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The Effect of Peatland Drainage and Afforestation on Runoff Generation

Consequences on Floods in River Glomma

HYDRA

-ET FORSKNINGSPROGRAM OM FLOM

HYDRA - et forskningsprogram om flom

HYDRA er et forskningsprogram om flom initiert av Norges vassdrags- og energiverk (NVE) i 1995. Programmet har en tidsramme på 3 år, med avslutning medio 1999, og en kostnadsramme på ca. 18 mill. kroner. HYDRA er i hovedsak finansiert av Olje- og energidepartementet.

Arbeidshypotesen til HYDRA er at summen av alle menneskelige påvirkninger i form av arealbruk, reguleringer, forbygningsarbeider m.m. kan ha økt risikoen for flom.

Målgruppen for HYDRA er statlige og kommunale myndigheter, forsikringsbransjen, utdannings- og forskningsinstitusjoner og andre institusjoner. Nedenfor gis en oversikt over fagfelt/tema som blir berørt i HYDRA:

- Naturgrunnlag og arealbruk
- Tettsteder
- Flomdemping, flomvern og flomhandtering
- Skaderisikoanalyse
- Miljøvirkninger av flom og flomforebyggende tiltak
- Databaser og GIS
- Modellutvikling

Sentrale aktører i HYDRA er; Det norske meteorologiske institutt (DNMI), Glommens og Laagens Brukseierforening (GLB), Jordforsk, Norges geologiske undersøkelse (NGU), Norges Landbrukshøgskole (NLH), Norges teknisk-naturvitenskapelige universitet (NTNU), Norges vassdrags- og energiverk (NVE), Norsk institutt for jord- og skogkartlegging (NIJOS), Norsk institutt for vannforskning (NIVA), SINTEF, Stiftelsen for Naturforskning og Kulturminneforskning (NINA/NIKU), Norsk Regnesentral (NR), Direktoratet for naturforvaltning (DN), Østlandsforskning (ØF) og universitetene i Oslo og Bergen.

HYDRA - a research programme on floods

HYDRA is a research programme on floods initiated by the Norwegian Water Resources and Energy Administration (NVE) in 1995. The programme has a time frame of 3 years, terminating in 1999, and with an economic framework of NOK 18 million. HYDRA is largely financed by the Ministry of Petroleum and Energy.

The working hypotesis for HYDRA is that the sum of all human impacts in the form of land use, regulation, flood protection etc., can have increased the risk of floods.

HYDRA is aimed at state and municipal authorities, insurance companies, educational and research institutions, and other organization.

An overview of the scientific content in HYDRA is:

- Natural resources and land use
- Urban areas
- Databases and GIS
- Risk analysis
- Flood reduction, flood protection and flood management
- Environmental consequences of floods and flood prevention measures
- Modelling

Central institutions in the HYDRA programme are; The Norwegian Meteorological Institute (DNMI), The Glommens and Laagens Water Management Association (GLB), Centre of Soil and Environmental Research (Jordforsk), The Norwegian Geological Survey (NGU), The Agriculture University of Norway (NLH), The Norwegian University of Science and Technology (NTNU), The Norwegian Water and Energy Administration (NVE), The Norwegian Institute of Land Inventory (NIJOS), The Norwegian Institute for Water Research (NIVA), The Foundation for Scientific and Industrial Research at the Norwegian Institute of Technology (SINTEF), The Norwegian Institute for Nature and Cultural Heritage Research (NINA/NIKU), Norwegian Computing Center (NR), Directorate for Nature Management (DN), Eastern Norway Research Institute (ØF) and the Universities of Oslo and Bergen.

SAMMENDRAG

Myrgrøfting øker den årlige avrenningen på grunn av at fordampningen blir mindre da overflaten blir tørrere. Hovedeffekten av grøfting på hydrologien er at grøfting endrer vannets veier gjennom naturen; endringer i vannbalansen er små. Grøftingens betydning for flom er komplisert fordi at konsekvensene beror på mange faktorer. Grøfting kan både øke og redusere avrenningstopper.

De hydrologiske endringene etter grøfting kan skje både direkte og indirekte. Den direkte "dreneringseffekten" beror på torvas hydrauliske ledningsevene, myrtype, hydrologisk situasjon og grøfteintensitet. Dersom grøftingen vellykkes vil dette føre til økt skogvekst, økt evapotranspirasjon, lavere snøsmeltingsintentitet, mindre snødybde og redusert avrenning.

Torvas hyrdrauliske egenskaper påvirker avrenningen. Dersom torva har liten ledningsevne, som ofte kan være tilfelle for grasmyr, vil dreneringen føre til relativt høgt vannspeil og liten magasinering som fører til rask overflatevrenning. På mosemyr med liten humifiering, stort fiberinnhold, lav densitet kan konduktiviteten og magasineringskoeffisienten være større som fører til lagring av vann og større flomdemping. Etter dreneringen kan torvas hydrauliske ledningsevne reduseres i tid på grunn av setning og økt nedbryting av torv som fører til raskere avrenning.

Konsekvenser av grøfting beror delvis på myrtype. På grasmyr med mye tilsig fra utmark kan dreneringen ha større konsekvenser dersom grøftene leder flommen raskere en myras naturlige drensystem. Grasmyr kan ha et stort nedslagsfelt og dermed vil drenering av grasmyr i visse fall kunne endre på avrenningen fra et stort areal.

Virkningen av drenering på avrenning er avhengig av den hydrologiske situasjonen. Ved små nedbørmengder på myr som ikke er vannmettet vil en større del lagres og avrenningen forsinkes. Ved større nedbørmengder og vannmettet torv vil dreneringen kunne føre til raskere avrenning. Under snøsmelting kan grøfting i noen tilfeller også føre til raskere avrenning. Forutsetning er at grøftene ikke er tettet av is og snø, at de naturlige helningene er små og utløpet har liten naturlig kapasitet for dreneringen.

Skogen reduserer avrenningstoppene under snøsmeltingen på grunn av stor skyggeeffekt. På vinteren kan det akkumuleres opptil 30 % mindre snø i tett skog på grunn av fordampning av snø. Snøsmeltingen i tett skogbestand er målt til 2 mm pr døgngrad. På snaumark stiger snøsmeltingen fra 3 til 6 mm pr døgngrad utover våren som i hovedsak beror på nedgang i snøens albedo. Differansen, som er 1-4 mm pr døgngrad, kan betraktes som effekten av skogen på redusert avrenning under snøsmeltingen.

Fra 1960 til 1985 var 8-9 % av skogarealet i Østfold og 3-4 % av skogarealet i Hedmark og Oppland grøftet. Fordi at grøftene ikke har blitt vedlikehold har sannsynligvis den hydrologiske situasjonen bilt delvis tilbakeført til det den var før grøfting. På grunn av økt skogproduksjonen kan det være sannsynlig at flomrisikon har blitt redusert. I mindre delnedbørfelt til Glomma der andelen grøftet grasmyr er stor kan grøfting ha bidratt til større flomtopper. Fordi en så liten andel av Glommas nedbørfelt er grøftet, er det lite sannsynlig at skoggrøfting har bidratt lite til økt risiko for *større* flommer.

ABSTRACT

Drainage of peatlands results in a drier soil cover, decreased evaporation and increased annual runoff. However, the main hydrological effect of mire drainage is to change the water pathways not to change the water balance. The effect on peak flow is not so straightforward as it depends on several factors. Drainage can both increase and decrease runoff peaks.

After drainage, changes in runoff can occur both directly and indirectly. The direct "drainage" effect depends on peat hydraulic properties, mire type, hydrological situation and drainage intensity. If the afforestation is successful the increased forest cover will indirectly result in decreased storm runoff due to increased evapotranspiration and decreased snowmelt runoff due to decreased snowmelting rates and less snow accumulation.

The hydraulic properties of peat effect runoff. In cases when the peat has a low hydraulic conductivity, which is often the case for fens, the drainage will result in a relatively high groundwater table, small water storage capacity and rapid runoff. On mires with a high fiber content, low density and degree humification, such as some bogs, the conductivity and storativity will be higher than on some fens and drainage will result in a lower water table, increased water storage and reduced floodpeak. After drainage, the peat hydraulic properties will be reduced due to compaction of the peat by subsidence and increased decomposition which increases runoff.

