

Technical Report

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**Wetlands – different types, their
properties and functions**

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This report concerns a study which was conducted for SKB. The conclusions and viewpoints presented in the report are those of the author and do not necessarily coincide with those of the client.

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Summary

In this report, different Swedish wetland types are presented with emphasis on their occurrence, vegetation cover, soil physical and chemical properties and functions.

Three different main groups of wetlands are identified: bogs, fens and marshes. The former two are peat forming environments while the term 'marshes' covers all non-peat forming wetlands. Poor fens are the most common type in Sweden but (tree-covered) marshes would probably be dominating large areas in Southern Sweden if not affected by human activity such as drainage for farming. Fens and bogs are often coexisting next to each other and bogs are often seen to be the next step after fens in the natural succession. However, the development of wetlands and processes of succession between different wetland types are resulting from complicated interactions between climate, vegetation, geology and topography. For description of the development at individual sites, the hydrological settings which determine the water flow paths seem to be most crucial, emphasizing the importance of geology and topography.

For modelling the growth of peat, simple models have so far dominated, but these are often restricted in general use. Therefore, more physical-based models have been developed, but the natural heterogeneity and climate shifts bring uncertainty of how they should be parameterised. The use of coupled groundwater-substance transport models have shown to help understand how water flows and soil chemistry are developing in response to different peat forms.

The peat is characterised by a high porosity (80–95%) which is decreasing with decomposition and depth. The most important change is then that the actively conducting pores are getting clogged and closed. The storage coefficient of the peat is then decreasing with decomposition as well as the hydraulic conductivity. The variation of hydraulic conductivity (k) can be large and current understanding of the connections between hydraulic conductivity and other properties is insufficient. Relationships can often be shown but it is found to be hard to evolve predictive functions for general use. However the hydraulic conductivity has shown to be a very important parameter for transport predictions since a layer with different k can link off the flow considerably.

Peat materials can very easily remove cations from solution and retain them. The humic substances are then considered to determine the binding capacity. The dominating exchange sites of peat compounds are the carboxyl ($-\text{COOH}$) and phenolic hydroxyl ($-\text{OH}$) groups. The ions bind to these by ionic attraction as well as by forming complexes of different strength. The active groups of humic substances shift character with pH, and the fixation and release of substances are depending heavily on pH and redox conditions since these also determine the properties of the transported substances. The transport of substances can be considered to depend on both dispersion-diffusion and advection, depending on the water flow direction. The predominant downward flow direction in bog mounds keep the minerals from getting transported into upper layers by diffusion/dispersion while these processes can cause some substances to get considerably enriched in fen peat where the water flow is horizontal. The modern coupled groundwater-transport models have shown to be useful tools to understand and describe the processes and to estimate parameters for prediction purposes. By successful use of these models, simpler general models may be evolved and used for long term predictions.

Sammanfattning

Våtmarker har vid de senaste säkerhetsanalyser av ett djupförvar av använt kärnbränsle visat potentialen att ge högre doser än många andra recipienter. För att förstå de viktigaste processerna i våtmarker och hur framtida dosmodeller ska utvecklas har denna litteratur-sammanställning gjorts.

I denna rapport presenteras fakta om olika svenska våtmarkstyper avseende deras förekomst, vegetation samt fysikaliska och kemiska egenskaper och funktioner i deras mark.

Tre huvudtyper av våtmarker identifierades: torvbildande våtmarker med de olika typerna mossar och kärr samt övriga icke torvbildande våtmarker. Fattigkärr är den vanligast förekommande typen av våtmark i Sverige men rikare slättlands-sumpskogar skulle förmodligen vara mycket vanliga i södra Sverige om de ej dikats ut och odlats upp. Kärr och mossar förekommer ofta sida vid sida i så kallade myrkomplex och kärr ses ofta vara naturliga föregångare till mossar genom torvtillväxt. Dock har man funnit att utvecklingen från en våtmarkstyp till en annan beror på ett komplicerat samspel mellan klimat, vegetation, geologi och topografi, varför någon generell succession med mossar som slutled inte kan sägas gälla. För beskrivning av våtmarksutvecklingen vid enskilda platser, anses det mest angeläget att kunna återge de lokala hydrologiska förhållandena.

Enkla modeller har hittills dominerat när det gäller att försöka återge torvtillväxt. Dock är ofta nyttan av dessa begränsad för generellt bruk, varför mer fysikaliskt baserade modeller har utvecklats. Den stora rumsliga variationen av egenskaper i naturliga miljöer gör dock parameterisering och användning av dessa modeller mer komplicerad. Användning av kopplade grundvattenflödes-transportmodeller har visat sig vara till hjälp för förståelse av hur vattenflöden och markkemi har utvecklats vid olika torvmarksformer.

Torv karaktäriseras av en hög porositet (80–95%), vilken minskar med ökad nedbrytning och ökat djup. Den största förändringen i struktur, som nedbrytningen medför, är att stora, kontinuerliga porer försvinner och strukturen övergår till att likna en massa. En ökad nedbrytning medför därmed att magasinskoefficienten och hydrauliska konduktiviteten minskar. Variationen av hydraulisk konduktivitet (k) i torv kan vara stor och nuvarande kunskaper om kopplingar mellan konduktivitetens storlek och andra egenskaper är otillräcklig. Samband kan ofta ses mellan olika egenskaper men det har visat sig svårt att utveckla generella, prediktiva funktioner. Hydraulisk konduktivitet är dock en mycket viktig parameter för beräkningar av ämnestransporter eftersom lager med annorlunda k kan avlänka flöden betydligt.

Torv kan fastlägga kat-joner mycket effektivt. Den stora bindningskapaciteten beror till stor del på mängden humusämnen. Humusens aktiva grupper för bindning av ämnen är främst dess karboxyl- ($-\text{COOH}$) och hydroxylgrupper ($-\text{OH}$). Dessa grupper binder joner såväl genom elektrostatiske attraktion som genom bildande av komplex av olika styrka. Humusämnenas aktiva grupper utgör syror och skiftar därmed karaktär med varierande pH. Fastläggning och lösning av ämnen beror därför till stor del på pH och även redoxförhållanden eftersom dessa också bestämmer egenskaperna hos de transporterade substanserna. Transporten av ämnen sker genom både advektion och diffusion/dispersion, beroende av vattnets flödesvägar. Det dominerande nedåtriktade flödet i mossar förhindrar ämnen att transporteras upp till de övre torvlagren genom diffusion/dispersion under det att dessa processer kan ge upphov till en betydande anrikning av ämnen i vissa kärr, där flödet är mer horisontellt. Moderna, kopplade

grundvattenflödes-transportmodeller har visat sig vara användbara för att förstå och beskriva våtmarkers funktioner och kan vara användbara för att uppskatta parametrar för prediktioner. Framgångsrika anpassningar och bruk av dessa modeller kan vara till hjälp för framtagandet av parameterisering för utveckling av enklare modeller för mer generell användning samt också för långtidsprediktioner.

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1 Introduction

In SKB:s previous safety assessment for a repository with spent radioactive fuel (SR97) wetlands were found to be able to accumulate radionuclides in higher amounts than other recipients /Bergström et al, 1999; SKB, 1999/. This report gives some more understanding of important processes in wetlands for future dose models with wetlands.

Wetlands are defined as land that have the water table at, near, or above the land surface or which are saturated for a long enough period to promote wetland or aquatic processes as indicated by hydric soils, hydrophytic vegetation, and various kinds of biological activity that are adapted to the wet environment. The term wetlands include several different environments with various properties and the only common feature is that the water table is situated, occasionally or constantly, close to the surface and hence there is a close connection between the groundwater and the biosphere. This close connection is significant for the impact of contaminants on the environment. The objective of this report is to present knowledge on wetlands, the basis for classification, their appearance and how they develop subjected to hydrology and climate, their hydrological functions and physical properties, their chemical properties and functions, the processes of substance transports and sorption and how wetlands are influenced by human activity. This knowledge of wetlands is crucial for estimation of long-term impact of contaminants on the environment concerning their transport with groundwater and possible enrichment and discharge within wetland areas. For estimating the long-term effect of possible transport of radioactive nuclides from deeply located nuclear-waste disposals up to the biosphere, the significance of wetlands' functions should be very high.

2 The different types of wetlands

2.1 Different wetlands, classification

Classification of wetlands can be made on several basis, depending on purpose. Most common are classifications based on ecological concepts, using the plant cover as indicator of hydrological and geochemical status. The occurrence of a peat layer thicker than a few decimetres is also used as a divider between peat forming types and other land forms. From some aspects (ecological and geochemical) it is also convenient to differ between salt-water and fresh-water wetlands. The Swedish Environmental Protection Agency used the following division in an inventory of Swedish wetlands /Löfroth, 1991/:

Series 1: Mires, i.e. peatforming wetlands not influenced by surface water from sea, lakes or streams:

- 1 a) bogs (ombrotrophic mires), forested or open,
- 1 b) fens (minerotrophic mires) open, shrub or treed,
- 1 c) mixed mires, mixed by fens and bogs.

Series 2: Shorelines, i.e. wetlands largely influenced by surface water from sea, lakes or streams. They can be open or forested. They are divided into

- 2 a) freshwater (limnogenous) wetlands,
- 2 b) salt water (marine) wetlands.

