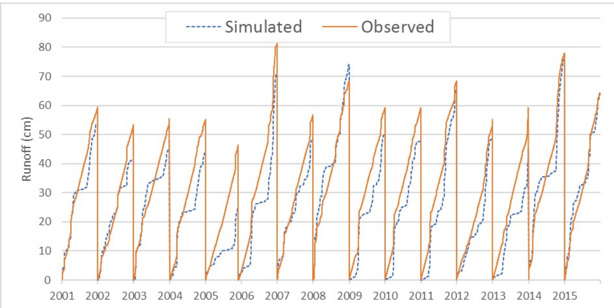

Figure 4.4: *Simulated and observed cumulative runoff over the validation period.*

4.5 More detailed analysis

Figure 4.5 shows the annual runoff. The years 1994 to 2000 are from the calibration run, 2001 to 2015 from the validation. There is a lot of variety between years, with 151 mm being the largest (2011) and 1 mm being the smallest (2014). DRAINMOD both under- and overestimated the runoff, however when only considering the averages the model underestimated with only 16 mm (503 instead of 519 mm).

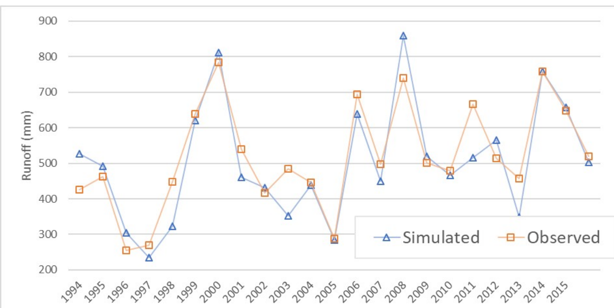

Figure 4.5: *Annual simulated and observed runoff, 1994-2015. The calibrated and validated results are merged into one figure.*

Based on literature, of the total runoff, 80% is drainage. On average, the whole catchment has 73%, for the whole period (Figure 4.6). Rk, He and Je have just over 80% (81-83%), whilst Fo has 51%, over the whole period.

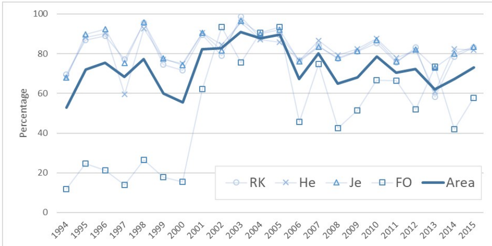

Figure 4.6: *Annual simulated percentage of drainage, compared to the total runoff, 1994-2015, per different soil type. Again, calibrated and validated results are merged into one figure.*

Literature also showed that hardly any surface runoff occurs in the summer, and April has a peak due to snowmelt. This can be seen in Figure 4.7, which shows the average surface runoff and drainage of the simulated years. There is hardly any surface runoff in the summer period, and there is a peak in April.

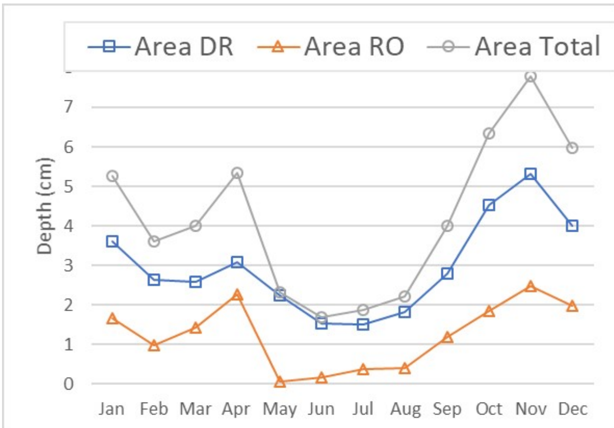

Figure 4.7: *Simulated monthly average drainage and surface runoff for the whole area, and total runoff.*

To see even more detail, both the wettest and driest years will be examined, as will the most extreme discharge event. Next to this the coldest winter and warmest summer will be taken. These events are shown below.

Wettest year: 2000

The year 2000 is the wettest year from this data, with 1200 mm over the whole year. Figure 4.8 shows the hydrograph this year for both simulated and ob-

Table 4.3: Monthly percentage of simulated and observed runoff, and precipitation, for the wettest year (2000).

Month	Simulated	Observed	Precipitation
Jan	7	4	3
Feb	4	3	4
Mar	4	5	3
Apr	5	9	7
May	2	4	7
Jun	1	1	4
Jul	1	1	7
Aug	0	1	6
Sep	1	2	5
Oct	23	22	19
Nov	38	35	26
Dec	14	13	9

Table 4.4: Monthly percentage of simulated and observed runoff, and precipitation, for the driest year (1997).

Month	Simulated	Observed	Precipitation
Jan	13	0	2
Feb	14	19	11
Mar	8	9	2
Apr	2	8	1
May	3	8	9
Jun	2	1	8
Jul	3	1	8
Aug	4	1	10
Sep	10	5	13
Oct	11	14	13
Nov	15	19	12
Dec	15	17	11

served, which are comparable. Most of the runoff occurs in the last few months of the year, which is also when most of the precipitation falls (Table 4.3). In the summer there is also precipitation, although the catchment hardly responds to it, as runoff remains relatively the same. The sudden rise in runoff in October is because the soil is saturated from the previous precipitation events and evapotranspiration, meaning more water stays in the soil.

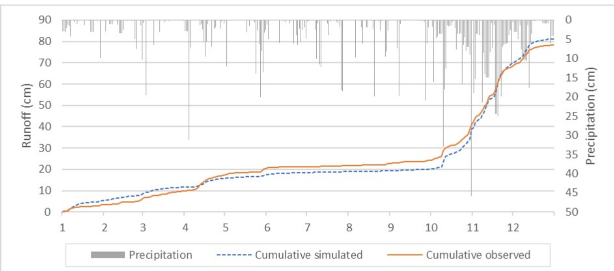

Figure 4.8: Cumulative simulated and observed runoff for the wettest year (2000).

Driest year: 1997

The year 1997 is the driest year, with only 651mm precipitation in total. This is nearly half that of the wettest year, resulting in also half the amount of runoff, the rest being evapotranspiration. The hydrograph in Figure 4.9 is similar to that of the year 2000, in that there is a peak in spring, then a flat summer and a rise again at the end of the year, although less steep. What can be noticed is that there are more step like rises in this dry year, which only occur after several precipitation events. Single events hardly influence the runoff, which is due to the available storage in the soil. Single precipitation events are quickly used by vegetation and evaporates. There is a difference between simulated and observed runoff in the nearly all seasons. The runoff in the first months is overestimated by DRAINMOD, meaning the temperature settings for winter may not be optimal. In the summer, the runoff is overestimated, as there is a rise in the simulated runoff and not in the observed runoff. The last months show an underestimation again, after a month of agreeable results in September. This could still be due to the evapotranspiration being higher than it should. Table 4.4 summarizes how much percent of the runoff or precipitation occurs per month. Compared to the wettest year, 1997 more runoff occurs in February and less in autumn.

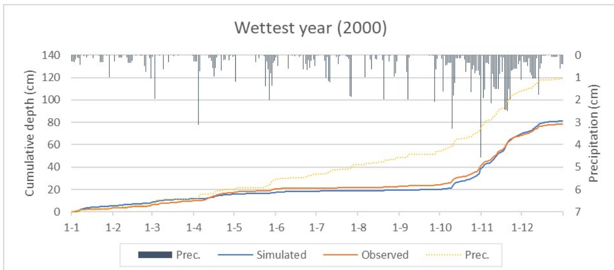

Figure 4.9: Cumulative simulated and observed runoff for the driest year (1997).

Extreme runoff: 16-1-2008

The most extreme runoff event occurred on 16-1-2008, with 39.3 mm (Figure 4.10). When more precipitation falls, runoff increases. Two minor peaks have been missed however, on the 11th and 19th of January.

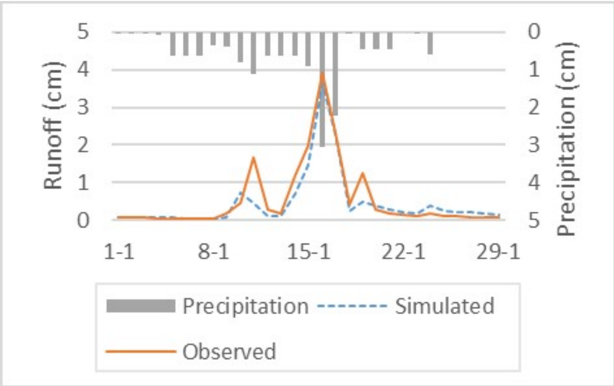

Figure 4.10: Hydrograph of the days around most extreme runoff event (16-1-2008), both simulated and observed.

Coldest winter: December 2009 - February 2010

Figure 4.11 shows the hydrograph and precipitation from December 2009 to February 2010, and the minimum, maximum and mean temperature of that same period. Most of the time, the temperatures are below 0 °C, except for the beginning of December, which resulted in some snowmelt or precipitation fell in the form of rain, meaning it could flow and leave the catchment as runoff. The rest of the year al precipitation fell as snow, and no drainage or runoff was possible because already present water is frozen, except in the deeper layers.

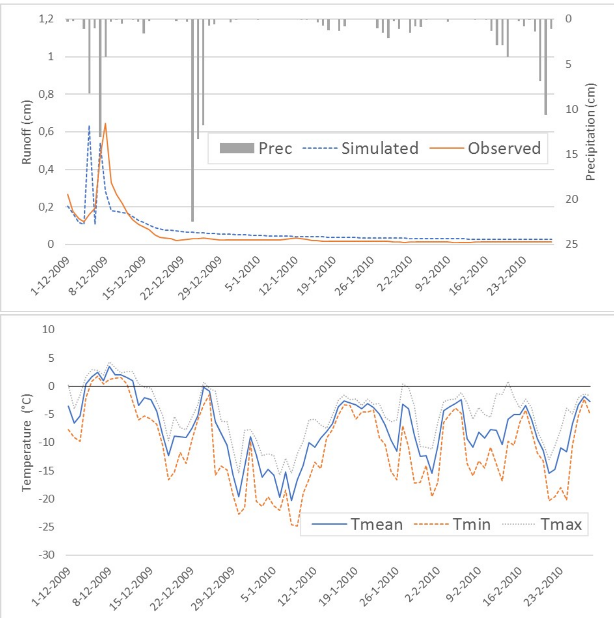

Figure 4.11: Hydrograph showing the simulated and observed runoff of the winter of 2009-2010 (top) and temperature graph of the winter of 2009-2010, showing minimum, average and maximum daily values (bottom).