The consequences of ditching depend on mire type. Ditching of fens may increase runoff from the uplands surrounding the fens whereas ditching of bogs cause changes only in runoff generation from the peatland itself.

The hydrological situation affects the consequences of ditching. During small rainfall events on unsaturated mires, the rainfall is stored and the runoff is delayed. Large rainfall events on saturated peat will result in a more rapid runoff production. Snowmelt runoff peaks are higher on drained areas if the outlet before ditching was not able to carry the meltwater and if the ditches are not blocked with snow and ice.

Increased forest production decreases snowmelt runoff. The snow accumulation can be reduced upto 30 % in forest compared to clearings due to interception evapotranspiration. Snowmelt in dense forest is about 2 mm/degree day. In clearings the melt increase from 3 to 6 during the snowmelt mainly due to increased albedo. The difference in meltrates, 1-4 mm, is due to afforestation.

In the Glomma catchment, forest drainage has probably not increased floods as only a small portion has been drained and the forest production has increased due to drainage. Between 1960 and 1985, 8-9 % of the forest area in Østfold and 3-4 % of the forest area in Hedmark and Oppland was drained. As the ditches have not been maintained, the hydrology in part of this area has probably been restored. In sub-catchements with a high proportion of drained peatlands the flood peaks might have somewhat increased.

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1. BACKGROUND AND AIM OF THE PROJECT

The HYDRA project was initiated after a large flood occurred in the 1995 in the south-eastern part of Norway in the catchments of rivers Glomma and Gudbransdalslågen. The flood was initiated by rapid snowmelt in the high altitude combined with 50-70 mm rainfall in the lowlands. During the flood peak several villages and more than 14 000 ha of agricultural land was flooded which resulted in considerable damage. The exceptional size of the flood volume and peak rose discussion on weather or not the peak could have been amplified by man made changes in land use that has occurred in the catchment.

The main objective of the HYDRA project is to increase knowledge on how watershed management effects floods and develop methods to reduce the flood risk. The effect of peatland drainage has been considered to be one of the reasons behind the exceptional size of the flood peak in 1995. In many Norwegian catchments a considerable amount of peatlands have been drained for agriculture and forestry and, therefore, drainage might have had some effects on hydrology.

The aim of this study is to evaluate the effect of peatland drainage on flood peaks. Previous literature on the effect of drainage is reviewed. Some calculations are presented on the effect of drains and soil moisture storage on runoff generation. Finally a conclusion is made on the effects of drainage based on the information on the area drained and on the results of this review.

2. PEATLAND DRAINAGE IN NORWAY

Drainage has been carried out to improve growing conditions in peatlands. Ditching increases oxygen, temperature and biological activity. In Norway cultivation of peatlands started in the 1700 century. Drainage continued in large scale with government support until the 1970's when subsidience were cut of and the drainage reduced. Altogether about 200 000 ha has been drained for agriculture and 400 000 ha for forestry and approximately 27 000 ha were drained for peat production. In some regions such as in Sørlandet, the southern most part of Norway, peatland occupy today as much as 25 % of the agricultural land.

The open forest ditch network has been quite irregularly dug. The drain distance has usually been 20-50 m. The agricultural drains have been placed more densely at 0.8 m depth and 7-8 m interval. The forestry drains have not been maintained to the same extent as the agricultural drains. The total proportion of drained area in Norway is only 20 % of the peatland area of 3 million ha of peatlands which is considerably less than in the other Nordic countries.

In the flood sensitive catchments of rivers Glomma and Gudbransdalslågen mires have mainly been drained for forestry. Drainage for forestry was particularly intensive in the 50's - 60's when 5000 ha was drained annually. Between 1960 and 1985, 8-9 % of the forest area in Østfold and 3-4 % of the forest area in Hedmark and Oppland was drained. The affected area is larger than the 892064 da ditched because ditching of fens might effect the hydrology of the land area surrounding the fen as discussed later (chapters 3.2 , 4.2 and 6). The statistics of the forest and agricultural drainage in the regions covering the flood sensitive catchments is given in Table 1. The percentage drained compared to the total forested area is largest in Østfold. The drainage has in most districts decreased steadily from the late 1960's. Today mire drainage is generally not allowed and only a small portion is drained for agriculture, peat production, sport facilities and other forms of urban development.

Table 1 Drainage in south-eastern Norway 1920-1997. Total forest area, total mire area drained for forestry (da), and total drained area as a percentage of total forestry area.

Region/ State	Total area of forests (da)	Accumulation of drained area (da)				Drained area as % of forested area			
	1997	1920	1950	1980	1997	1920	1950	1980	1997
Østfold	936000	1234	48902	117828	131123	0,1	5,2	12,6	14,0
Hedmark	11593000	21123	352820	803190	892064	0,2	3,0	6,9	7,7
Oppland	5465000	4778	128874	279118	293322	0,1	2,4	5,1	5,4

3. INTRODUCTION TO HYDROLOGICAL CHARACTERISTICS OF PEAT AND PEATLANDS

3.1 Definitions and characteristics of Norwegian peatlands

Peatland is one type of a larger group of landforms called wetlands. Marshes, coastal floodplaines and swamps have a mineral substrate and do not accumulate peat whereas peatlands form in moist areas where the rate of production of organic material exceeds the rate of degradation resulting in organic peat deposits. Mires are usually used as a synonym for peatlands which are by definition landforms with a peat layer exceeding 30-40 cm. Mires can be classified based on hydrological characteristics into fens which receive water input from rainfall and from the surrounding cathment and into bogs where the rainfall is the only source of water input. A schematic of the hydrological difference and the two characteristic layers, acrotelm and catotelm, are shown in Fig. 1.

In Fenno-Scandia mires formed after the last Ice age 10 000 years ago and reached their present appearance approximately 1000 BC. They were formed directly on mineral soils, from wet forests or from filling of lakes. Due to the variable climatic conditions in different parts of Norway, there exists an exceptional large variety of mires. Along the west coast large rainfall results in very wet conditions which sustain development of peat on relatively steep slopes and formation of blanket bogs which are not present elsewhere in Fenno-Scandia. The most common type of mire are fens (Aapa mires) which spread all over the country. In the North, ice formation and melting of these fens has resulted in large hummocks and in mires called Palsa mires. In parts of Norway with milder climates, the fens have grown and resulted in ombotrophy and formation of raised bogs. Bogs are most abundant in south-eastern part of Norway (Østlandet) and in the Trøndelag region. Direct age measurements have not been made in Norway but growth rate has been estimated to 0,2-0,4 mm/year (Johansen 1996). Based on Finnish estimates the mires have grown to a size of 60-70 % of their maximum size which might be used as an estimate for the size of the Norwegian mires. The average depth of catotelm that has been observed in mire inventories (Johanssen 1997) is 2 m.