In this report, some properties can be identified as significant for the fate of inflowing minerals. The presence of peat is recognised as significant, as well as the hydrological and geochemical settings and the vegetation properties. In practise the division between the different wetland types is delicate and in this report only the terms bog, fen and marsh will be used. The term marsh will then cover wetlands without substantial peat layers, while mires, that will say bogs and fens are characterised by peat forming vegetation. The definitions of these types are here:

Bog: peat covered area, ombrotrophic (vegetation only supplied with water from precipitation).

Fen: peat covered area, minerotrophic (vegetation supplied by at least some water from inflowing surface water or groundwater).

Marsh: Little or non-peat forming soil, minerotrophic, also open water (aquatic) wetlands are within this group.

These definitions agree largely with common international classification practise concerning the peat forming types, although the single mineral-soil type used here is often divided into different classes /National Wetlands Working Group, 1988; Löfroth, 1991; Mitsch and Gosselink, 2000; EPA, 2002/. Forested wetlands (also called swamps) are then often treated as a separate group within wetland science and management. They are here also presented separately but then considered to be within the three groups mentioned above, since they have the same soil conditions as open (treeless) wetlands but carry trees or woody shrub as dominant vegetation.

2.2 Main features of the different types

2.2.1 Bog

The bogs are characterised by peat soil, nutrient poor and acidic soil conditions, a low species diversity and low primary production. The central parts of the bog are supplied by rain water only (Figure 2-1). The surface layer is dominated by peat moss (*Sphagnum spp.*), which demands low pH and/or low concentrations of Ca in the substrate. The *Sphagnum* mosses also cause acidification by releasing hydrogen ions from organic acids, associated to the cell walls, in exchange for other cations. The present vascular plants on bogs endure the nutrient-poor conditions thanks to a slow growth rate, an ability to conserve nutrient resources, a long lifespan and a successful vegetative propagation /Backéus, 1985, cited By Rydin et al, 1999/. They need further to have an aerenchyma to supply the roots with oxygen for respiration or to grow on (at least partially) aerated parts, such as hummocks or fairly drained (sloping) border sites. The surface has often a small-scale varying topography, with hummocks, hollows and pools. Hollows can be further subdivided into lawns, wet carpets and mud bottoms. Hummocks are raised 20–50 cm above prevailing water-table level, and are therefore well aerated in the shallow layers. Their characteristic vascular plant species are *Calluna vulgaris*, *Empetrum nigrum*, *Rubus chamaemorum*, *Betula nana*, *Vaccinium oxycoccus* and *Andromeda polifolia*. On dryer sites may also low pine (*Pinus sylvestris*) trees grow. The hollows are soaked except during longer dry periods and host water tolerant species. Characteristic for bog hollows are *Eriophorum vaginatum* or *Trichophorum cespitosum* and in the wetter parts (carpets) *Carex limosa* and *Scheuchzeria palustris* are added. Mud bottoms have only few individuals of the carpet species but can host some microalgae. The processes and factors behind the development of these micro-relief elements are uncertain but the structure seem to be stable over time /Rydin et al, 1999/. Commonly are these micro-relief elements symmetrically organised in elongated forms of ridges and hollows perpendicular to the water-flow direction. Hence the surface patterns are likely initiated and developed through the combined effect of the hydrological and biological factors /Seppä, 1996/. There are typically three different main parts of a raised bog. The mire margin is often a fen, called lagg fen, as it is situated at the border and receives water from the surrounding mineral soil as well as it is draining the bog. From the margin, the bog is raising up to the bog expanse. The slope inwards from the margin is often better drained and hosts a low pine forest. At the centre mire expanse, water table is again continuously close to the surface and there are normally few trees. There are different forms of bogs, most likely depending on differences in climate, but topography may also influence the formation /Seppä, 1996/. Most common are small bogs with weakly developed domes, often tree-covered and constitute often components of larger mire complexes. Concentric raised bogs, with a dome-shaped cupola, occur mainly in the southern part of Sweden, with increasing height of the cupola with increasing excess of precipitation. In dryer climates the cupola is often weakly developed and tree-covered. With greater humidity, the occurrence of pools increases, while the shape of the bog turns from the domed into more irregular formes. Soaks (dråg), forming elongated channels carrying some fen species, can occur. Their possible function as drainage channels is yet unclear, although their fen species can be caused by a greater supply of minerals by the flowing water /Lonnstad and Löfroth, 1994/. An alternative explanation to the fen vegetation can be a formation of local flow cells /Siegel and Glaser, 1987/ which is described under point 5.5. In areas with small topographic variation, ‘plateau bogs’ with a steep margin slope and an almost horizontal central plateau are common. Where the topography is more pronounced, eccentric raised bogs evolve, with a unidirectional or a fan-shaped slope. Water from surrounding areas can enter and cross the mire area as sloping (elongated) flushes (soaks, dråg) with fen-type vegetation. In northern Sweden, bogs with thick peat layers are uncommon. Instead another, flat, type with thin peat is common in the north. These bogs are situated at level or slightly sloping parts of recharge areas /Rydin et al, 1999/.

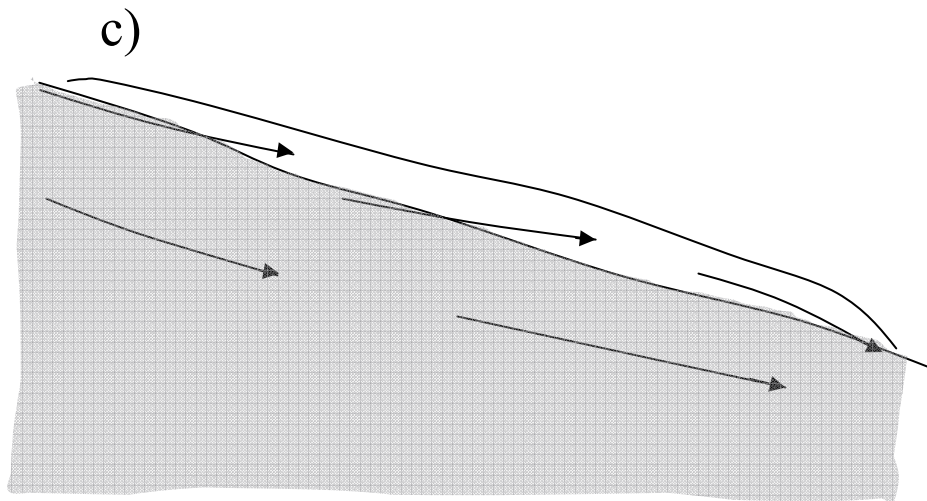
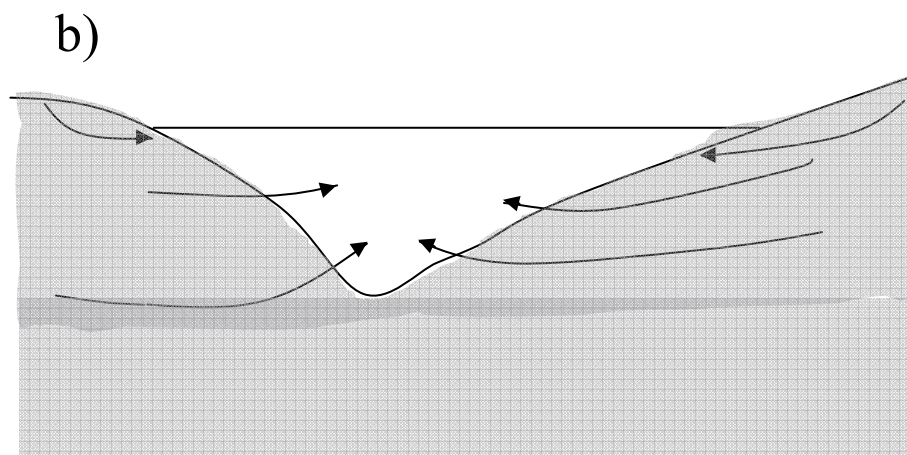
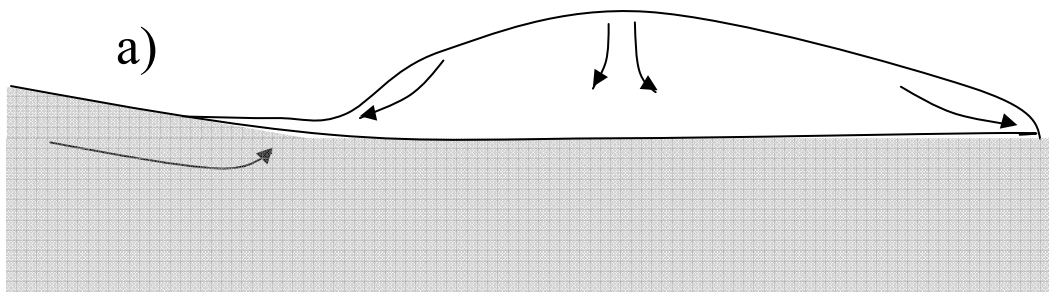


Figure 2-1. Some different mire formations with prevailing water flow directions through the peat layers and underlying mineral soil (shaded area). a) Bog (ombrogenous mire), b) valley fen (topogeneous), c) sloping fen (soligenous).