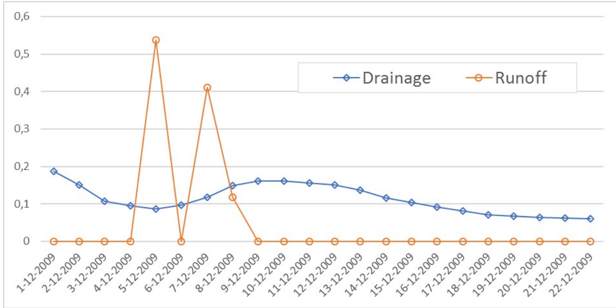

Figure 4.12: Simulated drainage and surface runoff of the winter of 2009-2010.

The simulated hydrograph of the first few days of December (Figure 4.12) show that when temperatures were above 0 °C (5 - 11 December) the contribution of surface runoff increased, and that of drainage decreased. As soon as the temperatures dropped again, only drainage contributed to the total runoff.

5 | Numerical experiments

This chapter will show the results of the numerical experiments of the following parameters: the surface storage, drainage coefficient, drain spacing, lateral saturated hydraulic conductivity and threshold temperatures. Instead of only showing yearly values or averages, certain events will be highlighted: both the wettest and driest years will be examined, as will the most extreme discharge event. Next to this the coldest winter and warmest summer will be taken. The calibration and validation chapter has already shown these events, and compared the model results with the observed values. This chapter will only compare simulated results with each other, assuming the calibrated and validated results are correct.

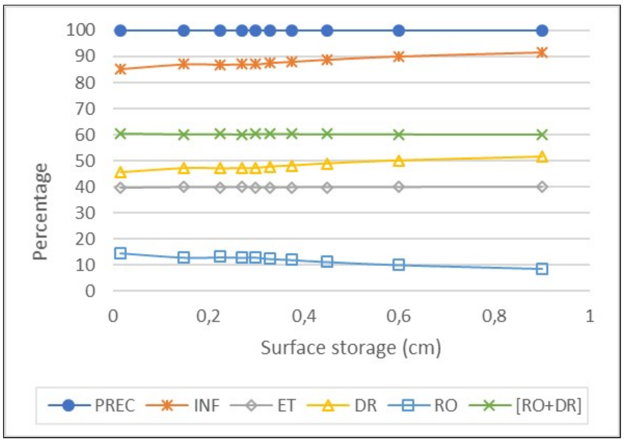

Figure 5.1: *Partitioning of water for different surface storage values (cm).*

5.1 Surface storage

As mentioned in chapter 1.2.5, the slope can be incorporated into other numbers, in this case the surface storage. Higher values mean flatter slopes, whilst lower values mean steeper slopes. For the numerical experiments, the following values will be used, all derived from the calibration value 0.3 cm: *0.015, 0.15, 0.225, 0.27, 0.33, 0.375, 0.45, 0.6* and *0.9* cm.

Partitioning Figure 5.1 shows the division of water between the different fluxes in percentage. Two fluxes remain the same throughout every experiment: the total runoff (=surface and subsurface combined) and the evapotranspiration. The total infiltration increases with increased surface storage, as does the contribution of drainage to the total runoff, whilst surface runoff decrease. This is as expected, as more surface storage means more water can be stored on the soil surface (ponding) before it starts to flow over the surface. At the same time, this ponded water has more time to infiltrate into the soil. Besides surface runoff occurring because of low surface storage, it can occur because the precipitation intensity is higher than the infiltration capacity (Horton overland flow). With more surface storage the effects is countered, as the overland flow cannot move away and stays on the soil surface until it can infiltrate. The ratio between subsurface and surface runoff is between 76/24 and 85/14, comparable to literature (80/20)

Wet & dry year Although the previous figure shows the main influence of surface storage on the different fluxes, it is comprised of all the years. In the wetter and drier years the catchment may respond different. Figure 5.2 shows the cumulative runoff (DR+RO) for the wettest and driest years, 2000 and 1997 respectively. In 2000 the simulation with the highest surface storage had the same runoff as the second lowest one, with the lowest one, in this case

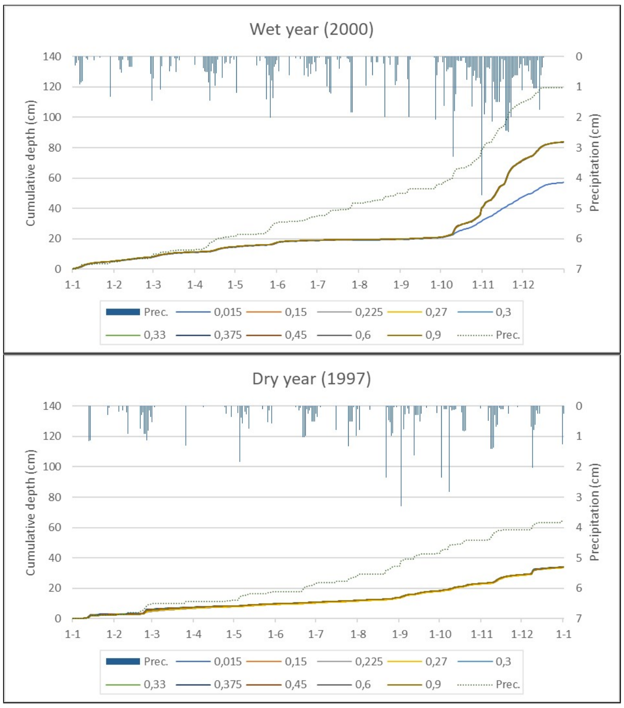

Figure 5.2: *Cumulative runoff for the wettest year and driest year, for the different surface storage values (cm).*

0.015 cm, having 27 cm less cumulative discharge. In 1997 there is hardly any difference between the different surface storages. In both hydrographs the runoff follows precipitation in the first few months. At the end of spring and during the summer the precipitation and runoff lines diverge, due to an increase in evapotranspiration. When autumn and winter arrive the runoff follows precipitation again. The difference between precipitation and runoff is approximately 35 cm for the wet year and 31 cm for the dry year at the end of the year, due to evapotranspiration. Evapotranspiration remains relatively the same, even though the input (precipitation) is nearly doubled.

Extreme runoff The extreme runoff event occurred because of the previous saturation of the soil by other precipitation events. In Figure 5.3 this can be seen, as surface runoff occurs on the 15^{th} of January, whilst drainage was already increasing from the 8^{th} onwards. The surface runoff peak is high for all surface storage values, the differences being 1 mm only, however a second, smaller peak also occurs. Here it is visible that high surface storage leads to low surface runoff, and vice versa. Although the surface runoff graph shows many similarities between the different simulations, the drainage simulations are very different. The only logical lines the ones with the highest and lowest surface storage, which are the two boundaries. The other lines do not behave as expected, namely less drainage with lower surface storage. But still, in general the rend is that higher surface storage values lead to higher drainage values

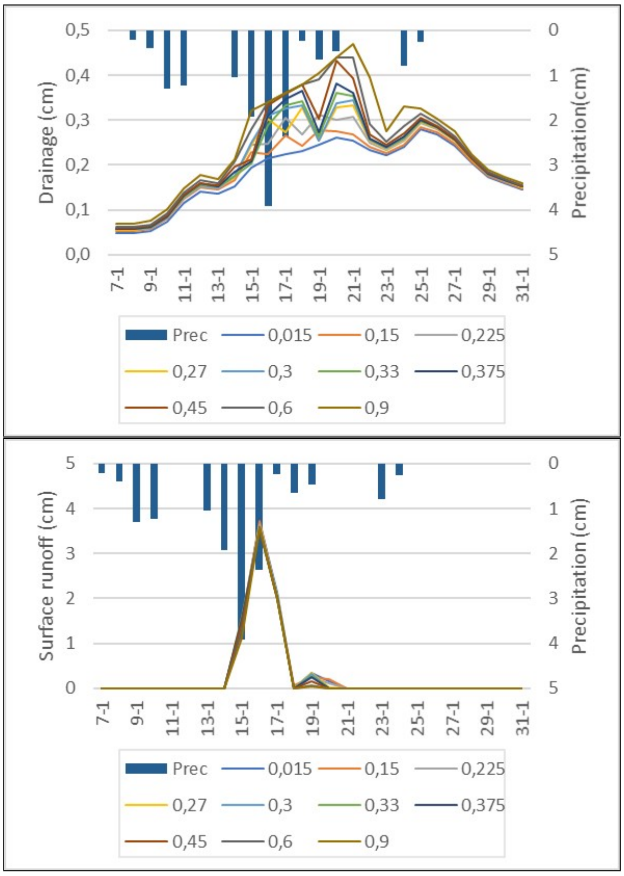

Figure 5.3: Hydrograph showing the drainage (top) and surface runoff (bottom) results of the highest runoff event (January 2008), for different surface storage values (cm).

5.2 Drainage coefficient

The following values will be used for the drainage coefficient: *0.05, 0.5, 0.75, 0.9, 1, 1.1, 1.25, 1.5, 2* and *3* cm d^{-1}.

Partitioning Figure 5.4 shows the ratio between the different fluxes, confirming that there is little difference after DC = 1 cm d^{-1}. With a low DC evapotranspiration goes up however, because water is stored in the soil longer before it can be drained away or used by plants. This can be seen in Figure 5.6, which shows the summer of 1997. With DC = 0.05 cm d^{-1} there is more drainage during the summer, until the rains start in September again (which is why there are peaks). The higher DC values have less drainage during these months. At the same time the figure shows that a value of 0.05 cm d^{-1} looks like, as that is where the drainage stops. The figure also shows that 60% of the water leaves the catchment as either drainage or surface runoff. Of that, 78% is drainage and 22% is surface runoff. The other 40% is evapotranspiration. This is not true for the DC = 0.05 cm d^{-1}, which has higher evapotranspiration and surface runoff, and lower drainage.

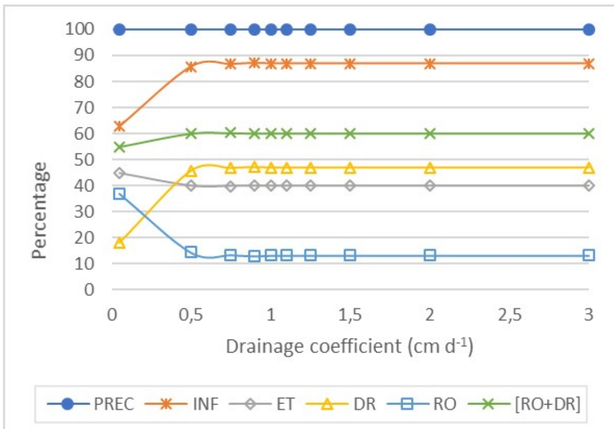

Figure 5.4: Partitioning of water for different drainage coefficient values (cm d^{-1}).