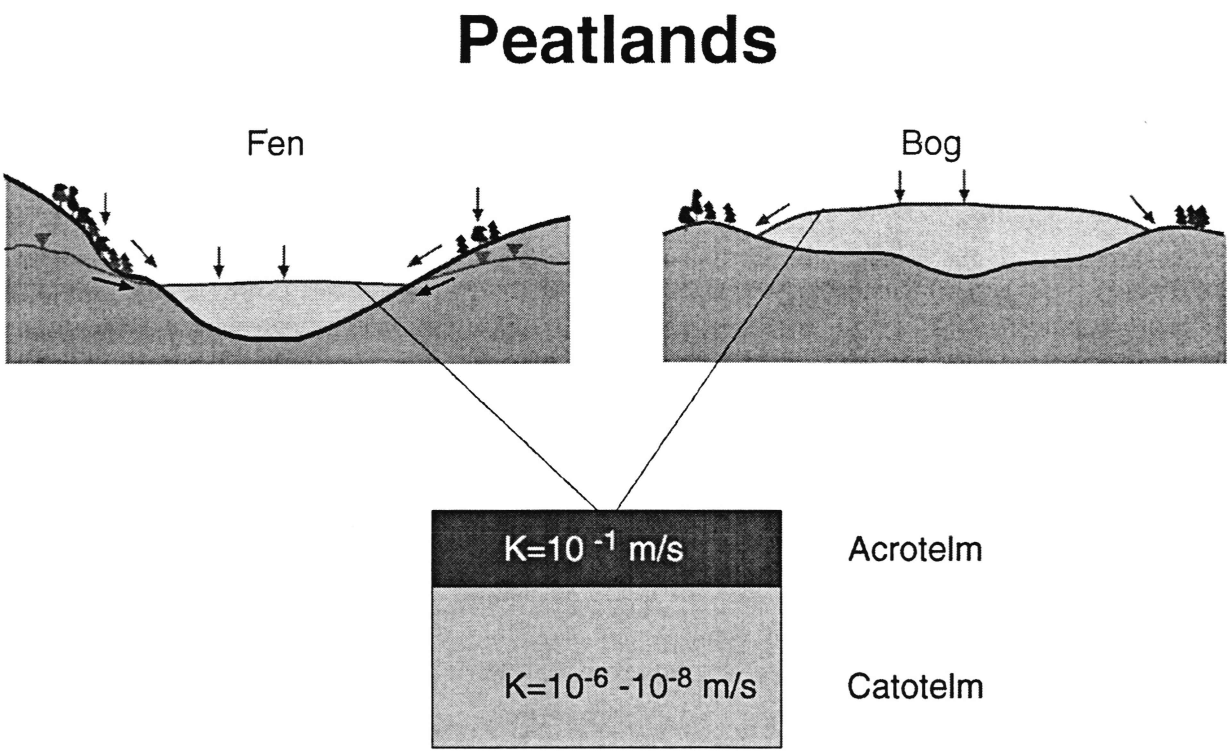

Figure 1 Schematic separation of mires into fens that receive water from the upland and bogs that receive water from rainfall alone.

3.2 Hydraulic properties of peat

Mires consist of two different layer of organic material. The living and rapidly decaying plant layer (acrotelm) overlay a compact brown layer of partly decomposed peat (catotelm). The transition from one layer to the other is rapid. The depth of acrotelm is usually around 0.3-0.5 m. The soil properties governing flow are very different between these two layers. The hydraulic conductivity of the upper layer is about 0.1 m/s (Burt et al. 1990, Hobbs 1986). The lower layer has a considerably lower conductivity ranging from 10^{-3} to 10^{-9} m/s (Burt et al 1990) a typical value being 10^{-7}-10^{-6} m/s (Clymo 1987). The hydraulic conductivity increase with increasing fiber content and decrease with increasing humification and density (Boelter 1969). The unsaturated hydraulic conductivity decreases rapidly with moisture content. This decrease is less for peat with an even pore size distribution and well humified fen peat might have a higher unsaturated conductivity than bog peat have (Eyzerman, 1993). The specific storage (S) of the upper layer is 0.8-1 and for the lower layer 0.13-0.26 (Kløve 1997) but these values are not so well documented. A large variation in the peat properties is due to plant composition, degree of humification, stratification of the peat and compaction. There has not been observed consistent difference between the hydraulic conductivities in the horizontal and vertical directions (Gillman 1994).

Some properties of peat change following drainage (Gregory 1988). Subsidence is a well-documented phenomenon. The primary subsidence of peat is due to loss due to compaction of pores as the water table is lowered. Peat might be further compressed after drainage and wheel traffic from forestry and agricultural machines (Gregory 1988) which might be particularly high in case of low peat shear strength (Kløve 1997). The secondary subsidence is caused by loss of carbon as CH_4 and CO_2 and due to leaching of carbon in runoff waters. An annual settling rate of 1.2-4 cm/year has been observed in Norway (Hovde 1996) on cultivated peat soils.

3.3 Hydrological features of boreal peatlands

Some general conclusions can be made about the hydrology of pristine mires. Mires form in regions where the annual evapotranspiration (ET) is lower than precipitation (P). The soils are relatively wet during periods when P exceeds ET because subsurface lateral and vertical water movement rates are variously limited by combinations of flat or low-lying terrain and low soil conductivity (Gregory 1988). Poorly drained wetland soils are usually saturated near the surface during winter and early spring (Gregory 1988). The low gradient and near-saturated state make it likely that extensive saturation-excess overland flow will be produced (Burt et al. 1990). Due to very little storage for rainwater, streamflow from peat bogs is poorly regulated (Burt et al. 1990). The old popular idea that mires act as sponges mopping up heavy rain is very limited if not entirely mythical (Clymo 1987). The first rain after a long dry period may be effectively absorbed, but once the acrotelm is recharged the ability to retain further water is small.

Storm runoff response is controlled by the layering of the peat into two hydrologically different layers. It has been shown in many studies that runoff from mires depend on groundwater level in the peat (Chapman 1965 ref. Gillman 1994). In response to rainfall, the water table may rise until it intersects the surface and the much higher permeability and allows rapid runoff (Chapman 1965 ref. Gillman 1994). If the groundwater table lies in the transition between acro- and catotelm, as often is the case, rainfall results in rapid flushy runoff response due to the very high conductivity of the uppermost layer. Due to low seepage rates from the catotelm the base flow production from peat-covered catchments is very poor (Burt et al. 1990). It has been observed that during dry periods in summer and winter the runoff might cease completely (Bay 1969, Burt et al. 1990, Sirin et al. 1991, Heikurainen et al. 1978). The outflow from some mires formed in valley depression such as some fens occurs much in the same way as for a lake i.e. controlled by the outlet configuration and the waterlevel in the peatland. Typical for these wetlands would be high runoff peaks during wet conditions and no flow during the summer dry conditions. It follows from the discussion above that virgin mires are characterised by a small portion of base flow and a flushy response to runoff.

Most hydrological studies of mires have been done on raised bogs where it has been showed that runoff depends on the interrelationship of ET and rainfall (Gregory 1988). During the summer there may be periods of several days without rain, during which the water table falls from day to day. The plant roots extract water from both saturated and unsaturated zones and water is re-distributed at night to restore equilibrium above the water table (Gillman 1994). Observation in the UK show that the transpiration compared to potential evapotranspiration is very low in the early part of the summer but picks up in late June and peaks in July (Gillman 1994). Gowin (1977 ref. Gillman 1994) found that transpiration was between 50-60 % of the potential during the summer months and postulated that the ratio would rise to 80 % in September and remain at 100 % over the winter, as interception losses and evapotranspiration took over from transpiration. Total runoff is usually low with relatively low peaks during the growing season when ET maintains high soil moisture storage. During the dormant season when ET is low and soils remain saturated and heavy rainfalls results in high peaks (Gregory 1988).

It should be noted that the hydrology of a mire is site specific as it depends on factors such as for example landsurface inclination and the geological setting were the peatland has been formed. Due to the influence of the upland catchments the hydrological characteristics of fens are site specific and not so well known as characteristics of bogs that receive water input from rain alone. In case of a large surrounding catchment area draining to a fen it is reasonable to assume that base flow is sustained throughout the summer and the runoff is more even than from bogs (Gregory 1988). On the other hand if the surrounding upland portion is small base flow might cease during summer as the water from the upland is evapotranspirated in the wetland. It has suggested that the evaporation from fens is larger than from bogs resulting in a smaller annual runoff from fens (Price and Maloney 1994).

4. REVIEW OF HYDROLOGICAL CONSEQUENCES OF PEATLAND DRAINAGE

4.1 Water balance on drained and undrained mires

Several authors report an increase in low flows following drainage (Seuna 1981, Heikurainen et al. 1978, and Sirin 1991). This is due to the decrease in evapotranspiration rates due to drier surface soil caused by the lowering of the watertable after drainage. In natural condition the evaporation has been found in some studies to be higher on fens than on bogs (Ingram 1983). Results by Eyzerman (1993) indicate that that the soil cover will be drier on fens than on bogs after drainage and therefore the decrease in evaporation might be highest from bogs. For drained bogs the general decrease in ET is almost 100 % in midsummer and is naturally less with developed tree stands (Sirin et al. 1991). Perhaps the most important increase in base flow observed in many studies (Sirin et al. 1991) is due to that the soil water storage capacity is increased and the release of water from this storage maintains a high runoff throughout the summer. The increased discharge of artesian groundwater may in some cases increase summer and winter low-water runoff (Sirin et al. 1991). An estimate of summer low-water increase is given as 50 % by Sirin et al. (1991) depending on the level of alteration in the drainage area.