2.2.2 Fen

There are several similarities between fens and bogs; there is a peat soil and the water table is continuously close to the surface, the surface has often a micro-relief of hummocks, hollows and pools. However, major differences can be seen in the composition of vegetation and in the peat, as a result from the influence of water with a terrestrial origin. As the water supply becomes influenced by laterally inflowing water, the vegetation and type of fen will be governed by interaction between the substrate and the chemistry of the water, especially the pH and the Ca content. Hence is the character of the fen depending on the geologic settings and the proportions of lateral inflow from surroundings compared to precipitation. The basic division is between poor fens, intermediate fen, moderately rich and extremely rich fens. The terms rich and poor relate to the richness of mineral elements and not to the all-round nutrient richness. Although the pH is very well correlated to the type of fen, some variation may occur and in fact is the type of fen given by its composition of plants. For example the fen limit indicators, which grow on poor fens but cannot breed on bog substrate, thus help to point out the borders to the bog. Useful species herein are *Carex lasiocarpa*, *C. rostrata*, *Eriophorum angustifolium*, *Menyanthes trifoliata* and *Sphagnum fallax*. The character of the surface of the poorest fens is then similar to the bogs, with a dominance of Sphagnum mosses and just a few more vascular species, mostly grasses and sedges (*Carex*). The poor fens are the most common in Sweden since the geology is dominated by the siliceous bedrock and its mechanically weathered products as the overburden of glacial till and glaciofluvial deposits. These can cover vast areas in humid regions, together with intermediate fens, common in boreal Sweden and found to have a similar electrolyte concentration to poor fens but slightly higher pH. Moderately rich fens occur where there is a great supply of non-acid terrestrial water or in areas with some calcium in the soil. Extremely rich fens are found in calcareous areas, i.e. where there is limestone or calcareous drift. The vegetation of rich fens can differ greatly from poor fens. The peat mosses dominance in the bottom layer in poor mires is taken over by brown mosses and there is an increase of vascular species, with e.g. more *Carex* species. A characteristic feature for the extremely rich fens is the occurrence of many orchid species. The species characteristic for each type is given by /Rydin et al, 1999/.

Fens are typically located in discharge areas, in valley bottoms (topogenous) but occur also along valley slopes, especially in wet areas (soligenous) (Figure 2-1). As with the bog hollow and hummock topography, fens often develop similar patterns with hummocks, lawns and carpets, and mud bottoms. In sloping fens, especially in boreal Sweden, they turn into patterned fens with elongated hummocks, called ridges or strings between which the surface is flatter and form poorly vegetated 'flarks', which are mostly waterlogged. Where the slope is negligible, the string pattern becomes a wide network. In cases where the strings develop to be high enough to be less affected by the through-flowing water, their vegetation become bog-like. The mire can then be referred to as a mixed mire, a type which is common in the boreal zone. Another type of mixed mire, very common in the boreal zone, has an irregular combination of bog and fen elements and can be referred to as a mosaic mixed mire /Rydin et al, 1999/. Typical mires, covering vast areas in Northern Sweden are actually complexes, composed by *Carex* lawn and carpet fens, flat bogs, string mixed mires and wet flark fens, open and forested parts. In the south-eastern lowlands, flat fens and other non-peat wetlands are more common. Quagmires (floating mires) are often developed along oligotrophic lakes by poor fen communities, and are kept floating by roots and rhizomes of *Carex* spp. *Eriophorum* spp. *Menyanthes trifoliata* and others carrying a *Sphagnum* spp. carpet.

The stability of water outflow depend much on the scale of the catchment for the flow into the fen. An inflow with regional source is more stable in time than a local source for both groundwater and surface water inflow.

2.2.3 Mineral soil wetlands, Marshes

Similar to mires, marshes or mineral-soil wetlands often has a substantial content of organic matter within their surface layers. However, the organic matter is not accumulating at a rate high enough to cause peat formation. The variation of water table level is larger than in the peatlands. The redox conditions are hence more oxidising in surface layers during the dryer periods and a more complete decomposition of organic matter is allowed. The variable water table level gives also a large storage volume and the dampening potential of water outflow can be high. Similar to the fens, the pH as well as nutrient status depends on the supplying water. Primary production can be high as well as the diversity of species. Since marshes cover a great number of different environments, they can exhibit various appearances. Common is a mixture of different types of marshes lying adjacent to each other, depending on the surface level in relation to water table. There are often successive changes in appearance along so called *ecotones*, e.g. from open water (or fen) into dry land forest.

For lakeside or seaside marshes along shorelines, the water table is often temporarily or seasonally varying. There is an influence from surface water during flooding periods, which can bring both dissolved and particular substances to the soil, but also can flush the sediments and upper layers as well as bring about erosion and export. The lower, aquatic parts are more part of the limnic or marine system, and if connected to the open lake, there is an exchange with the lake water. These are generally inhabited by *Phragmites* and *Scirpus* societies. A more periodically inundated area is situated in the shoreline at the level around the mean water level. These areas can host a certain, minor peat formation but the peat is often mixed with sediments. The dominance of *Carex* spp. and other submergence-tolerant grasses indicate more or less continuous anaerobic conditions. In the next level in the ecotone from lake to dry land conditions the moist shore meadow (*strandäng*) occurs. It is occasionally submerged but is sufficiently high above mean water table to sustain facultative or terrestrial species. Most of these shoreline areas have been used for grazing and harvesting and are thereby kept open. If this usage ceases, there will soon be an overgrowth of *Phragmites australis*, *Glyceria maxima* etc in wetter parts and of *Salix* spp. *Betula pubescens* and *Alnus glutinosa* in the slightly drier parts, allowing only a thin borderline for the *Carex* dominated shoreline vegetation /Jacobson, 1997/. The marine shorelines resemble the fresh-water shoreline wetlands but host salt-tolerant species in an extension depending on the salt concentration. These have a varying water table level as well and are inundated either by tide floods or irregularly by onshore wind. They are often initiated at sheltered bays in deposited sediments, inhabited and stabilised by fast establishing species such as *Salicornia europaea*, which also effectively catch sediments and detritus, enabling formation of a cover of humic matter. These tidal salt marshes cover extensive areas in southern North sea, but well-developed seaside wetlands do not cover large areas along the Swedish shoreline, except in the south-west. This is probably due to the land heave together with human activity (grazing cattle and harvesting) /Sjörs, 1956/.

The riparian wetlands (wetlands that are influenced by adjacent streams or lakes) extend over a wide range of environments, depending on the local groundwater-table regime, landform and parent material, the variability of timing, magnitude and duration of flooding (see e.g. Figure 2-3), and the water-quality yield from flooding water. It could be argued if stream-influenced riparian wetlands should be treated separately, since the processes are similar to other wetlands close to surface water bodies. It is also hard to denote the borderline between large-scale wetlands emanating along large floodplains and shoreline marshes. However, treating stream-influenced wetlands separately may be useful since the occurrence of streaming water can generate a greater transport of substances and flushing of surface layers than in calm water. Consequently the stream-influenced wetlands are often highly productive, although a negative effect on the productivity could come from the stream water if there is a large proportion of dystrophic (or ombrogenous) peat areas in the upstream zone /Fölster, 2001/.

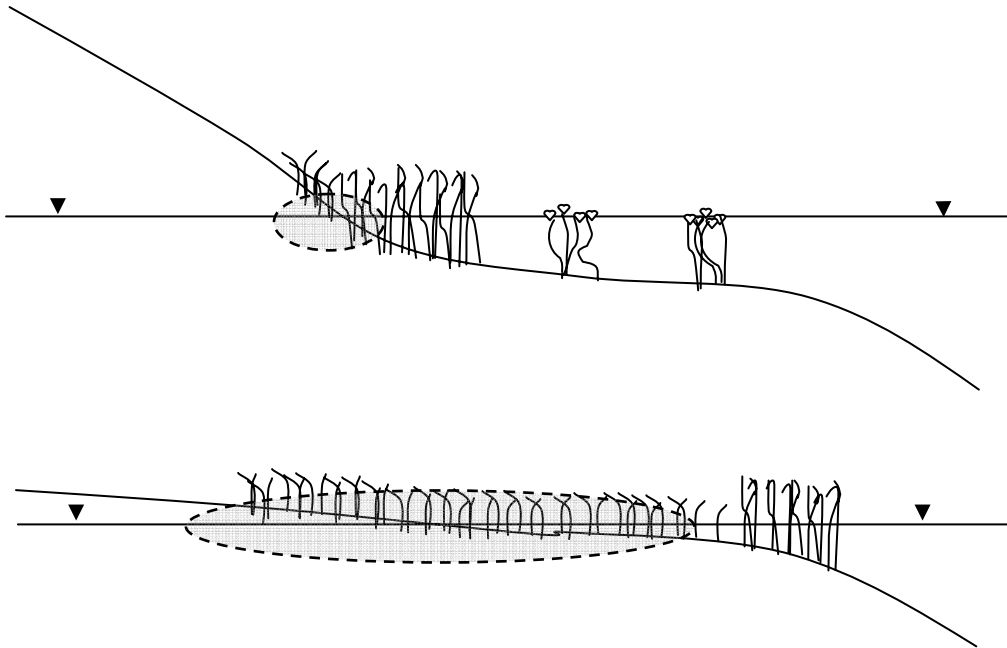


Figure 2-2. Shoreline wetlands with different topographic settings, leading to different wetland environments. The shaded areas denote potential area for peat formation.