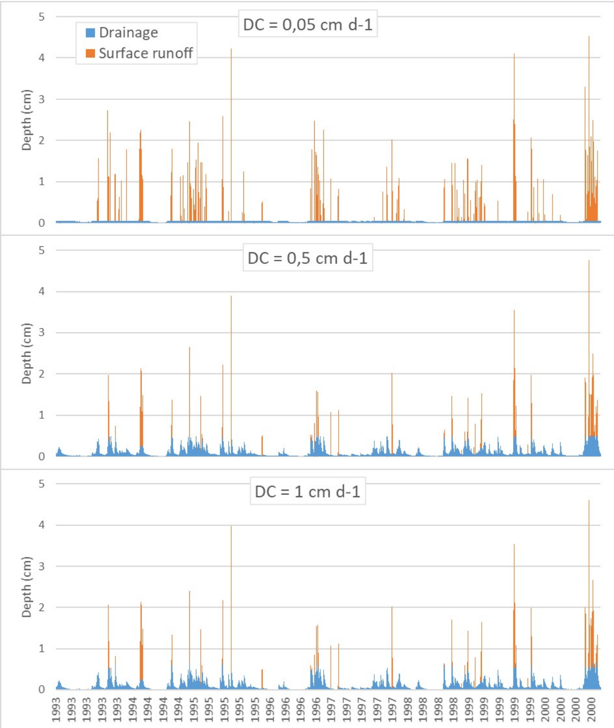

Figure 5.5: Hydrograph showing drainage (dr), surface runoff (ro) and precipitation (prec) for the wettest year (2000), for different drainage coefficients: 0.05, 0.5 and 1 cm d^{-1}.

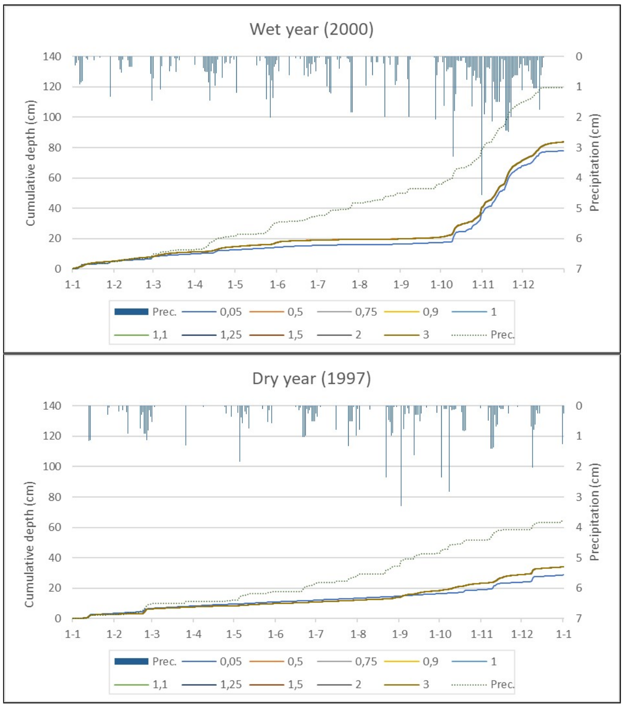

Figure 5.6: Cumulative runoff for the wettest and driest year, for the different drainage coefficient values (cm d^{-1}).

Wet & dry year Similar to the surface storage graph (Figure 5.2), Figure 5.6 also shows that runoff follows precipitation. In both years the lowest DC has the lowest runoff, whilst the other DC values are the same. This was also confirmed in Figure 5.5. The difference between the lowest DC and the others is 6 cm for the wet year and 5 cm for the dry year, quite similar. The evapotranspiration is 35 cm for the wet year and 31 cm for the dry year, for the higher DC values from 0.5 cm d^{-1}.

Figure 5.5 shows the drainage and runoff for the year 2000, for three different values of DC, to illustrate the effect that the drainage coefficient has on the partitioning of water flows. The other DC values are similar to 1 cm d^{-1} graph. The most difference is visible between 0.05 and 0.5 cm d^{-1}, where most of the precipitation leads to surface runoff for the former, whilst with the other DC values surface runoff only occurs at the end of the year, such as with the latter. What is also visible from the figure is that there is a threshold of drainage that needs to be met, before surface runoff can occur, namely the DC, however as the 1 cm d^{-1} show, after a certain value there is a different limiting factor, as the daily drainage in that same graph is not equal to DC.

Extreme runoff Figure 5.7 shows the drainage and surface runoff for the extreme event. The maximum daily drainage is around 0.35 cm, which is higher than the lowest DC, but still lower than the second lowest DC. This explains why all simulations are similar except for the lowest DC simulation. The maximum surface runoff is much higher than the maximum drainage, by a factor 10, for all DC values, even the lowest, although that one has more lower peaks. This means that there is another factor more limiting to drainage than the drainage coefficient.

5.3 Drain spacing

Simulations have been done with drain spacings of *50, 500, 750, 900, 1000, 1100, 1250, 1500, 2000* and *3000* cm.

Partitioning Figure 5.8 shows the general distribution between flows. As can be expected, when the drains are closer to each other more drainage occurs than with large drain spacings, whilst for surface runoff this is the other way around. Evapotranspiration also increases when drains are spaced further apart, because the infiltrated water in the soil stays in the soil longer, making it available for infiltration. As ET increases, the total amount of runoff has to decrease. What is noticeable is that a linear

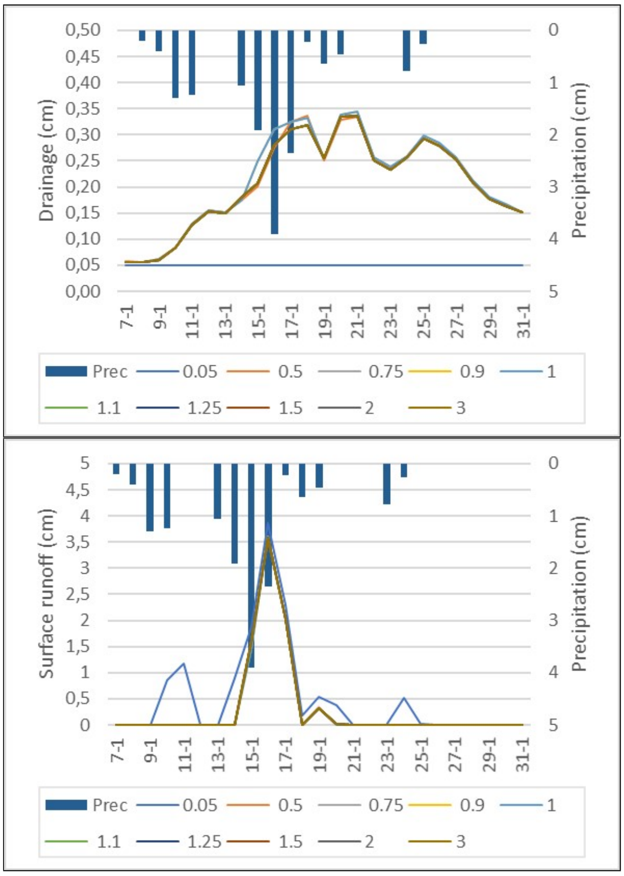

Figure 5.7: Hydrograph showing the surface runoff results of the highest drainage (top) and surface runoff (bottom) (January 2008), for different drainage coefficient values (cm d^{-1}).

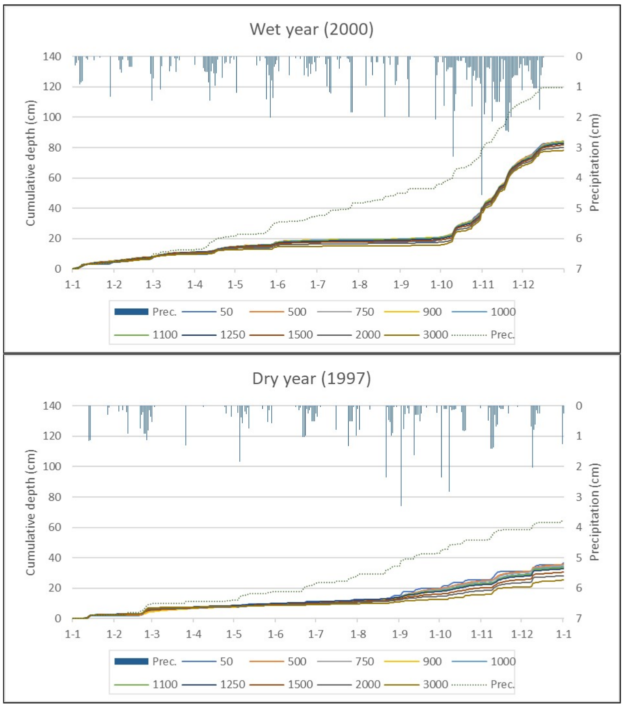

Figure 5.9: Cumulative runoff for the wettest and driest year, for the different drain spacing values (cm).

line can be plotted through the points, and that in principle a required percentage of annual drainage can be translated into the best drain spacing.

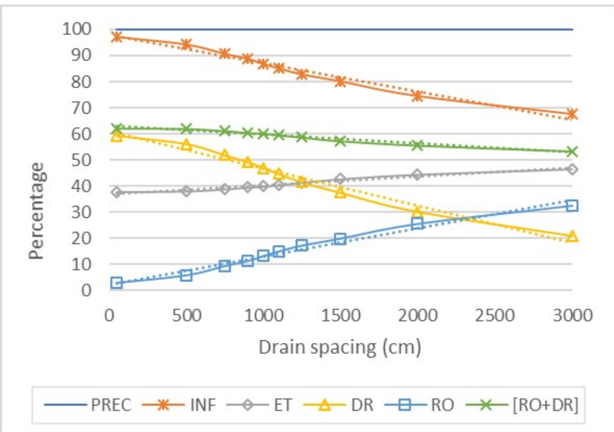

Figure 5.8: Partitioning of water for different drain spacing values (cm).