Most researches point out that drainage does not have a large effect on the annual runoff coefficient. The annual water balance from mires tend to be similar the overall regional waterbalance. Sallantaus (1983) found that the annual runoff from Finnish peat mines is equivalent to the annual runoff in Finland, which means that on average 46 % of the 660 mm of rainfall runs off. The proportion of rain that generates runoff during previously reported annual runoff coefficients from drained areas are: 47 (Burke 1975), 50 (Sallantaus 1994) and 69 (Panu 1988). These coefficients are in the same range as values previously observed from undrained areas: 16 (Hemond 1980), 36 (Verry and Timmons 1982), 50-62 (Burt et al. 1990), 58 (Branfireun et al. 1996), 59 (Arheimer et al. 1996), 60 (Devito et al. 1996), 73 (Panu 1988), 79 (Burke 1975) and 84 Sallantaus (1994). The variation in the runoff coefficients are partly due to difficulties in measuring the water balance form mires (Burt et al. 1990, Price and Maloney 1994). The losses of incoming precipitation through interception is difficult to quantify because of the rather variable vegetation cover and the density of near-ground vegetation

(Gillman 1994). Despite small changes in the water balance drainage changes water pathways (Burt 1995) and effects individual storm peaks (Kløve 1997).

4.2 Consequences of drainage on peak flows

Previous studies on the effect of drainage on peak flow is uncertain (Burt et al. 1990, Robinson et al. 1991, Dunne and Mackay 1996). In a majority of cases a decrease (Hill 1976, Burke 1975, Eggelsmann 1975, Lundin 1988, Baden and Eggelsmann 1964, Burke 1975, Heikurainen et. al. 1978, Robinson et al. 1991) in peak flows have been reported after drainage but an increase in floods has also been noted (Conway & Millar 1960, Krug 1993, Seuna 1981, Panu 1988, Konya et al. 1988). The divergent effects on flows are partly due to that drainage has both reducing and increasing effects on peak flows. Increased soil water storage followed by drainage can temporarily store part of the rainfall whereas the intense channel network and the higher hydraulic gradients result in a quicker runoff (Sirin et al. 19991, Gregory 1988). When additional factors affecting runoff are included such as interception, rainfall intensity and surface morphometry the assessment of drainage consequences is further complicated (see Sirin et al. 1991).

During the non frost season the important factor determining the peak on drained areas is whether overland flow in the acrotelm will occur or not. If the infiltration and storage capacities are not exceeded, the rain only results in a rise in the catotelm groundwater level and the runoff will be small compared to undrained cases where the runoff usually occurs in the acrotelm (Gregory 1988). If the groundwater level rises close to the surface on drained areas the runoff will be increased (David & Ledger 1988). On most forestry drained areas the catotelm storage will quickly be filled and result in rapid overland flow if the rainfall continues (Gregory 1988). Seuna (1981) and Sirin et al. (1991) have observed on forestry ditched areas that the moisture deficit in the peat is rapidly satisfied and runoff peak was not markedly reduced by increased infiltration at least not for heavy rainfalls. A 5 fold increase in peak runoff was observed on a Russian mire (Sirin et al. 1991) after ditching. An increase of 131 % in summer peak runoff and 31 % during snowmelt peak runoff was noted after drainage in Finland with 60 cm drain depth, 40 m spacing and 40 % drained area (Seuna 1981). However, 10-20 years after forest growth the drainage consequences were reduced. The spring maximum peak was 13 % smaller on the ditched area and the summer maximum was only 19 % larger on the ditched than unditched mire. The decreasing peak flow were related to much lower flood peaks in that period, to increased interception in the forestry canopy and to impairment of the ditches. A decrease in spring runoff after drainage when the canopy had developed as been seen by Heikurainen et al. (1978).

The effect of the location of the drained area within the river catchment on the peak flows and timing of the peak has received some attention in the literature. Mustonen and Seuna (1971) and Acreman (1985, ref. Sirin et al. 1991) have observed that peak runoff increase as the most when the drained area is situated in the upper part of the catchment. Sirin et al. (1991) showed with modelling that an even distribution of the drained area results in the lowest peaks and that highest peaks are observed when the drained area is close to the drainage area outlet (as often is the case when fens are drained). Seuna (1981) observed that the peak came 1.5 days earlier on drained areas compared to undrained catchments which was related to drainage and clearcutting.

Very little attention has been made to the effect of drainage on runoff from upland areas surrounding fen mires (Branfireun et al. 1997). This is unfortunate as fens, in particular, are due to the higher nutrient status suitable for drainage and forestry. In the wetland the upland water is partly intercepted and evaporated. After drainage the upland water is conveyed in artificial channels. Therefore drainage evidently increases runoff from the upland portion of the catchment. The effect on the floodpeak is probably small as the runoff from the mires are also quite rapid due to the high conductivity.

The hydrology of boreal mires is dominated by effects of snowpack and frozen soil that results in low runoff in the winter and high runoff during the spring snowmelt period (Gregory 1988). It has been suggested that the impact of forest drainage on snowmelt runoff is more complex than summer time runoff (Sirin et al. 1991) although hardly any published results exist on snowmelt runoff from mires. Results from modelling (Sirin et al. 1991) and observations (Starr and Päivänen 1981) indicate that the effect of drainage on snowmelt runoff is small and only small alterations in snowmelt hydrographs have been observed after drainage. Seuna (1981) noted a smaller increase in spring than summer time flood peaks in a ten-year period after drainage.

4.3 Effect of afforestation on runoff

Much of the research on the hydrological effects of drainage has not included the effects of forestry and tree development (Gregory 1988) despite the fact that usually peatland drainage is carried out to increase forest production. The experimental methods used have not made it possible to separate the hydrological effects imposed by ditching and the effect of tree development (Gregory 1988). It is well known that the vegetation has a strong influence on the water balance (Gregory 1988) and that forests have widely been claimed to reduce flooding downstream (Robinson et al. 1991). The presence of forest cover is associated with reduced annual water yield. This has been demonstrated repeatedly by comparisons of similar forested and non-forested catchment and by noting the effects of deforestation, reforestation, and afforestation (Lee 1980). According to Anderson et al. (1990) previous results show that afforestation of conifers increases water yield by 140-390 mm in climates with less than 1000 mm rainfall.

The developing tree stands decrease runoff by changing snowmelt conditions and increasing interception and evapotranspiration. The development of canopy increases the surface area from where water can more rapidly evaporate. Rainfall quantity, duration and intensity, as well as the state of the crop, all play a part in determining the amount of interception. Robinson et al. (1991) observed a halving of the runoff coefficient when the tree plants had grown from 2 to 22 years. Also Seuna (1981) observed a decreasing trend in runoff as the trees developed. The effect of forest on peak runoff from rainfall will depend on the soil moisture content. On mires with little seepage, increased evapotranspiration will result in increased soil water storage and reduce peak flows. During wet soil conditions, when precipitation exceeds evapotranspiration, the effects of forest in retarding peak flows is minimal as the canopy storage only takes up about 2 mm of precipitation (Lee 1980, Lundberg 1996).

Several studies show that forests reduce runoff peaks from snowmelt (e.g. Schmidt and Troendle 1989, Ohta et al. 1993). Observations show less accumulation of snow in forest than in clearings which has been related, in most cases, to the evaporation of intercepted snow (Lundberg 1996) or to wind redistribution of intercepted snow (Wheeler 1987). A maximum loss of 3.3 mm/24 h was found from two winter measurements in Sweden (Lundberg 1966). Some resent results in Sweden and Norway show that the interception of snow and the consequent loss in water yield can be up to 30 % (Sælthun 1998).