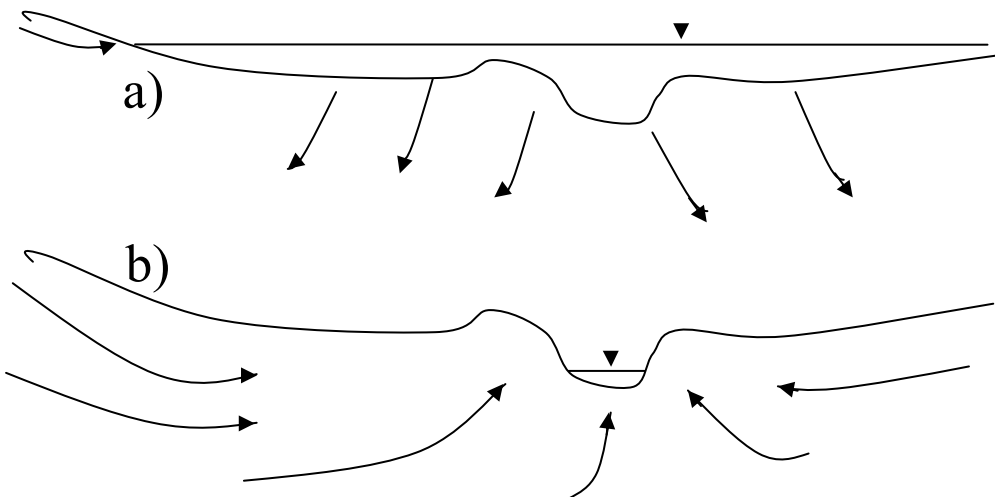


Figure 2-3. Principle view of the flow lines of water at a riparian wetland during a) stream high flow and b) stream low flow.

2.2.4 Forested wetlands, Swamps

Forested wetlands (swamps, according to American vocabulary /National Wetlands Working Group, 1988; EPA, 2002/) span over almost all types of wetlands, except for very wet or poor conditions, which preclude establishment and growth of trees. A prerequisite for tree growth is that the ground is sufficiently aerated during the main parts of the year, even if the growth can be significantly reduced. The forested wetlands are often characterised by a mixture of wetland and forest ecosystems where both species from wetlands and forests appear resulting in a very rich diversity for both vegetation and fauna. They are often part of larger systems within wetlands or forests. A common threshold for separating forested and open wetlands is a 30% cover by the treetops /Rudqvist and Fiskesjö, 2000/. Similarly to the

open wetlands, there is a wide range of vegetation species with different nutrient demands. In bogs, the outer parts are often covered with very slowly growing Pine, *Pinus sylvestris*, (and occasional birch, *Betula pubescens*) plants. As the access of mineral substances increases, the proportion of pine decreases while the proportion of spruces (*Picea abies*) and birch increase. In the richer forested wetlands, the deciduous trees dominate, with alder (*Alnus glutinosa*) and in southern Sweden also ash (*Fraxinus excelsior*). The growth of the trees, which largely sets the productivity of the wetland, is largely dependent on the water regime and peat depth. A consistently high water table will restrict the tree growth by anoxic conditions and a thick peat layer will obstruct the roots from effective nutrition supply. Generally the wetlands with a moving groundwater or surface water are more flourishing than if it is still, since a moving water gives a supply of nutrients and oxygen to the substrate but also transports organic matter from the substrate, leading to a less anoxic state. Sloping discharge areas with a shallow groundwater table or even a flowing spring water, as well as occasionally flooded riparian and shoreline wetlands with a highly fluctuating water table, are often both rich in species and productive since these types often give both a substantial supply of nutrients and oxygen. However, compared to mineral soil with moderately deep groundwater, the primary production is still lower. On the other hand, swamps with stagnant water are poor in species and have a very low productivity.

The alder swamps are typically covering intermittently flooded areas with mineral rich soils. During snow-melt or long periods of rain the ground is inundated or soaked, either by groundwater or by surface water from an adjacent stream or lake. The alder swamps are now covering only small areas along lakes and rivers and in smaller topographic depressions. However they are considered to have dominated the flat country of middle Sweden which nowadays is traditional farmland. This transformation has taken place since these areas often are associated with fertile soil and are accordingly brought into agricultural use and/or affected by drainage. To a large extent, current lake-side or sea-side marshes which are open and used for harvesting and grazing would also be forested wetlands, if not taken into use. In northern Sweden, the corresponding areas typically host societies of *Salix spp.* These are especially frequent in the subalpine and subarctic zones /Sjörs, 1956/. They occur primarily on flooded or spring-flooded areas, close to watercourses and at spring-influenced borders of mires. The substrate can be both peat, sediments or just ordinary mineral soil, but the *Salix* swamps seem to prefer at least a fair supply of mineral substances and are often more dense in richer sites.

3 Wetland development and forms

3.1 Influence by topography, hydrology and climate

There are some major factors behind formation and development of wetlands. The groundwater table has to be close to the surface at least some periods. There has then to be a water supply and topographic or geological settings which allow the water to come close to the surface. This could be solved by plentiful lateral water supply and/or a poorly drained surface. Most wetlands consequently occur at low points in the terrain where the drainage is impeded by low hydraulic conductivity or an absence of drainage into the surroundings.

In most of Sweden the geology is dominated by precambrian bedrock overlain by glacial till. The bedrock often has low permeability and so has the till, except the surface layers. This leads to 'bad drainage' and these areas often have the groundwater table close to the ground surface, meaning that the groundwater table closely follows the topography. Local heights are often associated with groundwater recharge areas and local hollows are associated with discharge areas. The local heights and hollows together are also part of a larger groundwater system where they constitute a recharge or a discharge area of a regional aquifer. Wetlands are normally only located to discharge areas, though they may occur at recharge areas where the water supply is larger than the ground's capacities of infiltration and drainage. The local and regional flow paths of the water flowing out at a discharge area set the water chemistry and the flow regime. Generally, discharge areas in the higher parts within a watershed receive less regional groundwater flow, whereas the lower parts receive more regional flow, with influence from deep sources. The regional flow involves longer flow paths and the water is more aged and has passed deeper layers in the ground. Hence are wetlands in the low parts of a watershed often more rich in minerals, even if they are not in direct connection with lakes or streams. Since the recharge area is larger, the regional flow is also more even than the local flow. Eskers and glaciofluvial deposits constituted by thick layers of material with high hydraulic conductivity have a deeper water table, which varies less with surface topography, compared to thin-layered till soils, and also often has an extensive recharge area. The discharge from these aquifers are therefore often stable with time.

The Swedish climate renders a surplus of precipitation over evapotranspiration (though there is often a deficit during summer months). This facilitates wetland formation without lateral water supply in conditions where the ground is badly drained, due to low infiltration capacity, low hydraulic conductivity and flat surface. In parts of Sweden, the potential for wetland formation is therefore large.

Near-coast marshes originate by effect of inundation by the sea and, depending on the salt content of the sea water, can thus have a somewhat salt-influenced ecosystem and soil chemistry. Along the eastern Swedish coastline there are only small effects of the tide, in short terms is the water table normally at a stable level, but can vary by a metre between different weather situations. In a long-term perspective, new shoreline wetlands form as the sea gradually rises and inundates the soil surface or as the sea lowers and uncovers new land surface. Since many of the shoreline areas raising from the sea are flat and covered by fine-grained sediments together with organic detritus, they stay badly drained and become often directly covered by peat /Jacobson, 1997/.

3.2 Resulting wetlands, Development

Succession between different wetland types is not following any universal pattern but varies with the impact from different factors. The 'final' distribution of different wetlands is often the result of the local climate and geological conditions. For the succession, and especially occurrence of peat formation, the concept of 'openness' (see section 5.6) can be useful. Where vertical fluctuations of water table are small, peat growth through autogenic succession is possible. Where water movement through the wetland is great or the oscillations of the water table are large, peat formation is limited by the high rates of decomposition of any organic material that begins to accumulate. In Sweden, the peat forming types prevail /Löfroth, 1991/. However, in East Sweden, and especially in the south-eastern part, the precipitation is probably too low to provide peat formation at locations that are not supplied with lateral water inflow.

3.2.1 Peat formation

Peat formation is dependent on several factors:

- **Climate:** the wetness (precipitation-evapotranspiration) is the major factor for peat forming.
- **Relief:** peat formation is enhanced by glacial landscapes with low permeable thin soil layers, glacial-lake beds and kettle-hole topography, gently sloping topographies with blocked drainage. The wet conditions necessary for peat growth can be found in discharge areas and in frequently surface-water flooded areas.
- **Geology:** igneous silica bedrocks generally have lower permeability and generate fewer minerals than sedimentary (especially calcitic) bedrocks, hence they are more favourable for peat forming environments.
- **Biota:** peat forming vegetation especially *Sphagnum* spp., sedges, wetland trees. At the upland margin, *Polytrichum commune* often indicates the beginning of peat formation.
- **Time since glaciation or emersion out of the sea.** Thicker peat layers seem to enhance a transition from fen into bog.