Wet & dry year Figure 5.9 shows the cumulative runoff and precipitation for a wet and dry year. Similar to the surface storage and drainage coefficient simulations, runoff follows precipitation quite well, with the summer creating a divergent. For the wet year, on average the is 36 cm less runoff than precipitation, for the dry year this is 32 cm. There are not many differences between the cumulative runoffs when changing the drain spacing in wet years, which we have seen in the previous Figure as well. However for dry years, the results are different. In dry years the spacing matters more, with the smallest spacing having the highest runoff, as can be expected. The difference between 50 cm and 3000 cm at the end of the year 10 cm runoff, which seems little but is almost 40% of the total runoff for the 3000 cm simulation, and 28 % for the 50 cm.

Extreme runoff Drain spacing has a big effect on the distribution of water between drainage or surface runoff, as can be seen in Figure 5.10 which shows the extreme runoff event. The smallest drain spacing, 50 cm, has the most drainage, and a faster response to the precipitation than the drains that are spaced further apart. The higher and steeper peaks are proof of this. At 30 m drainage is almost stable throughout this period, and additional precipitation is hardly reflected in peaks. The drainage for the other spacing values responds as expected, with smaller spacings having higher drainage peaks and steeper slopes. The opposite is visible at for the surface runoff, which is lowest for the 50 cm spacing and highest for the 3000 cm. Not all spacings have the same peaks. The ones furthest apart have more and higher peaks. Surface runoff is also bigger than drainage for most spacings (ranging from 2 to 8 times more), only the 50 cm spacing has higher drainage. What is remarkable is that even with spacings at a very short distance, the maximum surface runoff peak is still 3 to 15 times higher than that of drainage. This indicates that, similar to the previous experiments, there is another factor influencing the maximum drainage, although the drain spacing already influences it a lot.

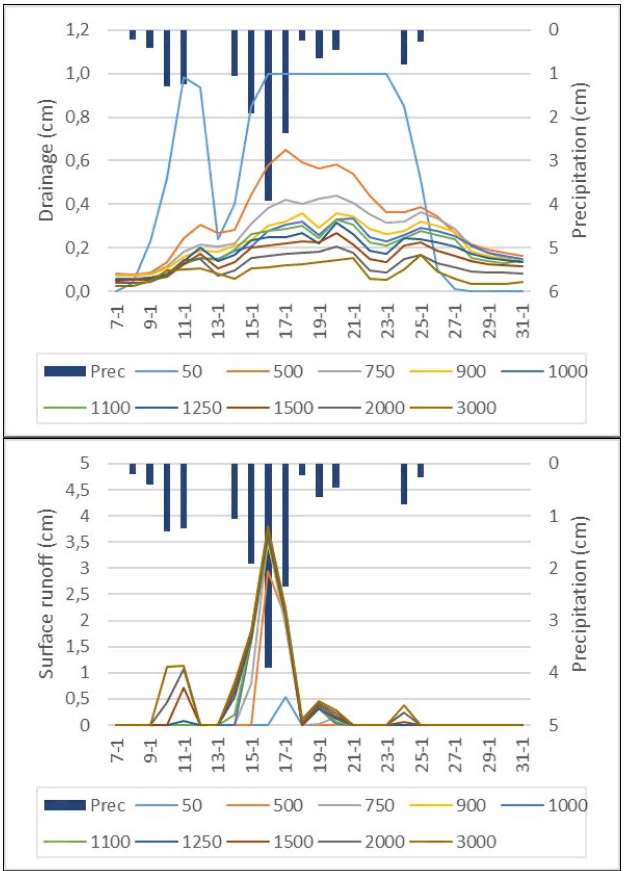

Figure 5.10: Hydrograph showing the surface runoff results of the highest drainage (top) and surface runoff (January 2008, bottom), for different drain spacing values (cm).

5.4 Lateral saturated hydraulic conductivity

The lateral hydraulic conductivity is not actually measured in situ, but assumed from the vertical conductivity. The used values during calibration for layers 1, 2, 3 and 4 were *1, 1, 0.1* and *0.1* cm h^{-1}. For the numerical experiments, per layer the following conductivities will be used: *0.01, 0.05, 0.1, 0.5, 1, 1.5, 2, 3, 5* and *10* cm h^{-1}. What this means is that per simulated layer, the other three layers will remain at their default value. The results of the four layers will be compared together.

Partitioning Figure 5.11 shows the partitioning of the water in percentages, when taking different hydrological conductivities per specific layer. All layers have similar results, thus only the fourth layer is shown. The bottom figure shows the same results but with a logarithmic scale. There seem to me two noticeable trends, namely between 0.01 and 1 cm h^{-1} , and from 1 cm h^{-1} to higher conductivity values. Starting with the latter, after 1 cm h^{-1} there is little variation in the partitioning. Surface runoff is low whilst drainage is high, when comparing to other experiments. The first trend is more interesting. The influence of the hydraulic conductivity is apparent here. Lower values means there is more resistance in the soil, in turn leading to lower drainage and higher surface runoff values.

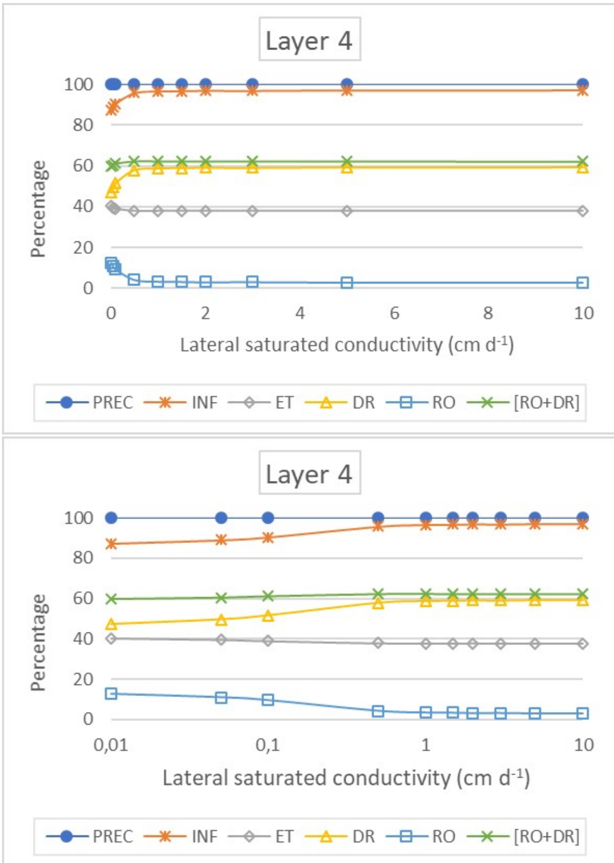

Figure 5.11: Partitioning of water for different lateral saturated conductivity values of layer 4 (cm h^{-1}), with normal and logarithmic scale.

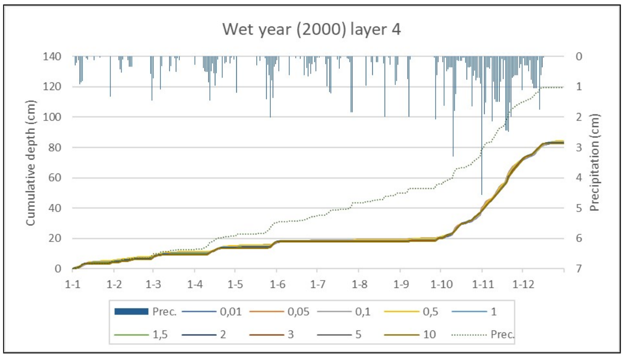

Figure 5.12: *Cumulative runoff for the wettest year (2000), for the different lateral saturated conductivity values (cm h^{-1}).*

Wet & year The cumulative runoff graphs are all similar for the different layers, thus only one will be shown for illustration purposes, namely layer 4 as it shows the most variation (Figure 5.12). As seen before, runoff and precipitation diverge during summer, with the eventual difference being around 36 cm. The simulated runoffs are nearly all the same. The cumulative graphs for the dry years are more interesting (Figure 5.13). When going deeper into the soil the total runoff graphs separate when using different conductivity values. In layer 1 there are hardly differences, whilst layer 4 has a difference of 4 cm. These differences occur mostly during autumn, when soil saturation plays a bigger role. However other than this not a lot can be concluded from these graphs.

Exreme runoff As with other the experiments, the higher conductivity values lead to more drainage and less surface runoff. Figures 5.14 and 5.15, show the hydrographs of the extreme event, per layer. First the difference in drainage between layers will be handled. The major difference is that when going deeper into the soil, the hydraulic conductivity plays a bigger role and increases the drainage. Ceilings of 1 cm d^{-1} are reached, equal to the drainage coefficient, in layers 3 and 4 whilst layers 1 and 2 do not result in this ceiling. During these days there are three separate precipitation events. The first event only causes a slight rise in drainage for layer 1, whilst the peak height increases when going deeper into the soil. When the second event occurs all layers show a big increase in drainage, however the peak is narrowest for layer 1 and broadest for layer 4, which even encompasses the third event. For the other layers this third event results in a slight increase in drainage. Even though layer 4 shows the biggest response to precipitation, within that layer

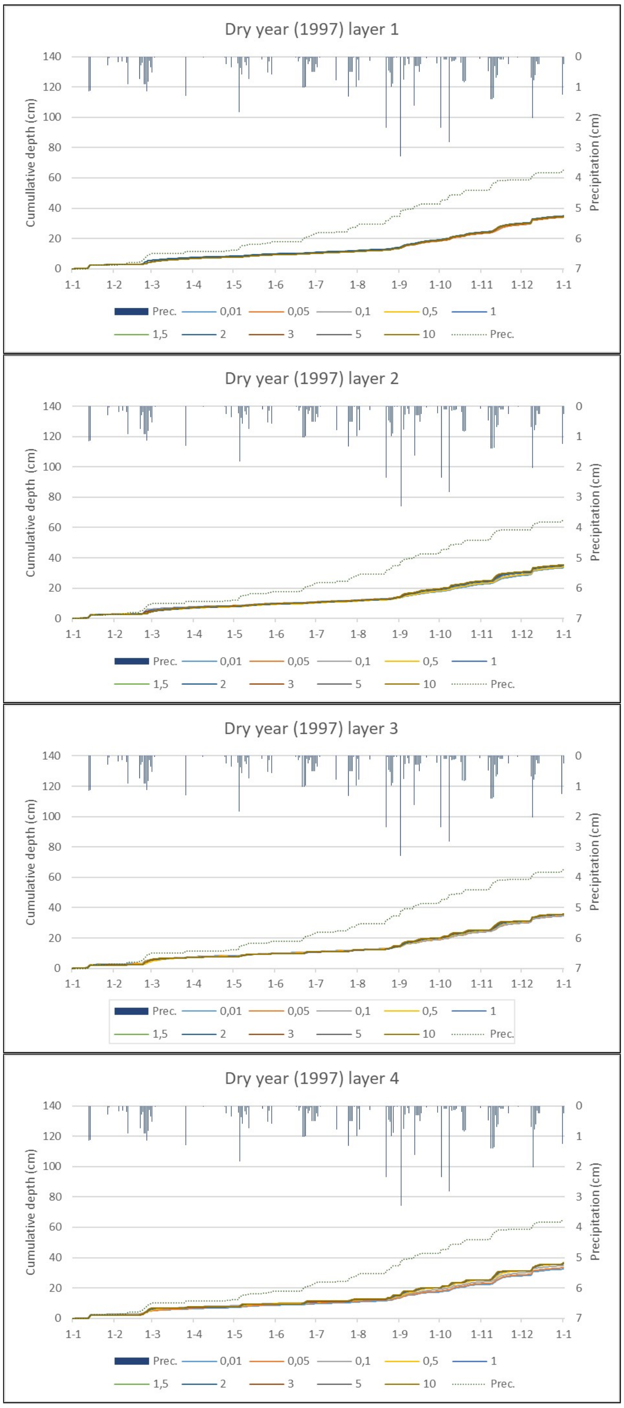

Figure 5.13: *Cumulative runoff for the driest year (1997), for the different lateral saturated conductivity values of all layers layer (cm).*