The snowmelting rates used in snowmelt calculations from forested areas is smaller than from open fields. Generally, 2 mm/degreeday has been reported from forests. The melt rate from open fields is larger and more variable than from forests (Lars Bengtsson, per. comm.). In early melt a the rate is about 3 mm/degree and increases towards 5-6 in the late snowmelt mainly due to decreasing albedo. This indicates that in the late snowmelt 40 mm less water will run off from forests during a day of 10 degree Celsius. Indeed timber harvesting in areas with substantial snow cover has been seen to increase snowmelt runoff (e.g. Troendle and King 1985 ref. Lundberg). Because the snowmelt from forested areas is smaller in magnitude and volume than from clear fields, it is logical to assume that afforestation of wetlands will result in smaller peak flows and smaller runoff volumes following snowmelt when the tree stand has developed.

4.4 Effect of increasing vegetation on runoff

The ditch depth on mires is reduced due to erosion, siltation, peat subsidence, freezing-thawing and vegetation Saarinen et al. (1998). The reduction in ditch depth is fastest during the first years after the ditching. The deeper the ditches are dug the quicker is the loss in ditch depth due to peat subsidence. The growth of *Sphagnum* and *Carex* in the ditches might decrease the ditch depth by 25 % in after 5 years (Saarinen et al. 1998). Eventually, without any maintenance of the ditches the, the hydrological situation will return to a natural state (Hytönen & Aarnio, 1998). Observations in northern Finland show a decrease in ditch depth from 70-80 cm to 30-40 cm in 30 years (Lauhanen et al. 1998). Information is not available on how this affects runoff.

5. EFFECT OF DRAINAGE ON RUNOFF: ANALYSIS OF GOVERNING FACTORS

Previous studies on drained and undrained areas show converging results on the effects of ditching on peak flows. This is due to that drainage has both decreasing effect on peak flow due to increased soil water storage and increasing effect due to the large channel network and higher hydraulic gradients (Sirin et al. 19991, Gregory 1988). The increased soil moisture storage capacity allows part of the rain to be temporarily stored in the soil which decreases flood peaks. On the other hand when the moisture storage is filled up, surplus rain results in rapid runoff as the increased channel network allows rapid overland- and groundwater flow enhancing floods. It is therefore important to evaluate the possible moisture storage in the peat caused by drainage.

If considering the hydrological system of drained mires there are three conditions that have to be fulfilled before large peak flows can be produced after rainfalls. These are:

1. Soil and canopy water storage filled up.

2. Rapid runoff from strips to ditches.

3. Efficient channels to convey the increased overland flow.

After the canopy water storage is filled up, excess water is infiltrated into the peat. When the field capacity has been reached, the excess precipitation results in an immediate increase in the groundwater level. The immediate increase in ground water levels can be derived from effective porosity and rainfall depth. An increase in groundwater depth results in an increase in runoff as the hydraulic gradient is increased. It is, for most cases, reasonable to assume that the drainage network is able to rapidly carry the excess water away. It is reasonable to assume that condition 3 does not usually control runoff. On some areas where the ditches have not been maintained, the carrying capacity might have been reduced due to a decrease in channel by peat subsidence depth and increased channel roughness due to vegetation, erosion and siltation. Conditions 1 and 2 (runoff from strips to ditches) are probably the most restrictive factors for rapid runoff generation. Next we will estimate the effect of conditions 1 and 2, calculate the moisture storage available after drainage, and estimate the effect of drains on peak flows.

5.1 The effect of soil moisture storage in attenuating peak flows

Drainage of peatlands decreases the groundwater levels and increases the depth of the unsaturated zone. This could have a large effect on runoff (Lundin 1972 ref. Gregory 1988) as the increased moisture storage allows rainfall to be temporarily stored in the peat. The soil water storage after drainage will be evaluated in this chapter.

The depth of the unsaturated zone after drainage depends on the hydraulic conductivity of the peat, drainage intensity and drain depth. In cases that evapotranspiration from acrotelm is less than the moisture transport from the saturated catotelm the soil moisture stays close to field capacity. The maximum rate of moisture transport depends on the depth of the groundwater table. Eyzerman (1993) has shown with the Darcy moisture transport equation that 50 cm and 70 cm ground water depths for high-moor and low-moor, respectively maintains a 3-4 mm/day moisture transport, i.e., if ET is below 3-4 mm/d the soil stays at field capacity. This is in agreement with field observation by Päivänen (1973) where he observed that when the distance to ground water level remains below 60 cm the soil moisture content follows the theoretical matrix suction corresponding to the distance of the ground water table. The general drainage norm in Norway has been that the depth of the distance to the groundwater surface should be 30 cm (Brække 1983). Usually forest ditches have been 50-80 cm deep at 15 to 30 distance apart. In agriculture a ditch depth of 80 cm at 7-8 m interval has been used in south-eastern Norway (Njøs 1998). Based on the relatively shallow drainage it is reasonable to assume that in case of forest drainage the peat stays at field capacity at least for groundwater fed fens that receive a constant seepage of water from the upland.

Some approximate calculations will be made on the size of the soil moisture storage after drainage when the peat is at field capacity. This is done based on observations by Päivänen (1973) on moisture in forestry drained Finnish peat soils. Using the data by Päivänen, we assume that the capillary fringe is 40 cm and at 81 % volumetric moisture content. The available moisture storage at different groundwater depths is then obtained by taking the average moisture content for each intervals above the groundwater level and summing the moisture content for each layer. The results of this is presented in table 2.

Table 2 indicates that the initial soil moisture storage is usually filled up by rainfalls below 10 mm indicating very little moisture storage on forestry drained mires. In most cases severe floods in the non-frost season occur due to periods of large rainfall. During these circumstance the moisture storage has not a large effect on floods. Also according to Sirin et al. (1991) peat soil storage has little effect during the hydrometrological conditions prevailing during flood periods. This agrees with studies in Finland on forestry drained peatlands where it has been observed that soil moisture storage has an attenuating effect only during small rainfall events and not on large events (Seuna 1981). Open drainage are well known to have little effect in drawing down the water table in the adjoining peat (Boelter 1972, ref. Gillman 1994). However, in some very special situations were the drain intensity is very large and the drains are deep the soils relatively permeable as e.g., on some cutover peatlands, the increased soil moisture storage will attenuate the peak runoff considerably as observed in studies by Kløve (1997).

Table 2 Available soil moisture content based on field measurements by Päivänen (1973) on soil suction and volumetric moisture content.

Distance to GW level cm	Water content vol.%	Soil suction kp/cm2	Available peat moisture storage mm
40	81	0.04	0,00
45	79	0.045	0,50
50	76	0.05	2,33
55	74	0.055	5,25
60	73	0.06	8,80
65	60	0.095	17,92
70	50	0.16	31,71
75	46	0.25	47,69
80	39	0.35	67,11
85	36	0.46	88,20

5.2 The effect of drains in generating peak flows from mires

Lähde (1969), Hemond (1980), Ingram (1987), Devito and Dillon (1993) and Branfireun (1996) have observed that increased groundwater elevation results in increased stream flow from boreal mires. Studies by Ingram (1982) and Burt (1995) show that a Darcian approach can be used to describe the hydrology of natural mires. David and Ledger (1988) show that stream flow peaks from heavily drained peatlands are due to groundwater contribution and on-channel rainfall. During large runoff events, the runoff is almost due to rapid groundwater discharge (Kløve 1997).

The peak flow from a given recharge can be estimated with an analytical model. When Darcian flow is combined with continuity, the flow towards parallel draining ditches is described by

$$S\frac{\partial h}{\partial t} = \frac{\partial}{\partial x}(T + Kh)\frac{\partial h}{\partial x} + r \tag{8}$$

Where r is groundwater recharge, h is the increase in groundwater level above the horizontal initial level as seen in Fig 2. K is the hydraulic conductivity of the soil above the initial water level and T is the transmissivity of the water transporting layer of thickness D below the initial groundwater level.