There are three principal types of mire formation:

1. **Primary mire formation,** the fresh soil surface is occupied directly by mire vegetation after emergence from water or ice. Primary peat production will occur at the spots that are wet during sufficiently long parts of the growing season to benefit mire vegetation. At the same time as the peat grows (in height), it most likely also expands laterally, either by lake filling (terrestrialisation) or by paludification. This mire formation is common along the coastal shoreline in Uppland and in northern Sweden.
2. **Terrestrialisation,** a hydrosere succession from an open water basin into a mire /Korhola and Turunen, 1996/. Two formations are given by /Mitsch and Gosselink, 2000/: terrestrialisation by filling in a lake basin, quaking peatland succession, where mosses together with floating or half rooted vegetation cover produces a peat mass which gradually develops from the edges towards the middle of the lake. The peat growth from the borders facilitates a colonisation of first shrubs and then trees, forming rings towards the centre. After some peat development it becomes thick enough to isolate the surface from groundwater influence. Some parts then get more nutrient poor and turn finally into a raised bog. The other terrestrialisation process mentioned by /Mitsch and Gosselink, 2000/, is the detritus sedimenting at the lake bottom, hence building it up enough that marsh vegetation can grow and continue the build up of peat, enabling a transition into

a fen and later into a bog. In fact many lakes are more or less evenly shallow, and an instant peat growth is possible. The floating-mat littoral with *Sphagnum* coexisting with the aerenchyma of higher plants is formed only in the smallest lakes, while in larger lakes emergent plants, e.g. *Phragmites*, grow in the sediments and form a border between the shoreline mire and the open water /Brunberg and Blomqvist, 2000/. The probable reason is that the shoreline of larger lakes more often get disturbed by wave action and the mire edge along shores need to be protected by the macrophyte vegetation to thrive. The mire colonisation is then taking place as a successive lake filling until the emerging plants take over the whole area or the lake area become so small that floating mat colonisation is possible /Brunberg and Blomqvist, 2000/. In addition to incoming (allochthonous) material, the material produced in the littoral zone is washed out and accumulates at the bottom of the lake. In small lakes in an area of northern Uppland /Brydsten, 2003/ found that the size of the lake catchments was of very little importance for the sedimentation rate, indicating that the intransport is of low importance and that most of the accumulating material is of lake origin (autochthonous). In shallow, not too poor lakes the transition into mires often involves a period of *Phragmites* or *Scirpus* in all parts of the lake as it may colonise the bottom at all depths less than 2 m. The latter form with a more or less concurrent mire formation all over the lake may form more horizontally homogeneous peat layers, whereas the infilling of a lake basin from the detritus of a quaking shoreline mire probably gives a laterally differing mud and peat constitution. The mean rate of lake filling is about 1 mm year⁻¹ in shallow oligotrophic-slightly eutrophic lakes in Uppland /Brunberg and Blomqvist, 2000/. The deposited sediments are dominated by gyttja or clay gyttja in lakes with abundant vegetation, whilst incoming water from areas rich in wetlands and forests often have a high content of humus which give cause to sedimentation of dy upon lake bottoms /Brunberg and Blomqvist, 2000/.

3. Paludification, the conversion of mineral soil ecosystem into a mire ecosystem. The terrestrial ecosystems are overgrown by mire forming vegetation. This process can be brought about by climate changes, geomorphological changes, beaver ponds, forest logging, but more often by a natural advancement of peatlands. The low permeability of adjacent peat layers contributes to a rise of the groundwater table, which supports the lateral expanse of mire vegetation. The paludification is considered as the predominant way of mire formation in the wetter parts of Sweden /Rydin et al, 1999/.

The factors that determine which of the different mire formations that dominates in different situations are not totally clear but generally the peat growth by terrestrialisation is greater during warm periods whereas the paludification is greater during cold and wet periods /Korhola and Turunen, 1996/. Correspondingly should the terrestrialisation process be more common in warm, dry regions whereas paludification dominates in colder and wetter climates. Climate, geologic and hydrological settings of a site thus constitute the basic conditions to be considered for estimation of peat growth and peatland development.

The growth of peat and mire development is, however, hard to predict. A natural succession from (lakes or marshes into) fens and into bogs as a final result seems to be valid in most areas in southern and mid Sweden /Rydin et al, 1999/ as well as in southern Finland /Korhola and Turunen, 1996; Rydin et al, 1999/. The influence by autogenic (internal factors such as production, decomposition and peat type) and allogenic (external factors such as climate, topography etc) factors for mire formation and peat growth have been studied and are often discussed. These factors are normally connected in different ways. The climate is often seen as a major factor for peat growth and was earlier used as an explanation of found variations of colour and degree of decomposition between different layers within the peat, suggesting that the change into a colder and wetter climate would cause a less decomposed and lighter peat /e.g. Frenzel, 1983/. However, dating techniques

have showed that these patterns are often different in different mires and even within the same mire, and therefore it is hard to couple certain type of peat formation with climate conditions, although the mire surface probably changes its properties and features with changes in the climate.

The type of mire, fen or bog, and the shape of the mire, which forms in the subsequent peat growth, depends largely on the hydrological settings. Almquist-Jacobson and Foster found, in a study of 7 mires in Bergslagen, that fen initiation appears to be primarily a function of local conditions of soil or “hydrogeography” rather than as a result of climatic conditions /Almquist-Jacobson and Foster, 1995/. A reduction in the permeability of the underlying soil e.g. by accumulation of humic substances could then have set on a fen initiation, by keeping a constantly high wetness. Both site topography and regional water balance affected the rate of the subsequent fen expansion. In their study, /Almquist-Jacobson and Foster, 1995/ claimed that fen peat growth was greater in early Holocene than in current time, i.e. the fen peat growth should be greater during moist climate periods, although they did not present any figures of fen peat growth. They also found that lateral peat growth over mineral soil was inhibited by too effective drainage at locations where the slope was more than 0.5%. The conversion of fen to bog likely results from a decrease in nutrient-rich slope-wash water reaching the site, which may reduce microbial activity and decomposition rates. As the fen peat grows thicker, it will impede the incoming groundwater and surface water flows and some parts will get successively less influenced by these lateral water flows compared to rain water supply. This situation brings on a poorer condition, that also facilitates a more competitive situation for the *Sphagnum* moss. This situation of hydrologic isolation may then, in addition to a decreasing gradient in hydraulic head (because of fen’s vertical growth), also be due to an increasing distance from the uplands to the mire centre (because of fen’s lateral expansion). Mires formed at the base of steeper slopes have a thicker accumulated fen peat before turning into bog than the mires lying on flatter substrates /Almquist-Jacobson and Foster, 1995/. However, /Almquist-Jacobson and Foster, 1995/ could not find any quantitative (predictive) relationships between fen age at the bog initiation and e.g. distance from the point to the uplands or hydrologic gradient from the upland to the mire centre. After establishment, the *Sphagnum* subsequently turns the peat to be more acid and hence even more favourable for the successive *Sphagnum* development. The greater influence by *Sphagnum* leads to slower decomposition and enhanced peat growth. The conversion into a bog may be seen as a second paludification where the growth of a bog cell locally builds up a groundwater mound. This brings on a rise in hydraulic head at the surface, which causes a downward gradient. Together with the acidification of the successively dominating *Sphagnum* moss it turns the surroundings more acid, facilitating the *Sphagnum* to take a more dominating role and hence slowly turn the surface into a bog.

In the study of /Almquist-Jacobson and Foster, 1995/, the bogs seemed to have expanded at the same rate since their start. Thereby, up to now, the local hydrology and topographical conditions seemed to have played a greater role than climate for the mire formation.

3.2.2 Models for peat accumulation and mire developments

The peat growth and mire formation is a result from complex links of processes and are difficult to predict. Conceptually there are two major aspects of peatland growth: control by external factors such as climate and local basin hydrology, and control by internal factors such as production and decomposition. Considering the production and decomposition, /Clymo, 1984/ suggested that the peat formation in the surface layer was fairly constant with time, i.e. that the primary production minus decomposition in the aerated zone was constant. The slow peat decay in the anaerobic zone beneath could then be assumed to be constant with time or decrease with peat depth (age). This approach agree with the

observations that in most mires there is a general pattern with faster growth in early years than in later periods, since the low rate of decay in the anaerobic zone earns more importance as the peat mound grows thicker. /Clymo et al, 1998/ used these assumptions for a conceptual model of peat growth when comparing data of numerous mires in Finland. For a constant anaerobic decay rate of all layers, the final peat depth will only depend on the ratio between the peat production in surface layers and decay rate in deeper layers. For a decay rate that decreases with age of the peat, where the decay rate of the peat is linearly proportional to the ratio between remaining mass and original mass, the dry mass per area unit M [mass/area] is expressed as /Clymo et al, 1998/:

$$M = (p/a) \times \ln(1 + at) \text{ and } \partial M/\partial t = p/(1 + at) \quad (3-1)$$

where t is time, p is rate of addition of dry mass [mass/(area×time)] at the surface layer and a is the (starting) proportional decay rate for the peat [time⁻¹]. The anaerobic decay rate aM will then increase with thickness of the peat and the total depth will finally (when t approaches infinity) be proportional to the logarithm of mire age. This simple model assumes constant biological activity with time (a and p constant with time). This assumption could not be rejected in the study of /Clymo et al, 1998/ though they identified differences in growth rate between different regions, where also the resulting mire types were different. An exponential increase of deep peat decay with average temperature should then be expected, whereas /Clymo et al, 1998/ suggested a linear dependence between p and yearly degree-days i.e. accumulated daily values of temperatures above 0°C. This model can be useful, although there is lack in knowledge on how to accurately estimate assimilation and decomposition in environments with different nutrient supply. /Frolking et al, 2001/ developed this concept further into a more parameterised operative model, which considers vascular and non-vascular plants, differences in climate and production. This model is still static with one-way relationships and constitutes a very simplistic description of peat growth. Better models with a more dynamic description of peat production, climate and water table depth are coming /Hilbert et al, 2000/ but for long term modelling predictions, the uncertainty of climate variation makes the addition of better parameterisation useless.