there are big differences when considering the hydraulic conductivity itself. Comparing 0.01 to 10 cm d^{-1} shows that the former has approximately 3 to 5 times more drainage. Drainage also decrease to 0 cm within a day after the last precipitation event for 10 cm d^{-1} conductivity, whilst the 0.01 cm d^{-1} takes another few days. The other hydraulic conductivity simulations show peaks and regressions in-between

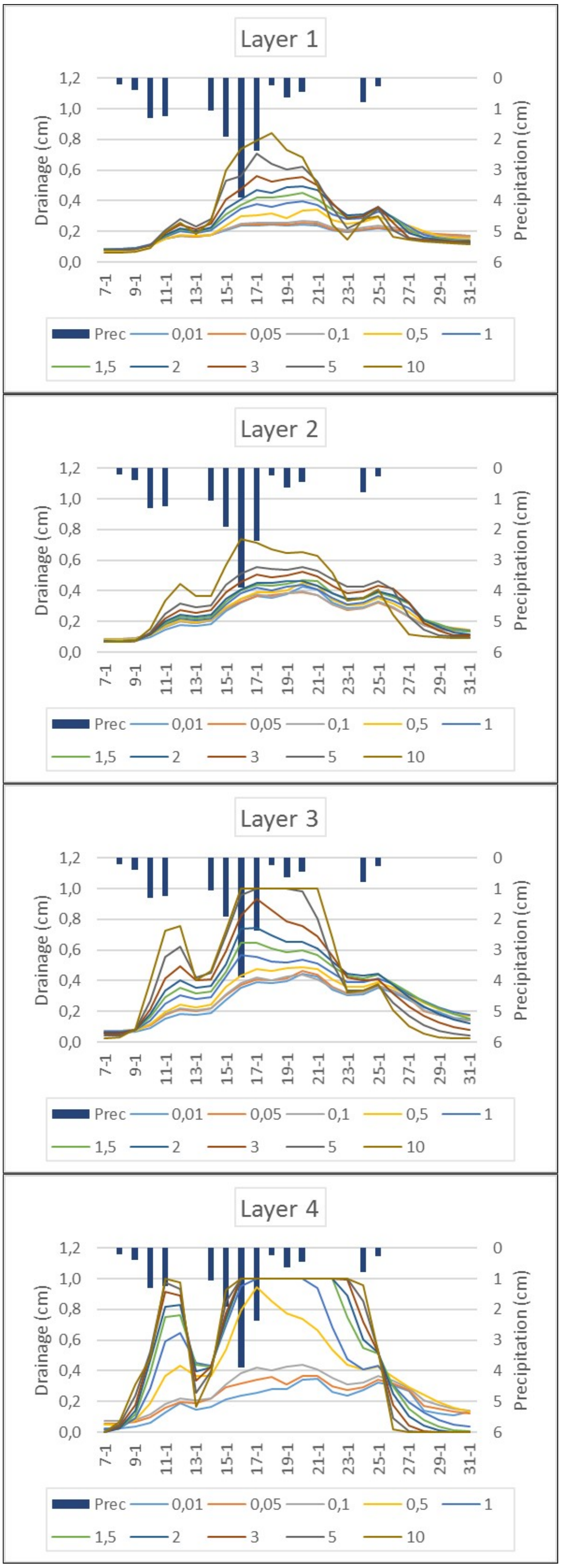

Figure 5.14: Hydrograph showing the drainage results of the highest runoff (January 2008), for different lateral saturated conductivity values (cm h^{-1}).

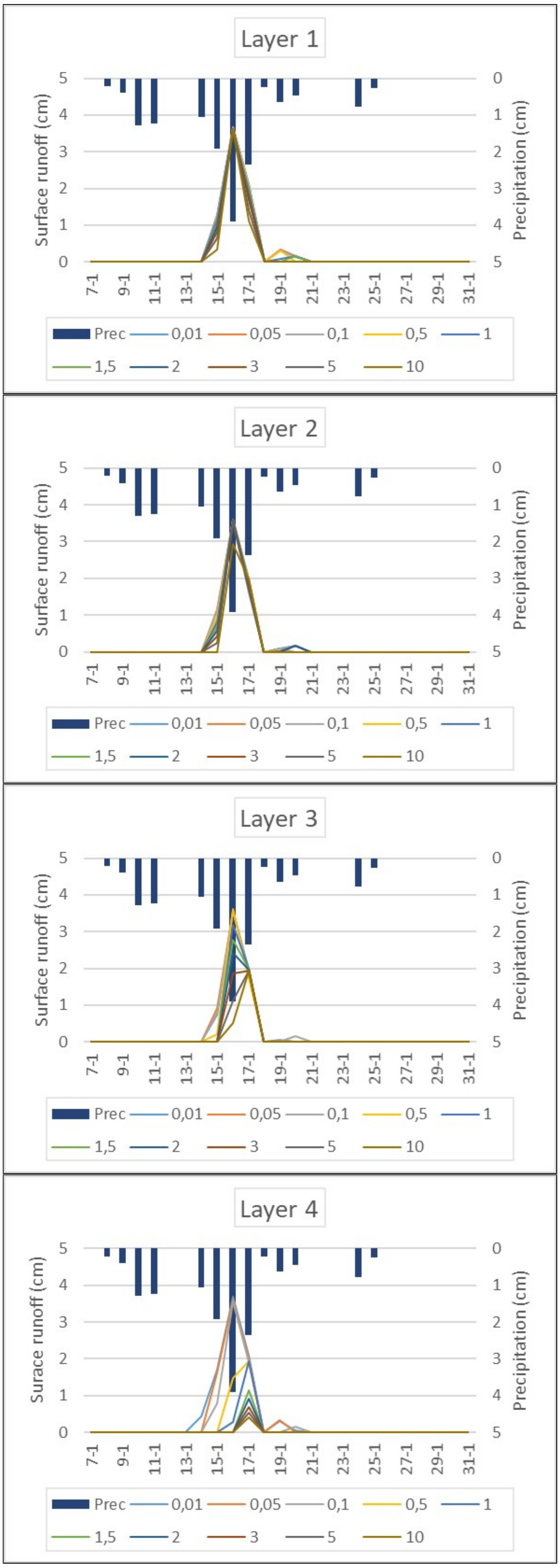

Figure 5.15: Hydrograph showing the suface runoff results of the highest runoff (January 2008), for different lateral saturated conductivity values (cm h^{-1}).

these two boundaries. For the other layers the differences between 0.01 and 10 cm d^{-1} are most visible halfway the second precipitation event, and show similar results but less intense. The difference is that the simulations of the first layer do not reach 0 cm drainage. Concerning the surface runoff, the peak of layer 1 is the highest, whilst that of layer 4 is the lowest. Layer 1 also shows that all simulations have similar surface runoff, whilst in layer 4 there are differences between the conductivities. The highest conductivity has both a lower and narrower peak. Compared to the other experiments, adjusting the hydraulic conductivity has a lot of effect on the partitioning of water, until the DC limits the maximum drainage.

5.5 Threshold temperatures

Within DRAINMOD there are four temperature settings which will be tested. The rain-snow dividing temperature and snow melt temperature will be simulated in temperature steps of *0.5* °C, from *-1* to *3* °C. The snowmelt coefficient will have the same steps, however starting from *0* °C mm^{-1}. The critical ice content is simulated in steps op *0.1* ^{3} cm $^{-3}$ starting from *0* to *0.9* cm^3 cm^{-3}.

Partitioning On the overall scale, very little differences in partitioning are visible (Figure 5.16). For the rain-snow partitioning and snowmelt temperatures, the higher threshold temperature result in more surface runoff, however only in the order of a few centimeters. The snowmelt coefficient threshold shows more difference between 0 and 0.5 °C mm^{-1}, however after that changes are relatively small. Surface runoff and drainage both increase a little, whilst evapotranspiration decreases. The reason that there is a big jump from 0 to 0.5 °C mm^{-1} is that snow cannot melt with 0 °C mm^{-1}, instead it all evaporates, hence the high evapotranspiration value. The critical ice content also has similar partitioning, however at 0 $cm^3 cm^{-3}$ there is no infiltration when it freezes, which is why the infiltrated precipitation is lower and surface runoff is higher than with the ice content values. All in all not much can be said from these graphs. Hence the winter of 2009-2010 will be shown.