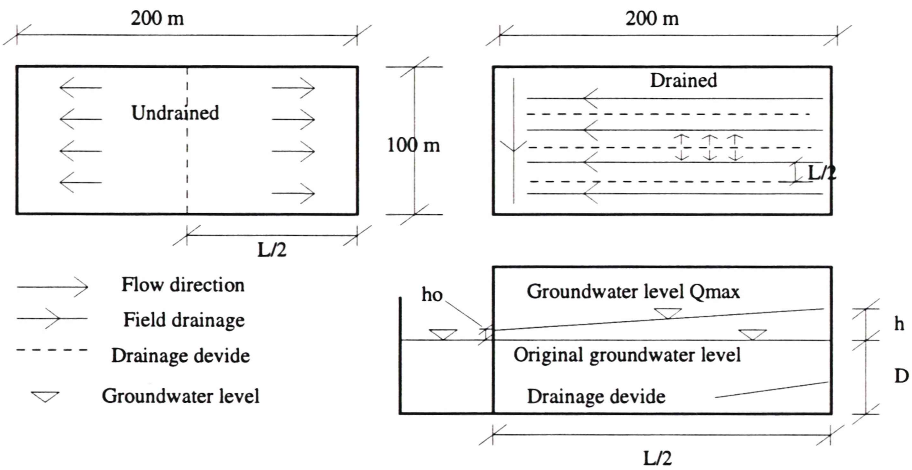

Fig. 2 Boundary conditions for the calculation of maximum expected runoff from drained an undrained areas.

Krajenhoff (1958, see Wesseling 1973) gave an analytical solution to the non-steady flow equation (8) assuming a constant transmissivity in time and horizontal streamlines. A constant value of transmissivity can be used if the changes in h are small compared to D. This assumption is approximately true for the situation depicted in Fig. 2 where the drains reach an impervious layer and the drain depth is large. The flow per unit area is calculated as

$$q = \frac{8}{\pi} r \sum_{n=1,3,5}^{\infty} n^{-2}\left(1 - e^{-n^2 t\alpha}\right) \tag{9a}$$

with the response parameter

$$\alpha = \frac{\pi^2 T}{S L^2} \tag{9b}$$

Where L is the distance between the draining channels and S is specific storage.

The Krajenhoff method can be used to evaluate runoff peaks from different drain intensities assuming that the assumptions mentioned above are applicable. The thickness D depends on the drainage practise and varies usually from 30 to 150 cm. The assumption of small changes in h compared to D holds for the cases when the water level in the peatland ditches are fairly high.

On natural mires the water table is drawn down in the summer after periods of drought and low flow. When the water table rises in acrotelm, increased storm flow is generated due to high conductivity. At high water levels the storage coefficient is rather close to unity, which means that very large storms only cause a small rise in groundwater levels. Due to the high conductivity of the acrotelm, the transmissivity of the catotelm need not be accounted for. Since the water level only rises some few centimetres into the acrotelm and varies along the water flow path, the Krajenhoff solution is very approximate for natural mire conditions.

Forested mires are in their hydraulic behaviour either similar to natural non-forested mires or similar to heavily drained cultivated peatlands, depending of whether the groundwater table lies in the acrotelm or in the catotelm. The watertable is generally in the catotelm during low flows. If the ditches are deep enough, recharge will not fill up the catotelm storage, the groundwater will fluctuate in the catotelm of low hydraulic conductivity and the runoff will be similar to the runoff from cultivated and heavily drained peatlands. If the ditches are shallow, which is usually the case, the recharge will eventually fill the catotelm storage, the groundwater will rise into the acrotelm and increased recharge will be generated in the acrotelm as in natural mires. Because of the dual property of the forestry drained areas, soil properties similar to natural as well as heavily drained conditions needs to be used in the Darcian runoff calculation.

The hydraulic parameter used to evaluate possible runoff peaks for different management scenarios are given in Table 3. Previous studies (e.g. Burt et al 1990, Hobbs 1986) show that the conductivity of the acrotelm exceeds 10^{-1} m s^{-1}. A specific storage of 0.8 has been noted for living Sphagnum by Dooge (1975). On the natural mire and for the forested scenario where the runoff generation occurs in the acrotelm, the transmissivity is equal to the product of hydraulic conductivity and the thickness of the water transporting layer, which is estimated as the groundwater level in the acrotelm when the rain immediately recharges the groundwater.

On heavily drained peatlands the groundwater level remains in the catotelm. A conductivity of 10^{-6} m s^{-1} is a typical to values observed at 1 m depth in the catotelm (Clymo 1987). However, much lower values have also been observed (Hobbs 1986 and Melantie 1988). The specific storage (S) can be estimated to 0.1 (Dooge 1975, Kløve 1997). The depth of the layer D conveying the water usually varies between 0.4-0.8 m at heavily drained areas; slightly less variation is assumed at forestry drained areas where the drains are usually shallower. The drain spacing is typically 7-8 m on agricultural areas and 20-40 m on forestry drained peatland. The total width of a typical natural mire of e.g. 20 ha could be about 300-500 m depending on where the water divide is located. Representative numbers of hydraulic characteristics for different mires are given in Table 3.

The peak flows from the different scenarios calculated from eq. (9) are shown in Table 3. The calculations show that drainage can either reduce or increase peak runoff. The main factor controlling whether the runoff increases or decreases is the location of the groundwater table before recharge occurs. If the mires are drained only to shallow depth, as is usually the case with forested mires, runoff will be generated within the acrotelm and drainage will increase the runoff peaks by almost one order of magnitude from a peak runoff being about 10 % of the daily rain intensity to a runoff peak corresponding almost to the daily mean rain intensity assuming that channel network is able to carry the storm water. The increase in peak flow is due to increased hydraulic gradients imposed by drainage. A high runoff from forestry drained and natural mires are in agreement with data presented by Lähde (1969). His data show that the groundwater table lies close to the soil surface for both drained and undrained mires during autumn rainfall events, which indicates rapid generation of runoff. The results in Table 3 show that runoff peaks from natural mires, about 7-20 % of the daily mean rain intensity, tend to be larger than peaks from deeply plough-drained sites, the runoff peaks from the scenarios being about 7 % of the rainfall intensity. If the channels are deep enough, the groundwater table will fluctuate in the catotelm where the hydraulic conductivity is low, so when the groundwater level controls the runoff, the peak discharge rates will always be smaller on deeply forestry drained areas than on natural mires. It should be noted that on some cultivated peat sites the soil surface is lowered, compacted and the storage reduced so that the groundwater level may reach the soil surface and initiate surface runoff which greatly increases peak flows.

Table 3 Soil characteristics and evaluation of peak runoff from natural and forestry drained areas and deeply drained cutover-peatlands.

Management practice	L (m)	Depth of active flow layer (m)	K (m/s)	S	qmax/P P=20mm/d	P=50 mm/d
Natural state	300	0,025	0,1	0,8	0,132	0,1948
Natural state	400	0,025	0,1	0,8	0,092	0,1458
Natural state	500	0,025	0,1	0,8	0,0735	0,1166
Forested (shallow drainage)	40	0,025	0,1	0,8	0,865	1,008
Forested (shallow drainage)	40	0,02	0,1	1	0,74	0,964
Forested (deep drainage)	40	0,4	1E-07	0,1	0,0297	0,02968
Forested (deep drainage)	40	0,6	1E-07	0,1	0,03205	0,03204
Cutover peatlands*	20	0,4	1E-07	0,1	0,0682	0,068
Cutover peatlands*	20	0,6	1E-07	0,1	0,0729	0,07292
Cutover peatlands*	20	0,8	1E-07	0,1	0,07685	0,0768

*Cutover peatlands for agriculture, peat mining etc with deep drainage.