There are also models describing size and shape of peat mound of bogs /e.g. Ingram, 1983; Kneale, 1987/, taking the groundwater hydrology as a starting point. Already /Granlund, 1932/ noticed a connection between yearly precipitation and bog surface shapes, with more convex mires in more wet areas. This can theoretically be described. A groundwater mound between some boundary sinks is formed when recharged water from the surface is draining laterally to the sinks. For a bog, the mound is sustained at a high level by the low conductivity of the peat. The parameters controlling the height of the mound are then the net recharge and the hydraulic conductivity. By using standard saturated flow theory based on Darcy's law and using the Dupuit-Forchheimer approximation (the flow is assumed to be limited to the horizontal direction), the shape of a groundwater mound can be described for different geometrical shapes of the borderlines. For a circular peripheral border the maximum height Z_m of the bog mound is given by /Ingram, 1983/:

$$\frac{U}{k} = \frac{2Z_m^2}{L^2} \quad (3-2)$$

where U is net annual recharge into the peat (i.e. precipitation – evapotranspiration – surface runoff), k is peat hydraulic conductivity and L is the radius of the circular mire. Some scientists have studied the possibilities to predict the size and shape of current bogs with this simple approach /Kneale, 1987/. However, although there are strong correlations between the factors mentioned above and the peat mound size, the natural variation of surface conditions makes this approach too simple. The approach is resting on several simplifying assumptions, among which some are: the underlying ground must be

impervious and horizontal; the hydraulic conductivity is homogeneous in all directions; steady state must be valid i.e. the inflow has to be in equilibrium with the outflow. These assumptions are very seldom fulfilled but the concept of hydrological functions setting the conditions for peat mound formation may still be useful. /Bromley and Robinson, 1995/ recommend numerical distributed groundwater models as a powerful and comprehensive tool for evaluating the hydrology and its influence on the processes responsible for growth and stability of peat systems. However, since the understanding of mire processes is still on a conceptual level, the models need to be calibrated for real case studies. /McNamara et al, 1992/ examined a small wetland complex, containing a diversity of peatland types, by measurements of hydraulic head gradients, water chemistry, peat physical properties and surface topography. From the peatland measurements, they could calibrate the settings of the mineral subsoil flows and heads. In the results they found very large spatial differences in surface chemistry between closely situated rich-fen areas and bog or poor fen spots. They could show by modelling that bog initiation started by a fen growing and forming a mound, where the water table was kept a little higher than in the surrounding fen. Hereby, only a 20 cm high mound would cause a vertical hydrologic reversal and divert the regional groundwater discharge, even when the surrounding mineral-soil water table was meters above the fen surface. The soil chemistry at the top of the mound then becomes gradually more influenced by precipitation than by groundwater, and bog vegetation is initiated. Since bog growth is documented to be significantly faster than fen peat growth, it has been suggested that the fen is 'bogged' down, i.e. the fast growth of the bog mound is successively making the fen surface layers poorer, and then initiating bog vegetation all over.

/Brydsten, 2003/ found that terrestrialsation of hard-water lakes in northern Uppland, around the area of Forsmark, can be predicted by a simple model of lateral peat growth. The model describes the succession from sedimentation of sea bed, isolation into lake and subsequent terrestrialsation by *Phragmites*, *Carex* etc, and finally *Sphagnum* peat. The threshold depth for onset of terrestrialsation was selected to 2–2.1 m because this is the maximal depth for *Phragmites* colonisation. After the lake isolation, the time to *Sphagnum* colonisation is 3–5000 years. The constitution of the material finally filling up the lake basin were about 10% outwashed sand, 40% gyttja clay and 50% peat. After *Sphagnum* colonisation, the lateral peat growth was linearly related to the original size (area) of the lake basin. The topographic and geological settings are very similar among these lakes and are probably the cause to the success of this relatively simple model and the close relationship between lake size and peat growth can most probably be found in that fact.

4 Physical properties of peat

4.1 Degree of decomposition

The physical properties of peat are largely connected to botanical composition and degree of decomposition. During decomposition, the peat changes structure from the fresh plant material successively into amorphous humic matter. In the first stage, the degree of decomposition of the peat matrix is governed by the nature of the environment in which the plant material was initially deposited and by the subsequent position of the water table. In the second stage, when the peat is permanently waterlogged, the decomposition is very slow and the degree of decomposition depends mainly on the rate and duration of decomposition in aerobic conditions when the peat in question formed the upper layers of the mire. There are many methods for determining the degree of decomposition, more or less objective and reproducible /Eggelsmann et al, 1993/. The most well-known method is the von Post humification scale /von Post and Granlund, 1926/, which is used in various applications. This technique involves taking a sample of raw peat which is squeezed in the closed fist to determine if water, of different colour and turbidity, or slurry-like material is expressed between the fingers. The result is expressed in degrees of humification on a 10-point scale (Table 4-1). Although it is a subjective method, its simplicity, speed and ease to use in field has made it most common. Taking the heterogeneity of peat into consideration, it has proved to be adequately reproducible.

Table 4-1. Von Post humification scale /from Clymo, 1983/.

Scale number	Description	Proportion of "dy"	Plant structure	Expressed fluid	Peat lost by squeezing	Peat retained in hand Consistency	Colour
H1	Completely unhumified	None		Colourless, clear	None		
H2	Virtually unhumified	None		Yellow-brown, clear	None		
H3	Little humified	Small		Noticeably turbid	None	Not porridgey	
H4	Poorly humified	Modest		Very turbid	None	Somewhat porridgey	
H5	Fairly humified, distinct structure	Fair	Plain, but somewhat obscured	Strongly turbid	Some	Very porridgey	
H6	Fairly humified, less distinct structure	Fair	Indistinct but still clear		Up to 1/3	Very porridgey	
H7	Quite well humified	Considerable	Still visible		About half	Gruel-like	Very dark
H8	Well humified	Large	Vague		About 2/3	Only fibrous matter and roots remain	„
H9	Almost completely humified	Most	Almost none visible		Almost all	More homogeneous	„
H10	Completely humified	All	None visible		All	Porridge	„

Other methods try to quantify the decomposition by visual estimation of the relative proportions of plant structures and humic substance. This is the most common method in Russia and other eastern-Europe countries, but also in North America, where content of fragments > 0.1 mm determine a division between Fibric (> 2/3), Hemic (1/3–2/3) and Sapric (< 1/3) peat /Farnham and Finney, 1965/.

4.2 Botanical composition and plant remains

Generally, bog peats are less decomposed than fen peats. Apart from the common shallow aerobic zone, the slow decomposition of bog peat come from the acid conditions and low abundance of nutrients. Peat mosses have shown to keep their shape and cell structure in weakly decomposed peat, wherein vascular plant parts can leave more decomposed strings and lenses. In more decomposed peat, only stems and leaf ribs are left of the mosses too. For the higher plants, most material is readily decomposed except the robust, thick-walled tissues, such as the outer bark of the roots, the epidermises of plant tissues. Wood remains often seem completely unaltered but cell walls are often significantly thinned and the wood is found to be fragile and shrinks considerably at drying /Eggelsmann et al, 1993/.

Mosses die at their base as their tops continue to grow. The dead lower parts sink down into the underlying peat layers while they are pressed together. The erect and tightly interwoven moss plants growing in hummocks often keep the structure even if they are squeezed vertically and sometimes crumble irregularly under the downward stress. Moss species in hollows' carpets often form a peat layer with lying parallel moss stalks. Depending on the shifting surface conditions with time, the structure of the bog peat can vary with depth, with more or less decomposed mosses oriented in different directions, varied with layers of more amorphous character, but also containing woody debris of stalks and root parts. The above-ground parts of vascular plants are seldom preserved, since these are deposited and readily decomposed in aerobic conditions. What is left then is the below-surface parts of the stems and roots, especially those who penetrate into the anaerobic layers. There are sometimes layers with extensive root remnants of *Eriophorum vaginatum* and/or *Scheuchzeria palustris*, which make them appear similar to Sedge fen peats.

In fens, whose vegetation is dominated by higher plants, the peat accumulation is not confined to the peat surface as the roots penetrate into old peat and then form a 'mixed' peat with new and old organic matter. The decomposition is more intensive than in bogs because of less acidic conditions and more available nutrients. The structure in peat becomes a mixture of highly humified matter and clearly identifiable remnants. Occurrence of woody debris can be scarce or plentiful, mostly only smaller roots and stumps appear. Fen peats can be divided into reed peats, moss peat and sedge (root-mat) peat. Reed peat is characterised by the roots and rhizomes of *Phragmites australis*. These are often weakly decomposed (\approx H4) and found at the 'border' between peat and mineral soil, especially at terrestrialised mires. The term sedge peat is often applied to a peat type in which fine rootlets are found and where it is likely that *Carex* spp. have been a dominating feature. Often can also Bogbean (*Menyanthes trifoliata*) and Horsetail (*Equisetum* spp.) remnants be found, which can cause the peats to be called after these species.