Cold winter Figure 5.17 shows the drainage hydrographs of these different simulations, for the winter of 2009-2010, which was the coldest of the measured period. All are quite similar in that they start with a regression until the precipitation occurs, after

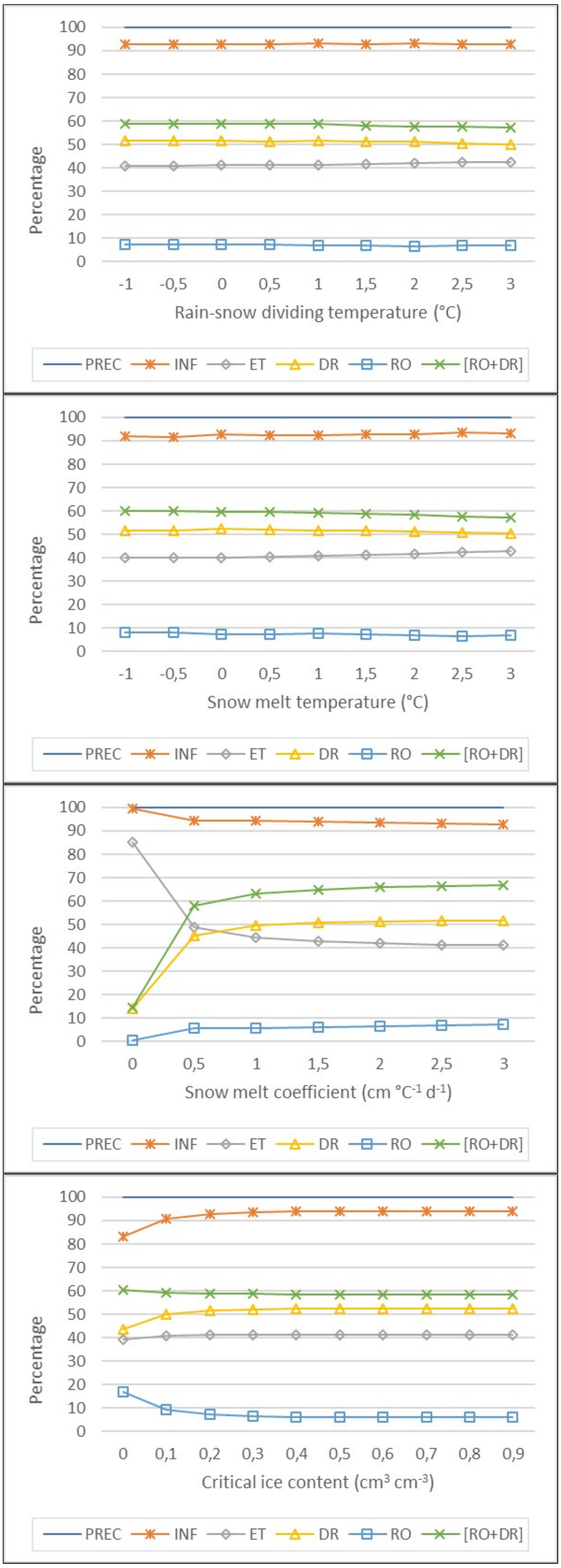

Figure 5.16: Partitioning of water for different threshold temperature related simulations.

which a bigger peak occurs. After this a long period of regression (two months) without precipitation ensues.

Rain-snow dividing temperature In the rain-snow dividing temperature graph, the biggest peak arrives on the 9^{th} and 10^{th} of December, depending on the temperature threshold (Figure 5.17). The simulated temperatures -1 to 1.5 °C have both a higher peak and occur on the 9^{th} of December, with approximately 0.3 cm drainage, whilst the temperatures 2 to 3 °C occur on the 10^{th} of Decmber and have only 0.11 to 0.15 cm drainage. During these days the temperature was around 1.5 to 3.3 °C, which explains the differences in the peaks. Simulating with 3 °C as threshold temperature means only on these days precipitation can fall as rain, whilst a threshold of 0 °C means much more precipitation falls as rain as there are more days with temperatures above 0 than 3 °C. Although lower threshold temperatures mean more drainage, surface runoff also increases. There is only one day of simulated surface runoff, namely on the 5^{th} of December. Between 0.46 and 0.6 cm of surface runoff occurred, while 0.8 cm of precipitation fell. This is also the first day that temperatures are above freezing, resulting in additional snow melt. What also causes melt is warm rain falling on snow. However the peculiar thing is that simulations with threshold temperatures of 1.5 to 3 °C have no surface runoff at all. Thus these thresholds are too high, resulting in snowfall instead of rainfall, even though the temperatures are above freezing.

Snowmelt temperature The effect of the snowmelt temperature threshold is also quite straight forward, lower temperatures meaning more snowmelt (Figure 5.17). As mentioned earlier, the days of the peak drainage occurred when temperatures were slightly above freezing. With a snow melting at temperatures above -1 °C more days are present for snow to melt than when 3 °C is the lower limit. The peaks range from 0.25 to 0.31 cm, however unlike the snow-rain temperature graph, the regression of the snowmelt results are all very similar. The surface runoff peak is high as well, ranging from 0.46 to 0.6 cm, and a second peak is present on the 8^{th}, but only for the temperatures 2.5 and 3 °C, with 0.21 and 0.1 cm respectively (not shown). The reason for this is again, that around the 8^{th} of December temperatures are high enough for snow to melt, when using high threshold temperatures. Why there is no surface runoff for

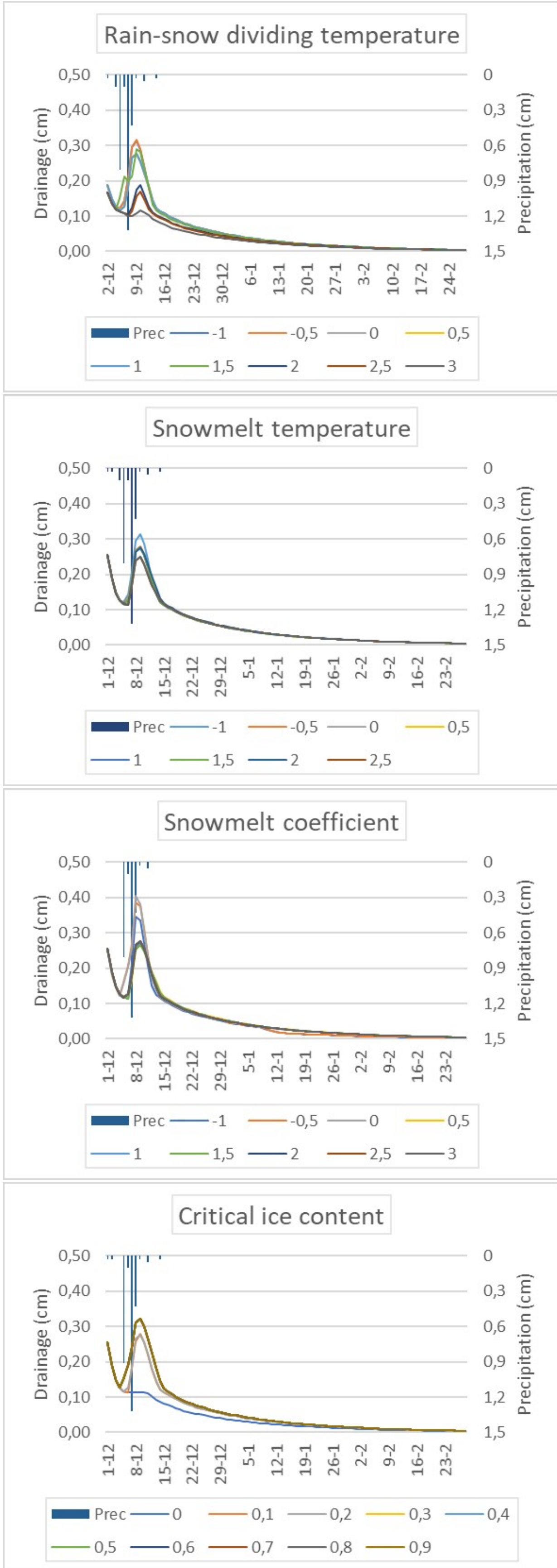

Figure 5.17: Hydrographs showing the drainage in the winter of 2009-2010, for different rain-snow dividing temperatures, snow melt temperatures, snow melt coefficients and critical ice contents.

the lower threshold temperatures is because the precipitation falls as rain instead of as snow, and infiltrates into the soil. Simply put, there is no snow to melt.

Snowmelt coefficient The melting coefficient value also behaves as expected (Figure 5.17), higher values meaning more snow melts per increase in °C. Drainage ranges between 0.27 and 0.41 cm on the 8th, and surface runoff occur on the 5th, ranging from 0.47 to 0.6, the higher values the result of high snow melt coefficients.

Critical ice content The simulations about the critical ice content above which infiltration stops, behave as expected, with higher values leading to more infiltration and drainage. There seem to be three categories, as seen in Figure 5.17: no peak (value 0 cm^3 cm^{-3}), medium peak (0.1 and 0.2 cm^3 cm^{-3}, with 0.25 cm) and a higher peak (the remaining, with 0.3 cm). Concerning the 0 cm^3 cm^{-3} simulation, if the ice content is above 0 cm^3 cm^{-3} no infiltration occurs, which is why the line has no peak. Drainage is the result of already present soil water or when the ice in the soil melts and infiltration is possible. Regarding the difference between the two peaks, the shift happens between 0.2 and 0.3 cm^3 cm^{-3}, which indicates that the latter may be too high as all precipitation infiltrates.

Additional remarks To conclude, when spring comes the differences are leveled out and the temperature thresholds do not matter any longer. Concerning evapotranspiration, it is only simulated for the first two weeks of this period with 1.02 cm, the other days have 0 cm, for all simulations. Based on these results it is difficult to know which values are suitable for further experiments, although the extremes are probably not suitable.

5.6 Evapotranspiration

Assuming deep and lateral seepage do not occur, the ET can be calculated as a result of the water balance: ET = P - RO - DR, with ET becoming the actual evapotranspiration (ET_{act}). This is done for the period starting from May 1 (start of growing season) to April 31 for 1994 to 2014, from observed data (Figure 5.18). The assumption that the growing season starts on May 1 assumes that the soil moisture storage is the same every year. Precipitation and discharge fluctuate a lot through the 21 years, as does ET_{act} (between 230 and 450 mm). These values already indicate that the ET is quite low but between the limits, comparable to literature values (see Chapter 1), which indicated 230 to 500 mm y^-1. The average ET_{act} is 350 mm, with five years having ET_{act} values higher than 400 mm and only one year having a lower ET_{act} value than 300 mm, being the 230 mm. On average the ET_{act} is 40% of the precipitation. In general the wetter years have more runoff and less ET_{act}, in percentage of precipitation, whilst drier years have less discharge and higher ET_{act}, in percentage of precipitation. Although these values are a theoretical result of the water balance and not measured, it does give an indication of the simulation ET.