Since the assumption of constant transmissivity (Kh + T in eq. 8) is not very valid when the groundwater level reaches into the acrotelm, another way of qualitatively relating the peak flow from different areas would be a kinematics approach. Assuming that the groundwater flow is not controlled by the water level in the ditches, it can be further assumed that the gradient, I, dividing the flow towards the ditches is constant. The flux moves with the particle velocity ($v_p = K \cdot I \cdot S^{-1}$) for Darcian flow. Since half the distance between ditches is L/2, the time prior to that rain on an entire basin contributes to runoff (t_c time of concentration) is

$$t_c = \frac{SK}{2KI} \tag{10}$$

The intensity, p, of extreme rainfalls with the same recurrence period is inversely proportional to the rainfall duration, t_d,

$$p \sim t_d^{-b} \tag{11}$$

where the coefficient b from many studies has proven to be about 0.7, e.g. Bengtsson (1991). The maximum runoff peak is produced from a rain of duration equal to the concentration time, provided that the soil is very wet so that all the rain water recharges the groundwater. For those situations, the runoff peak corresponds to the rain intensity, and the peaks from two different areas are related as

$$\frac{\text{peak}1}{\text{peak}2}=\frac{p_1}{p_2}=\left(\frac{t_{c2}}{t_{c1}}\right)^b \tag{12}$$

with index 1 and 2 for different areas. For two areas with the same hydraulic characteristics, S and K, but with different drain spacing as for a natural and shallow drained mire, assuming the hydraulic gradient to be the same, the peak flow ratio is

$$\frac{\text{peak}_{\text{drained}}}{\text{peak}_{\text{natural}}}=\left(\frac{L_{\text{natural}}}{L_{\text{drained}}}\right)^{0.7} \tag{13}$$

where b = 0.7 has been inserted. When the spacing ration is 400/40 =10, the peak ratio is 5. This is lower than the value 10 estimated from the Krajenhoff approach, but clearly shows that shallow drainage causes increased peak flows.

6. EVALUATION OF THE EFFECTS OFF PLOUGH DRAINAGE OF PEATLANDS ON FLOOD PEAKS IN NORWEGIAN CATCHMENTS

Based on the literature review and the theoretical calculations it seems like peatland drainage for forestry can both increase and decrease the flood peaks. The effect obtained depends on the geological structure, geography, climate and ditching practices. The most important conclusions on this study are:

- The ditches increase unsaturated zone and therefore allow more rain to be temporarily stored. However, as this storage is rapidly filled more runoff will be generated due to steeper gradients after drainage. This dual effect results in a reduction of small peak flows and an increase of intermediate peaks.
- The increased tree growth increases evaporation. Evaporation of intercepted snow result in less snow accumulation in forests than clearings. The reduction in snow volume might be up to 30 %.
- The tree stands reduce snowmelting rates from 3-6 mm/degree-day to approximately 2 mm/degree-day resulting in a smaller runoff peak after afforestation.
- Ditching of fens will probably increase the peak flow from the upland. Most mires that have been drained in Norway are fens with a considerable portion of upland area. This implies that a is considerably larger area than just the mires (percentages given in table 1) is affected by drainage. After ditching the runoff will increase if the runoff peak rates from upland was prior to ditching at least partly controlled by poor carrying capacity of overland flow through the natural wetland. The importance in this effect is probably not very large as the hydraulic conductivity in acrotelm is large and the runoff also in a natural state flushy.

• The increase in intermediate peaks might result in changes in channel morphology in such a way that channels become deeper, the flow resistance smaller and the largest peaks become larger (Shawn 1998, per.com.).

In the Glomma catchment, in the south-eastern part of Norway, the proportion of drained mires in large catchments is generally less than 10 %. It is known that the drains function not more than approximately 30 years. It is reasonable to conclude that the drained area is smaller than that given in table 1 and that the area of drained mires will be reduced in the future. If the drained area returns to natural condition after 30 years the drained area is decreasing as most of the drainage was carried out more than 30 years ago. Fig. 3 shows the accumulation of ditched area in different regions when the drain age is assumed to 30 years. It can be seen that the drained portion is usually less than 5 % which indicated that the ditches will not have a large effect on runoff.

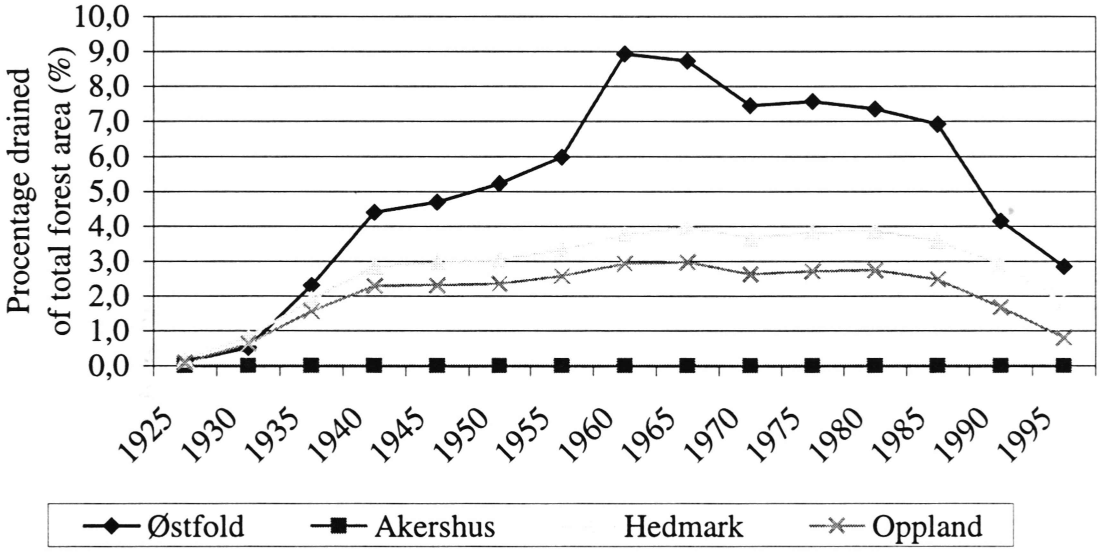

Fig. 3 The accumulated amount of ditched area as a percentage of forest area in different municipalities assuming that the drains are effective for 30 years.

REFERENCES

Anderson, A. R. Pyatt, D. G & Stannard, J. P. 1990. The effects of clearfelling a sotka spruce stand on the water balance of a peaty gley soil at Kershope forest, Cumbria. Forestry, Vol. 63, No.1.51-70.

Arheimer, B. Andersson, L. & Lepistö, A. 1996. Variation of nitrogen in forest streams-influences of flow, seasonality and catchment characteristics. J. of Hydrology, 179, 281-304.

Baden, W. & Eggelsmann, R. 1964. Der wasserkreislauf eines nordwestdeutshen hochmoores. The hydrological cycle of a raised bog in northwest Germany. Schriftenreihe de Kuratorium fur Kulturbauwesen, 12, 1-155.

Bay, R. R. 1969. Runoff from small peatland watershed. J. of Hydrology, 9, 90-102.

Bengtsson, L. 1991, Effective concentration time for design storms in complex urban basins. J. Atmosph. Res. 27, 137-150.

Boelter, D. H. 1969. Physical properties of peats as related to degree of decomposition. Soil Sci. Amer. Proc. 33: 606-609.

Branfireun, B. A. Heyes, A. & Roulet, N. T. 1996. The hydrology and methylmercury dynamics of a Precambrian Shield headwater peatland. Water Resour. Res., 32, 1785-1794.

Burke, W. 1975. Aspects of the hydrology of blanket peat. In: Hydrology of marshridden areas. The Unesco Press, IAHS, 171-182.

Burt, T. B. 1995. The role of wetlands in runoff generation from headwater catchments. In: J. Hughes and L. Heathwaite (eds), Hydrology and hydrochemistry of British wetlands, John Wiley, pp. 21-38.

Burt, T.B., Heathwaite, A. L. and Labadz, J. L., 1990. Runoff production in peat covered catchments. In: M. G. Andersson and T.P. Burt (eds), Process studies in hillslope hydrology, John Wiley, pp. 463-499.

Clymo, R. S. 1987. The ecology of peatlands. Sci. Prog.,Oxf. 71, 593-614.

Conway, V. M. & Millar, A. 1960. The hydrology of some small peat covered catchments in the Northern Pennies. J. Institute of Water Engineers, 14, 415-424.