4.3 Physical properties

The structure of the peat depends naturally on the parent material, the subsequent mixing and decomposition. Generally is the volumetric proportion of peat material small and porosity of natural, undrained peat is normally 80–97% of the volume. Decomposition diminishes the proportion of large continuous pores by tearing down the loose structure and increasing the proportion of fine particles. Thus the pore structure and pore size distribution changes. With a concurrent increase in pressure as the peat gets buried beneath newer layers, the porosity also decrease. The decrease of large pore proportion makes the peat tighter (less permeable) with decomposition. This can also be expressed as the dry bulk density ($\rho_b = \text{mass of solids} / \text{total volume}$) increases. The porosity of peat is often determined by determining the dry bulk density since the pore volume is difficult to deduce in other ways. The porosity p is then:

$$p = 1 - \rho_b / \rho_s \quad (4-1)$$

where ρ_s is the solid particle density. The particle density of organic matter is not absolute but variation is not great. /Päivänen, 1973/ found a highly significant correlation between water content of saturated peat (y , in volumetric percent) and its dry bulk density (x , in g cm^{-3}) as

$$y = 97.95 - 79.72 x \quad (4-2)$$

The organic matter both contains pores within the dead plant tissues (internal) and forms pores between them (external). The internal pores in fresh peat correspond to the pores within plant tissues and external pores the spaces between plant parts /Figure 4-1; Romanov, 1968/, whereas in decomposed peat it is harder to make this distinction between these two types. The higher the decomposition of the peat, the lower the volume of internal pores and the higher the proportion of “true” external pores /Hobbs, 1986/. For water flows, one interpretation is that the fast transferring large pores decrease with decomposition but also that the ‘locked’ internal pores decrease as the plant structure collapses, which mean that a larger fraction of the pores take part in the flow which becomes slower but concurrently more evenly distributed. However the more common view is that the part of dead-end pores increases with decomposition /Hoag and Price, 1997/. This is probably an effect of clogging by the increased proportion of humic particles with decomposition, which makes only some pores effectively conducting /Loxham, 1980/. /Loxham, 1980/ claimed that decomposed peat is a media with plant debris intermixed with amorphous peat particles together forming a complex structure with many dead-end pores and with few continuous pores of larger diameter. Besides, woody material residues can make the peat more heterogeneous. Cavities can occur in association with these and, although fen peat is more decomposed and dense compared with bog peat, it can contain more wood pieces and hence more continuous, large pores than bog *Sphagnum* peat. Typical values for particle density, dry bulk densities and porosities are given in Table 4-2.

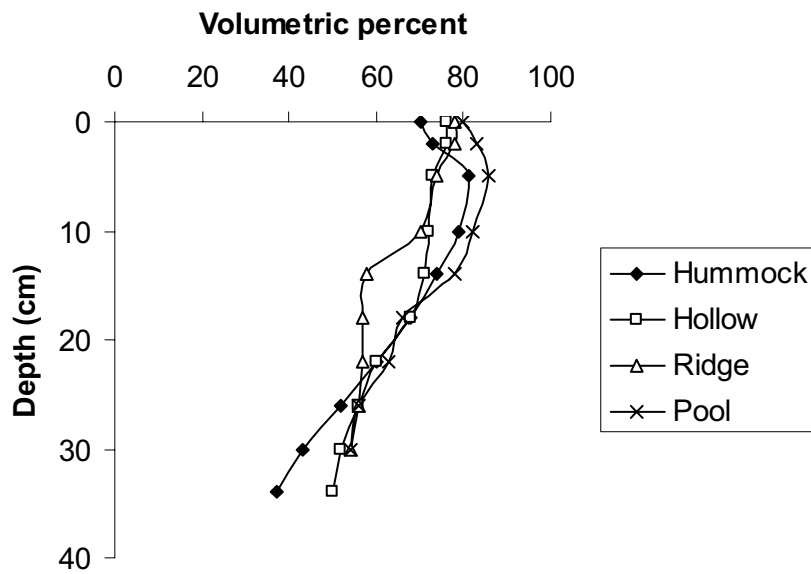


Figure 4-1. External porosity in *Sphagnum tigr* in the upper layers. Data from a hummock-hollow bog and a ridge-pool complex mire, both in Russia. Data come from /Romanov, 1968/, who claimed that the decrease of porosity with depth is caused only by compaction, not by decomposition processes.

Table 4-2. Typical values for particle density, dry bulk density and porosity in peat soils /after Eggelsmann et al, 1993/.

von Post number	Compaction	Dry bulk density ρ_b (g cm ⁻³)	Particle density ρ_s (g cm ⁻³)	Porosity (%)
H1	Loose	0.05	1.4	96
H4	Loose to dense	0.10	1.4	93
H6	Dense	0.15	1.4	89
H9	Very dense	0.20	1.4	86

5 Hydrology

5.1 Water budget equation

For most wetland systems the hydrology is of uttermost importance. The hydrology sets the conditions for the wetland by the water level and by different transport mechanisms. The hydrologic state of a wetland can be summarised as the result of the balance between inflows and outflows of water. The hydrological system can be described by the water balance equation, which can be simply expressed as:

$$DS = P - E + SI - SO + GI - GO$$

Where DS is water volume stored (change in water volume), P is precipitation minus canopy interception (interception is the part of the precipitation that is caught by the canopy and evaporates back into atmosphere without reaching ground), E is evapotranspiration, SI and SO is surface-water inflow and outflow, respectively and GI and GO is inflow and outflow by groundwater.

5.2 Storage and water table

The water volume storage is tightly coupled to the level of the water table (identical to the water level in inundated wetlands). The relation between storage change and change in water table level can be expressed by the storage coefficient (or storativity) $SC = DS / DWT$, i.e. the relation between the change in amount of water per unit area and the change in water table level. The storage coefficient decreases normally with decreasing soil particle diameter and it is smaller for soils with high water holding capacity like clays and silts and low for coarser soils. In wetlands, the storage coefficient is not constant with water table level, but generally it decreases with increased water table depth (when water table is above ground surface, SC is in principle unity). For peat soils, it decreases with degree of decomposition and with depth (Table 5-1).

Table 5-1. Variation of storage coefficient with peat type.

Author	Peat quality or type	Storage coefficient
/Boelter, 1969/	Fibric bog peat	> 0.42
/Boelter, 1969/	Mesic bog peat	0.15–0.42
/Boelter, 1969/	Sapric bog peat	< 0.15
/Päivänen, 1973/	H1 ($\rho_b = 0.05$)	0.60
/Päivänen, 1973/	H4 ($\rho_b = 0.1$)	0.36
/Päivänen, 1973/	H6 ($\rho_b = 0.15$)	0.22
/Päivänen, 1973/	H9 ($\rho_b = 0.2$)	0.18

There are two complications. The mentioned values of storage coefficient in table are established from measurements on samples taken into laboratory. In these systems with water table close to the surface, the ground surface is seldom completely flat. Consequently, while parts of the wetland surface can be some decimetres over water table, others can be inundated. The wetland-scale storage coefficient will then be different from that estimated for the single samples. The other complication originates from the elasticity of the peat. In submerged conditions, the soil is partly carried by the water by buoyancy forces. When some of the surface layers rise above the water table during drying sequences, there will be shrinkage in these layers partly because of the absence of support from the free water and partly because of increased tension within the soil matrix. However, another (and probably the largest) effect of compaction will take place in the still submerged part, since these layers also experience a greater pressure from the unsaturated layers. The effect of this phenomenon will be that the surface follows the movements of water table. The total storage coefficient SC can thus be divided into specific yield S_y (the depleted water is substituted by air in the pores) and compressive specific storage S_s (volume change in peat matrix).

$$SC = S_y + b S_s \quad (5-1)$$

where b is peat thickness. There are few measured values of the size of S_s . It probably depends on the type of peat and probably also shifts with time. It have been found to vary in the range 10^{-4} – 10^{-3} cm⁻¹ /Knott et al, 1987; Nuttle et al, 1990; Schlotzhauer and Price, 1999; Kellner and Halldin, 2002/, although the factors influencing on these values are still unknown. The effect of the compressive storage is thus difficult to predict but /Kellner and Halldin, 2002/ found that about 40% of the summer-season storage change in a mire with 3.5 m thick peat layer took place through compressive storage. For a typical summer water deficit (evaporation exceeds rain) of 100 mm, only 60 mm was depleted from the surface layers and the rest from compression of deeper layers. The groundwater table lowered then 250 mm i.e. the total storage coefficient SC was about 0.4, of which specific yield was 60 mm / 250 mm = 0.24.

A good estimate of SC is necessary as well as considering the peat compaction when using the water balance equation, since the water table level is crucial for the geochemistry and for the vegetation development and plant ecology.

5.3 Precipitation and Evapotranspiration

Precipitation minus evapotranspiration govern the upper border conditions for the hydrological system of the wetland. In Sweden, precipitation follows a slight seasonal variation with maximum in July–September. The greatest amounts appear later in the coastal areas and in the western, Atlantic areas the maximum is reached in late fall. Generally is also February–April the driest period during the year. However, the variation from year to year is very large, in both timing and amount of the precipitation. It is also hard to tell if there are any significant long term changes of precipitation since the methods of measurements have changed over time /Bergström, 1994/. The precipitation pattern is similar along the east coast from Kalmar to Luleå as well as the yearly amount, the increase with altitude is 10–20% /100 m. The distribution between rain and snow is though different with about 20% of the precipitation coming as snow in the southern and about 40% in the northern part of eastern Sweden. In northern Uppland, the percentage is 25–30%.