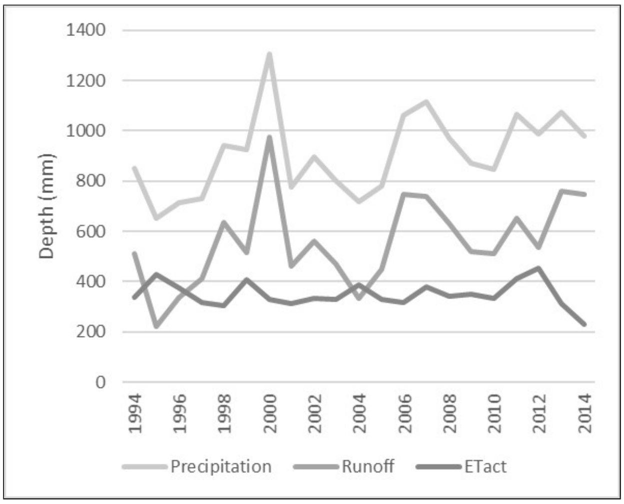

Figure 5.18: *Yearly (=May 1 to Aprill 31) ET_{act} values, compared to precipitation and discharge (mm).*

PET calculation methods Different simulations have been performed to evaluate the effect of different PET caluclation methds, being:

- Thornthwaite (*TH*);
- FAO Penman Monteith (*FAO*); and
- FAO Penman Monteith PET crop coefficients (FAO_{crop})

More simulations have been performed, this time with monthly factors which can be used to adjust the PET. The standard method of ± *10%, ±25%, ±50%, -95%, +100%* and *+200%* as monthly adjustment factors have been taken, to visualize additional adjustment effects within the weather tab in DRAINMOD.

The TH method was calculated by DRAINMOD itself, which uses daily maximum and minimum air temperature inputs. The crop adjusted factors (K_c) for FAO_{crop} are 0.25 for bare soil outside the growing season, and between 0.30-1.15 for when the crop grows, which is from May to August for wheat (Figure 5.19, Table 5.1). These K_c factors can be found in FAO manual 56 (Allen et al., 1998), and were explained in the FAO Penman Monteith section in the Introduction. With this method the PET depends on the crop and in what stage the crop is, hence the different factors for different periods. DRAINMOD has the function to adjust the root depths, which has also been done in the simulation.

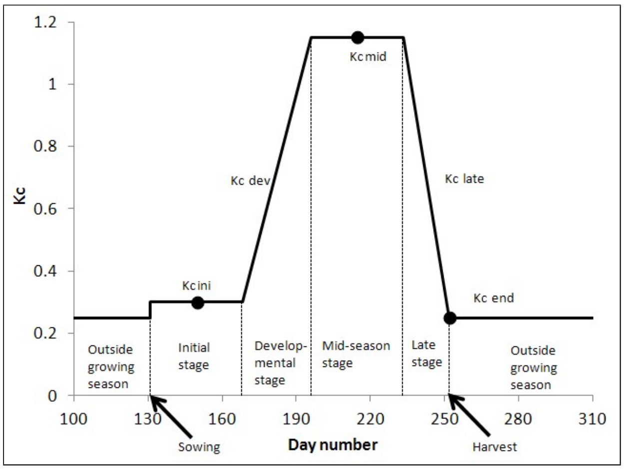

Figure 5.19: *Schematic for crop growing stages and crop coefficients.*

Table 5.1: *Crop growing factors.*

Crop factors	Value
Sowing date	1 May
Harvest date	29 Aug
Length of growing season (days)	120
daynumber sowing	121
daynumber ini-dev	136
daynumber dev-mid	161
daynumber mid-late	211
daynumber harvest	241
kcini	0.3
kcmid	1.15
kcend	0.25

Figure 5.19 shows the development of the crop-coefficient during both the growing and non-growing seasons for the calculation of the FAO_{crop}, which depend on the different crop stages: the initial stage (starting at the day of sowing), the developmental stage, the mid-season stage and the late-season stage (ending at the day of harvest). According to the Allen et al. (1998), $Kc_{initial}$, Kc_{mid} and Kc_{end} can be approximated to 0.3, 1.15 and 0.25 respectively for a situation with sowing in April and a growing season of 130 days. Kc_{dev} was determined by linear interpolation(Allen et al., 1998; Kværnø, 2014).

Table 5.2 gives a short overview of the different yearly calculated ET for the three different methods, compared to the ET_{act}, for the period May 1 to April 31. The level of detail required per method increases, resulting in a decrease in yearly ET.

Table 5.2: *Yearly simulated evapotranspiration for the three calculation methods, compared to ET_{act}*

	PET method			
Year	TH	FAO	FAO_{crop}	ET_{act}
1994	860	746	561	338
1995	1095	835	508	429
1996	956	902	518	374
1997	1025	1005	687	317
1998	1060	961	578	306
1999	1092	1047	635	409
Average	1015	916	581	362

Table 5.3 shows the average yearly (Jan - Dec) results of infiltrated precipitation (INF), drainage (DR), surface runoff (RO) and evapotranspiration (ET) for different PET calculated files. As can be expected from Table 5.2, the ET of FAO_{crop} is the lowest, followed by FAO. On the other hand, DR and RO are highest for FAO_{crop}, but INF is lower.

Table 5.3: *Average yearly (1994-2000) simulated drainage (DR), surface runoff (RO) and evapotranspiration (ET) results, for the three PET calculation methods as input for the model.*

Method (cm h^{-1})	INF (mm)	DR (mm)	RO (mm)	ET (mm)
TH	833	318	41	515
FAO	831	376	43	455
FAO_{crop}	814	528	59	288

Table 5.4: *Average yearly simulated drainage (DR), surface runoff (RO) and evapotranspiration (ET) results for the FAO PET calculation method, with monthly adjustment factors.*

Monthly factor	INF (mm)	DR (mm)	RO (mm)	ET (mm)
0,05	850	824	24	25
0,5	854	608	20	245
0,75	855	493	18	363
0,9	856	438	17	418
1	861	409	13	452
1,1	861	379	13	482
1,25	860	339	14	520
1,5	861	304	12	557
2	864	264	9	597
3	865	208	9	656

Monthly adjustment factor Next to the comparison between methods the influence of the monthly factor function within DRAINMOD was taken into consideration. Skaggs et al. (2012) recommends not adjusting the monthly value more than ±0.15 from the original value, and seeing the results, monthly factor 0.9 has more effect on the distribution of water than 1.1. The monthly factor does not influence the INF or RI, however DR and ET are highly affected. Doubling the adjustment factor does not double the ET or decrease the DR by half, although a decreasing monthly factor has more effect on decreasing the ET than increasing the monthly factor. For example, monthly factor 0.5 shows that ET is 207 mm less than the default 1, and factor 1.5 has 105 mm more than the default 1.

Forest PET An issue that has come past a few times so far, regarding PET, is that of the forest. No actual measurements have been done in the forest, and taking the forest into account with calibration is difficult but needs to be done. Some data is available on the outflow of the forest, but not detailed enough. The forest covers 129 ha (30%) of the catchment thus it certainly has some influence on the hydrology. Next to using a different PET file for the forest simulations, which is higher than the agricultural PET, a drainage system design with wide spacing will be included to simulate the forest hydrology. Not all precipitation in the forest evapotranspirates, thus some outflow is present, hence the drainage system design.

Conclusion The PET of TH is highest, followed by FAO and last the FAO_{crop} method. FAO_{crop} also comes closest to ET_{act}, however it is still more than 100 mm to high. The average simulated ET is high for TH and FAO_{crop} (455 and 515 mm, respectively), as literature values indicate from 250-400 mm per year. Based on this the FAO_{crop} PET file will be used. During all the past experiments in this chapter, the ET was similar to the found results here, with approximatley 310 to 350 mm.

6 | Discussion

This chapter summarizes the main discussion points: the available data, assumptions made, method & model, and the results.

Available data Three sets of observed data were available for this thesis, namely measured soil physical characteristics (pF curve and hydraulic conductivity) and the observed precipitation and runoff. First is the soil data. The hydraulic conductivity was measured in the horizontal direction, giving the vertical conductivity. The lateral conductivity was not measured in the field however, so assumed is that the lateral conductivity is lower than the vertical. Other field measurements included the pF curve. The measured values, however, may be too high. A conductivity of 40 cm h^{-1} has been found, and although possible, seems too high for a heavy clay. Secondly, the meteorological station is located some kilometers away from the catchment, and on the other side of the hill. This means uncertainties in the precipitation measurements might be present, including effects of wind direction and elevation, but this is not taken into consideration. However, as the catchment is relative small, differences in local precipitation may not be big either. Lastly, the observed discharge was for the whole catchment, which includes the urban area. For the comparison of the model results, the urban discharge was subtracted from the observed discharge, assuming that water does not infiltrate or flow elsewhere then the catchment outlet. This was done to make simulations easier, otherwise an additional soil with a low hydrological conductivity had to be included to resemble impermeable surfaces and small patches of grass. This would include many assumptions, thus this was a more reliable method. Another key factor regarding the runoff is that only the total daily runoff was observed, whilst the model simulates drainage and surface runoff. As the results showed, partitioning of water was mostly between these two water flows when adjusting parameters. Literature which observed similar catchments showed that drainage contributes to approximately 80% of the total runoff. Thus, without available data, this assumption has been used as a guideline to find the best results. For the visualization of the hydrology only the results of the Rk soil have been showed. Other soils respond the same but with different magnitudes. For the calibration and validation three different soils have been used, which represent 31 classified soils in Skuterud. Using all 31 soils is labor intensive and unnecessary, however 3 soils may be on the low side. On the other hand, 1 soil could be enough as well. The reason is that soils are heterogeneous, and the observed values for the pF and conductivity are different every cm. It is impossible to know these values for every square centimeter without disturbing the soil. However techniques exist which can overcome this, which are mentioned in the recommendation chapter.

Assumptions Deep and lateral seepage can be simulated within DRAINMOD, however due to the soil type (low lateral conductivity at deeper depths) of the catchment, this had not been considered. Assumed is that the soil becomes impermeable after 2-meter depth. This leads to another assumption, namely that water does not leave the catchment as groundwater, but only as runoff or evapotranspiration. These two assumptions lead to the following water balance: P = RO - DR - ET. This water balance is also used to calibrate the ET, by taking yearly ET_{act} values instead of monthly values due to snowmelt in March and April, which causes the ET_{act} to be negative. Thus, all precipitating leaves the catchment as either surface runoff, drainage or evapotranspiration. A further assumption relates to the previous paragraph, namely that of the available data. This available data is assumed to be homogeneous over the whole catchment (for weather) or soil type. Thirdly, an area weighted average runoff was made, using each soil type, and multiplied by the total catchment area, which gave the catchment discharge. This could be compared with the observed discharge to calculate the Nash-Sutcliffe values. Residence time and delays are not considered, however the observed data showed that when rain falls in the catchment there is an immediate response in runoff on the same day. With the area weighted average, the assumption that each soil contributes proportionally to the (sub-)surface runoff was taken, depending on their own specific soil physical characteristics. Each soil was simulated separately, of which the results were used for the weighted average. Next, the observed data needs to be put in a format DRAINDMO can read. The precipitation input files show the daily amount of precipitation and the hours it falls. This length and intensity is different per event, and available online, however in this thesis a simpler method has been chosen. The daily values are evenly distributed over a certain number of hours, starting from a certain time, in this case 10 hours from 10 AM onwards. This may result in a rainfall intensity that is smaller than it actually was, causing the model to simulate that all rain infiltrates into the soil, instead of runoff occurring. A final small assumption is that water that enters the drains is directly conducted away and without interference, such as backing up of water in the drains which decreases the drainage intensity.