Coulson, J. C. Butterfield, J. E. L. & Henderson, E. 1990. The effect of open drainage ditches on the plant and invertebrate communities of moorland and on the decomposition of peat. J. of Applied Ecology, 27, 549-561.

David, J. S. & Ledger, D. C. 1988. Runoff generation in a plough-drained peat bog in Southern Scotland. J. of Hydrology. 99, 187-199.

Devito, K. J. and Dillon, P. J. 1993. The influence of hydrologic conditions and peat oxia on the phosphorus and nitrogen dynamics of a conifer swamp. Water Resources Research. 29, 8, 2675-2685.

Dunn, S. M. & Mackay, R. 1996. Modelling the hydrological impacts of open ditch drainage. J. of Hydrology, 179, 37-66.

Dooge, J. 1975. The water balance of bogs and fens (review report). In: Hydrology of Marshridden Areas. The Unesco Press, IAHS, 233-272.

Eggelsmann, R. 1975. The water balance of lowland areas in North-Western regions of the FRG. In: Hydrology of Marshridden Areas. The Unesco Press, IAHS, 355-368.

Eyzerman, N. I. 1993. Physical Properties and Drainage Standards of Forest Soils. Eurasian Soil Science, 25, 104-110.

Gillman, K. 1994. Hydrology and wetland conservation. Institute of hydrology, Wallingford, UK. Wiley, 101 p.

Gregory, J. D. 1988. Hydrologic impacts of forest water management. In: D. D. Hook et al.(eds), The Ecology and Management of Wetlands, timber press, pp. 137-147.

Heikurainen, L. Kenttämies, K. & Laine, J. 1978. The Environmental effects of peat drainage. Suo 29, 49-58.

Hemond, H. F. 1980. Biogeochemistry of Thourau's Bog, Concord Massachusetts, Ecol. Monogr. 50, 507-526.

Hill, A. R. 1976. The Environmental impacts of agricultural land drainage. J. of Env. Management. 4, 251-274.

Hobbs, N. B. 1986. Mire morphology and the properties and behaviour of some British and foreign peats. Quarterly Journal of Engineering Geology, London, 19, 7-80.

Hytönen, L. A. & Aarnio, J. 1998. Profability of ditch-network maiontaenance on some oligotrophic pine mires. Mires and Peat (Suo) 49, 87-99. (In Finnish with English abstract).

Ingram, H. A. P. 1987. Ecohydrology of Scottish peatlands. Transac. Royal Society of Edinburg: Earth Sciences, 78, 287-296.

Ingram, H. A. P. 1983. Hydrology. In: Mires: Swamp, bog, fen and moor, General studies, 67-158,Gore, A. J. P. (ed). Elsevier. 67-158.

Johansen, A. 1996. The extent and use of peatlands in Norway, In:Global peat resources, Eino Lappalainen (ed.). International Peat Society, pp. 113-117.

Johansen, A. 1997. Myrarealer og torvresurser i Norge. 21 s. ie

Kløve, B. 1997. Environmental impact of peat mining, development of storm water treatment methods. PhD thesis, Lund University, 159 p.

Konya, K. D. Robbins, K. D. & Skaggs, R. W. 1988. Evaluating peat-mining hydrology using DRAINMOD. J. of Irrigation and Drainage Engineering, 114, 490- 504.

Kraijenhoff van de Leur, D. A. 1973. Rainfall-runoff relations and computational models. In: Drainage principles and applications, II theories of field drainage and watershed runoff, Intern. Inst. for Land Reclamation and Improvement, Wageningen, The Netherlands, Public. 16. vol II, 246-320.

Krug, A. 1993. Drainage history and land use pattern of a Swedish river system- their importance for understanding nitrogen and phosphorus load. Hydrobiologia 251, 285-296.

Lauhanen, R. Piironen, M-L, Penttilä, T. & Kolehmainen, E. 1988. Evaluation of the need for ditch network maintenance in northern Finland. Mires and Peat (Suo) 49, 101-112. (In Finnish with English abstract and summary).

Lee, R. Forest hydrology. 1980. Columbia Univ. Press.

Lundin, L. 1988. Impacts of drainage for forestry on runoff and water quality. Proceed. Intern. Symp. On the hydrology of wetlands in temperate and cold climates, vol. 1, The Academy of Finland, 197-205.

Lundberg, A. 1996. Interception Evaporation, processes and measurement technigues. PhD Thesis, Luleå Univ. of Tech., Div. of Water Res. Eng. 1996:196 D.

Lähde, E. 1969. Biological activity in some natural and drained peat soils with special reference to oxidation-reduction conditions. Acta For. Fenn. 94, 1-69.

Mustonen, S. & Seuna, P. (1971). Influence of forest draining on the hydrology of peatlands. Publ. of the water Res. Intitiutte, 2. (In Finnish with English summary).

Ohta, T. Hashimoto, T. & Ishibashi, H. 1993. Energy budget comparison of snowmelt rates in a deciduous forest and an open site. Annals of Glaciology, 18, 53-59.

Panu, U. S. 1988. Environmental assessment of peatland development operations in Newfoundland. Proceed. Intern. Symp. On the hydrology of wetlands in temperate and cold climates, vol. 1, The Academy of Finland, 291-300.

Price, J. S. & Maloney, D. A. 1994. Hydrology of a patterned bog-fen complex in south-eastern Labrador, Canada. Nordic Hydrologym 25, 313-330.

Päivänen, J. 1973. Hydraulic conductivity and water retention in peat soils. PhD-thesis, Acta Forestalia Fennica, 129. 70 p.

Päivänen, J. & Paavilainen, E. 1996. Forestry on peatlands. In: Peatlands in Finland, Vesander, H. (ed.). Finnish peatlands Society. 72-83.

Robinson, M. Gannon, B. & Schuch, M. 1991. A comparison of the hydrology of moorland under natural conditions, agricultural use and forestry. Hydrological Sciences, 36, 565-577.

Saarinen, M. Silver, T. & Joensuu, S. 1998. Ditch dimensioning in ditch-network-maintenance areas. A literature review. Suo, 49, 75-85.

Sallantaus, T. 1983. The load to watercourses from peat mining. M. Sc. thesis, Helsinki University, Department of Limnology. 122 p. (In Finnish with English summary).

Sallantaus, T. 1994. Response of leaching from mire ecosystems to changing climate. In: Kanninen, M. & Heikinheimo, P. (eds) The Finnish research program on climate change, second progress report. Publ. of the academy of Finland 1/94. 298-303.

Schmidt, R. A. & Troendle, C. A. 1989. Snowfall into a forest and clearing. J. of Hydrology. 335-348.

Seuna, P. 1981. Long-term influence on forestry drainage on the hydrology on an open bog in Finland. National Boards of Waters, Finland, 43, 3-14.

Sirin, A. Vompersky, S. & Nazarov, N. 1991. Influence of forest drainage on runoff: Main concepts and examples from central part of the USSR European territory. Ambio, 20, 7, 334-339.

Saarinen, M. Silver, T. & Joensuu, S. 1998. Ditch dimensioning in ditc h-network-maintanance areas. A literature review. Mires and Peat (Suo) 49, 75-85. (In Finnish with English abstract).

Starr, M. R. & Päivänen, J. 1981. The influence of peatland forest drainage on runoff peak flows. Suo 32, 79-84.

Verry, E. S. & Timmons, D. R. 1982. Waterborne nutrient flow through an upland-peatland watershed in Minnesota. Ecology, 63, 5, 1456-1467.

Wesseling, J. 1973. Subsurface flow into drains. In: Drainage principles and applications, II theories of field drainage and watershed runoff, Intern. Inst. for Land Reclamation and Improvement, Wageningen, The Netherlands, Public. 16. vol II, 1-56.

Wheeler, K. 1987. Interception and redistribution of snow in a subalpine forest on a storm-by-storm basis. proc. 55th Western Snow Conf., Vancpouver, B. C. , pp. 78-87.

UTGITTE NOTAT I HYDRA-SERIEN

1/97 **Metoder for kvantifisering av hydrologiske prognosefelts representativitet.**
Kai Fjelstad, NLH (diplomarbeid).

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