Method & model For the simulations a warmup period of one year was take, and 7 years for the calibration. The Nash-Sutcliffe efficiency (NSE) was used to assess the model. Results higher than 0.40 were considered acceptable, however values of 0.75 or higher were preferable. Individual years differed sometimes, but overall the results were 0.40 or higher, making them acceptable. Using only the NSE as criteria instead of multiple criteria is tricky, as the NSE is biased towards high runoff whilst low runoff hardly influences it. That is why two visions of the NSE were used, to compensate for that. But the issue remains that more criteria could have been used. The reason why more were not used is that this this thesis focused on the model and how parameters influence the water balance instead of how well the model can simulate 23 years of data. That being said, the model should still be able to correctly simulate the hydrology of a catchment, otherwise the whole point of trying to figure out how parameters influence the water balance means nothing. Nevertheless, using the NSE was the minimum to make sure DRAINMOD works correctly. The calibration and validation results show that over the years the model both under and over predicts the runoff, and that individual events are captured but not always with the correct intensity. As some results showed, the model had a slight bias towards wetter years. The used parameters values were based on literature, with the assumption that those values were plausible. After the model was calibrated and validated successfully, different simulations were performed by adjusting certain parameters. These steps in this order should give reasonable results which can be compared. However assuming that the used parameters from the other study were correct may result in similar results instead of showing where adjustments are needed. To show how the parameters worked, a single extreme runoff event was shown per simulation. However using other events can give additional insights. Using average events and events of low intensity precipitation might show that the model behaves quite differently. This has been done in (Sloan et al.) for example, where the larger of two discharge events showed similar results for different simulations, whereas the second, smaller, event showed distinct differences. During all the simulations it became apparent that adjusting certain parameters, such as surface storage depth and drain spacing resulted in the same outcomes, however through different processes. Both increased the infiltration of precipitation, however the surface storage increases infiltration time whilst drain spacing emptied the soil storage. Besides parameters having the same effect, there is no certain way to know which parameter settings are correct, as there are no separated surface runoff and drainage measurements. As these two parameters are not known, it means that DRAINMOD becomes more of a 'black box' than is desirable, despite it having lots of possibilities to adjust parameters.

Results This paragraphs reflects on the results.

The main result from the surface storage simulations is that there are not many differences, only when going to extremely small values such as 0.015 cm. However this is not realistic. The wet and dry years showed the same result. The surface storage will only change the partitioning between surface runoff and drainage, thus this is a parameter that can be used for final adjustments, when the actual distribution of water is known. Adjusting the surface storage is also a method to include slopes, however even with the lowest storage of 0.015 cm, 1.5 mm, the amount of surface runoff on the 15th of January was not bigger much than that of 0.9 cm.

From the drainage coefficient results it became apparent that there was no significant additional drainage after using 0.5 cm d^-1 as it was in the order a few centimeters at the end of the simulation. Even between 0.05 and 0.5 cm d^-1 the results were not extremely different.

Drain spacing on the other hand has considerable influence on the partitioning of water, with hardly any surface runoff with 50 cm spacings to more surface runoff than drainage with 3000 cm spacings. Adjusting this parameter is the most effective for calibration, however, as mentioned multiple times already, without knowing the distribution of water it is difficult to know the true value. The drain spacing in the fields range from 800 to 1100 cm, and the results show that approximately 80% of the total runoff is drainage. The model is slightly biased towards wetter periods, as differences in 2000 were less than in 1997, however this is mainly due to the warmup and calibration period, and less because of the model structure.

Conductivity values starting from 1 cm d⁻1 to higher values do not show much variation in the total runoff or the partitioning of water. Smaller conductivity values do show more variation, but from 0.01 to 0.1 cm d⁻1 the results are the same again. However, the highest runoff event shows that there is much variation, both when comparing different layers as well as within a layer and different conductivities. The surface runoff was effected most by changes in the fourth layer, with peaks at 3.5 cm. This is because this is the layer where the drains are situated in.

There are not many differences in the temperature simulations, although the rain-snow dividing temperature simulations do show there are differences. Norway has winters with snow and freezing temperatures, so the hydrology is definitely affected by this, however the model does not optimally show this.

Evapotranspiration can be calculated in different ways. For this thesis, the Thornthwaite and FAO Penman-Monteith methods were used as inputs. Data for the calculation was also from the measuring station, a few kilometers from Skuterud. It would be preferable to have field measurements, instead of theoretical values. The average simulated ET was higher than ET$_{act}$, and considering the effects of snow melt causing the winter period to have negative ET$_{act}$, the ET$_{act}$ method should only be used as a reference. A problem with the Penman-Monteith method may be the high level of detailed information needed to calculate the PET, whilst the ET is calculated with more basic equations. For the forest, an adjusted PET file was used, which accounted for the tree height and wind resistance. The FAO$_{crop}$ method for the evapotranspiration was the most in line with the assumption that ET is the rest product of the water balance, however as DRAINMOD has monthly adjustment factors these values are easily manipulated. This is exactly what has been done in this thesis, as the different soil type simulations did not have the same required ET, thus adjustments were made to better fit the model. The whole principle of the FOA$_{crop}$ method using many inputs for more reliable results does not make any sense anymore.

7 | Conclusion

Ockham's Razor mentions: "Everything should be made as simple as possible, but no simpler". This applies to the assumption that P = ET + DR + RO, which although simple, may be good enough to explain the hydrology in detail, at least enough to base decisions on. Of this simple water balance, only the precipitation and total runoff together is known. On the other hand, within the model's interface there are many parameters that can be used to tweak the outcome, some with the same result, which decreases the simplicity. This thesis focused on the effects of adjusting different parameters on the water balance, of which the conclusions are below. These will be presented as answers to the questions asked in the introduction.

How do changes in DRAINMOD parameters influence the hydrological processes and water balance for the prevailing combinations of soil type and land use combinations at field scale? The surface storage, drainage coefficient, drain spacing, lateral saturated hydraulic conductivity and threshold temperatures have been simulated with different values. The runoff does not respond equally to all parameters. The surface storage does not affect the total runoff, however it does affect the distribution between surface runoff and drainage, and can be used to tweak the final partitioning. The drainage coefficient also hardly affects the hydrology, only for very small coefficients. This means that there are other factors more limiting than these two. Drain spacing is one of them. It responds as expected, with larger spacings having more surface runoff. The lateral hydraulic saturated conductivity is the second factor, and of the four layers in the soil the fourth one specifically, as it is there that the drains are situated. Finally there are the temperature threshold simulations, however these do not influence the hydrology much. The rain-snow dividing temperature and snowmelt coefficient show the most variation in the results, as they determine when precipitation is snow or rain and how fast the snow melts. However, the hydrographs show that the peak always arrives on the same day, with different intensities, but also that after the peak drainage continues in a steady regression, hardly affected by the temperature.

How can these combinations at field scale be combined to accurately predict the outflow at catchment scale? Most of the parameters were set the same for all soils, however the temperature and drain spacing were adjusted to better simulate the hydrology. The Rk soil needed additional evapotranspiration, and the easiest and effective way was by changing the monthly value to 2. An evapotranspiration value of 350 mm per year was used as a guideline. The other soil vegetation combinations needed a reduced monthly factor of 0.85, as their initial evapotranspiration was higher than 350 mm per year and needed to be lowered. The most difficult to simulate was the forest. Evapotranspiration was aimed at 500 mm per year, and was done by using an adjusted PET input for the forest. The forest simulation also used drain spacing of 100 m instead of 10 m. These different results were combined by using the weighted average per soil, which showed Nash-Sutcliffe (NSE) values above the 0.40 value classified as acceptable.

Can DRAINMOD simulate the water balance over a longer period of time? The calibration period showed that the model had some difficulty in predicting over longer periods of time, slightly because of a bias towards wetter years. The NSE showed acceptable results, although the model had difficulty grasping the summer period and under estimating runoff. So yes it can, but not extremely well with these parameter values.

Is DRAINMOD suitable for predicting the water balance of small, agricultural dominated catchments in Norway? Deelstra et al. (2010b) simulated the hydrology of Skuterud using DRAINMOD, from 1993-2008, with satisfaction. Further analyses was suggested, for additional improvements of the results, especially in the winter season. This has been done in this thesis. The results were not that different, nevertheless more insight is provided as certain events highlighted the differences that were present, especially in the distribution of surface runoff and drainage. An encountered constraint of DRAINMOD is the slope. Skuterud has slopes, in all directions and gradient's, which need to be captured into one single value. This has been done by adjusting the surface storage, with smaller values representing steeper slopes. However the surface storage eventually has to become so small, in the order of tenths of millimeters, that the accuracy and reliability is questioned. Besides this, not knowing the actual distribution of surface runoff and drainage at the catchment outlet is a major constraint in trying to simulate the hydrology. This thesis also tried to give insight in the evapotranspiration in Norwegian catchments. Calculating ET as remainder in the simple water balance may give as much information as using DRAINMOD on the larger time scale. To conclude and reflect back on Ockham's razor, DRAINMOD may be too simple to simulate the hydrology of small catchments in Norway, especially in hilly catchments.

8 | Recommendations

A good summary for a recommendation is that data precision needs to be increased. As mentioned before, soil physical characteristics are available, but not all. The infiltration parameters are calculated by DRAINMOD now, instead of taken from field measurements. This will result in better infiltration simulations and partitioning of water between surface runoff and drainage. An infiltrometer, reversed auger hole method or the BEERKAN (Lassabatere et al., 2006) method can be possible ways to measure infiltration. Soil moisture storage may also change during the season and years, however it has not been considered, and neither is soil surface storage. More precise soil data can also be found by using a technique called kriging (Goovaerts, 1999), however the details need to be further investigated. Next to this, using hourly climate data can give additional insights in the processes, which are now summarized into one day or peak.

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