There is always some part of the precipitation which is not reaching the ground but is caught by the vegetation and then evaporates back into atmosphere. The amount of water caught by this process, called interception, differs between different vegetation types. More living

biomass, i.e. trees and herbs, generally means that a larger part of the precipitation does not reach the ground. The size of the part of the precipitation that is intercepted depends also on the conditions of the rain storm, since a lesser part of a single but heavy rain storm is stored in the canopy than of several small rain amounts. There are no studies on the size of interception especially made on wetlands. In studies on growing-season averages of other ecosystems, the size vary between <5% for sparse low vegetation as heather (*Calluna*) to 40% in a dense forest /Lundberg, 1996/. In a mature coniferous forest outside Uppsala the interception losses during the period May–October is 25–30% of the total precipitation /Grelle et al, 1997/. Since the storage capacity of snow is significant for coniferous trees, the interception evaporation can be substantial also during winter if precipitation falls as snow. Several investigations stress the importance of snow-interception evaporation in coniferous forests and evaporation fractions of gross precipitation as large as 20–50% have been observed by investigators in Scotland, Canada, and Japan /Lundberg and Halldin, 2001/. The influence on wetlands by this difference in winter precipitation is not known, the largest difference between snow and rain is the larger storage of snow on ground and in vegetation, which cause a delay in groundwater recharge and may increase the evaporation losses as by interception mentioned above. Hence, there could be a possible effect in a different water table regime.

The size of evapotranspiration from wetlands is as various as the variation among non-wetlands. The general factors influencing on evapotranspiration are the surface conditions such as roughness, temperature and dryness, together with the air temperature, humidity and solar radiation. There are very few measurements of wetland evapotranspiration and they cover only short periods. Since the climatic factors vary much, it is hard to generate absolute values without a great uncertainty.

Evapotranspiration for different surface types is often estimated by calculating potential evaporation, PE, (the evaporation from a water surface or a moist grass lawn, where the supply of water is not limiting), and then to multiply PE with a factor which express the evaporative ability of the surface /Bergström, 1994/. Mostly is the formula by /Penman, 1953/ used for calculating PE. Measured evapotranspiration and the quotient E/PE is given in Table 5-2 for some different wetland types. In Sweden, E is similar to PE during wet conditions and in cold weather but can differ significantly from this during warm and dry conditions in the growing season. During the growing season, a wet surface or a dense canopy give an evapotranspiration close to PE, whereas a sparse canopy over a semidry moss surface as in poor fens and bogs often has lower evapotranspiration. Evapotranspiration from forested wetlands are normally similar to other forested land which is not hampered by dry soil and can be assumed to be close to PE during the growing season. However tree transpiration could be substantially lower if the root functions are reduced by flooding or other impacts.

Table 5-2. Examples of evapotranspiration (E) for different wetland types.

Reference	Wetland site, type, vegetation	Period	E mm/day	E / PE
/Price, 1994/	Lakeshore Typha marsh, Lake Ontario	June–August	4.8	0.97
/Lafleur, 1990/	Lakeshore Carex marsh, Southern Hudson Bay	May–August	2.75	0.8
/Souch et al, 1996/	Lakeshore mixed open marsh, Lake Michigan	June	3.3	1.0
/Kim and Verma, 1996/	Poor open fen, Minnesota	May–October	(0.2–4.7) 3.0	1.0
/Lafleur and Roulet, 1992/	Poor open fen, Southern Hudson Bay Lowland	July–August	2.5	≈0.7
/Kellner, 2001/	Open bog, Central Sweden	June–August	2.1–2.4	0.7
/Moore et al, 1994/	Mixed open mire, Northern Quebec	June–August	2.8	0.8

5.4 Soil water characteristics above water table

The pore size distribution of a soil determines its water holding capacity in unsaturated conditions (when the soil is above water table and hence subjected to water pressure less than atmospheric). A pore's ability to keep water against suction by capillary forces is inversely proportional to its diameter. Therefore, coarse soils keep less water than fine-grained soils when subjected to increasing tension /e.g. Grip and Rodhe, 1988/. Peat soils' pore size distribution depend on original plant material, bulk density and degree of decomposition (Figure 5-1) /Boelter, 1968; Päivänen, 1973; Letts et al, 2000/. In addition to a great heterogeneity, hysteresis effects make it difficult to predict soil moisture distributions of the peat top layers. From the aspect of geochemistry and transport, the region above the water table where the pores are still saturated, (*i.e.* the largest pores ability to keep water is enough to withstand the tension from the sinking water table) should be the most important. This region, called the capillary fringe, rather marks the limit of saturation than the water table. This zone is almost non-existent for undecomposed, fresh surface peat layers, but is often 20–40 cm thick for more decomposed layers. The peat is therefore seldom aerated deeper than 10 cm beneath the mean water table level /e.g. Kellner and Halldin, 2002/.

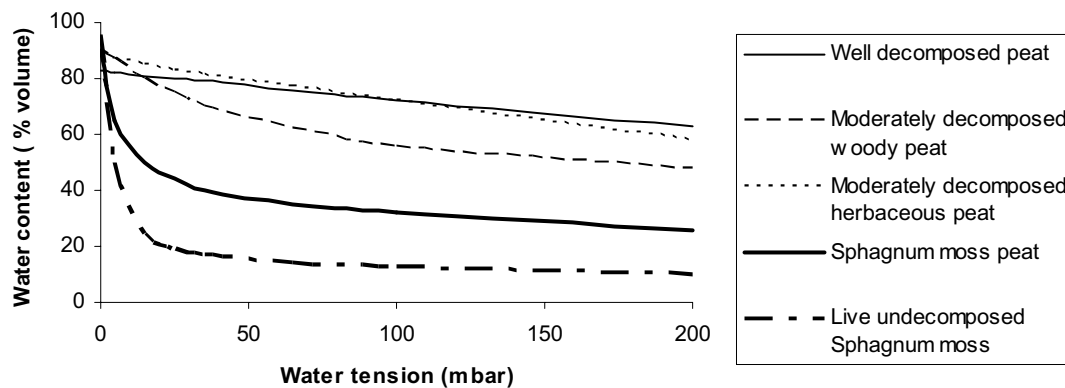


Figure 5-1. Variation of volumetric water content with water tension in different types of peat. From /Boelter, 1968/.

5.5 Groundwater

5.5.1 Darcy's law

The flow of water through a saturated medium is commonly described by Darcy's law:

$$Q = -kA \frac{\partial h}{\partial x} \quad 5-2$$

where Q is the volumetric flux through a cross section area A , k is the hydraulic conductivity (permeability) of the medium, h is the hydraulic head and x is the distance. Hydraulic head represents the energy level, it is generally expressed as energy per unit weight of water to give units of length. The total hydraulic head h is

$h = z + \varphi$ where z is the level above an arbitrarily defined reference datum plane, of the point where h is to be determined, whereas φ express the pressure potential (both h and φ are expressed in length units, e.g. cm). Basic knowledge of groundwater flows and substance transports is given by e.g. /Freeze and Cherry, 1979/.

5.5.2 Hydraulic conductivity

The hydraulic conductivity is often identified as the most important parameter determining subsurface flow in wetlands /Hemond and Fifield, 1982; Knott et al, 1987/. The hydraulic conductivity is highly dependent on the pore size distribution, since the flow velocity is proportional to the fourth power of pore diameter, but also on the regularity in the structure, the tortuosity and continuity of the pores /Hillel, 1998/. Therefore it is well correlated with soil texture and structure. The presence of 'biopores' or macropores such as root channels, worm holes and pores in connection with elongated wood parts are important. Measurements of k in gyttja-soils and semi-organic soils in marshes are scarce /Berglund, 1996/. The published measurements of hydraulic conductivity in peat differ greatly. The traditional view is that k decreases quickly with depth from the surface in mires and that k in the surface peat is thousands to tens of thousands times the k of deeper layers /Ivanov, 1981/. This has shown to be true in the surface layers (0–50 cm) where the conductivity is very high at the surface and then decreases sharply in the 'active' zone of fresh litter with intensive decomposition and often aerobic conditions /Boelter, 1968; Romanov, 1968; Ivanov, 1981/. The size of k in the top layers has been found to be well correlated with degree of decomposition, bulk density and depth (these three factors are in fact well correlated, since both the decomposition and bulk density increase with depth), as well as the plant material (Table 5-3, Figure 5-2), although the variation is great /Päivänen, 1973/. The variation within the plant and decomposition classes indicates that existing methods of classification do not cover the various properties that cause the physical behaviour of the peat. In addition to these variations, there is also often a great spatial heterogeneity which is contributing to the large differences in the size of k , both within studies and among different authors, and difference in measurement methods also influence /e.g. Päivänen, 1973/. Therefore it has shown to be difficult to predict k values in peat only from the information by taking samples and classify them.

Maybe more important for the surface layers, the 'effective' hydraulic conductivity depends much on the surface microrelief, i.e. on the elongation and thickness of ridges and strings and the interconnection between hollows and pools. Such effective values are given by /Ivanov, 1981/ for surface-layer flows in different mire surface types, although they can be expected to vary a lot depending on local conditions.

Table 5-3. Values of saturated hydraulic conductivity (k) measured by auger holes in field.

Reference	Peat quality observation	k (m/s)
/Boelter, 1968/	Fibric bog peat	$>1.8 \times 10^{-5}$
/Boelter, 1968/	Hemic bog peat	2.1×10^{-7} – 1.8×10^{-5}
/Boelter, 1968/	Sapric bog peat	$<2.1 \times 10^{-7}$
/Baden and Eggelsmann, 1963/	Sphagnum H3–H5	1×10^{-6} – 5×10^{-6}
/Baden and Eggelsmann, 1963/	Carex H3–H5	8×10^{-6} – 3×10^{-5}
/Baden and Eggelsmann, 1963/	Phragmites H3–H5	10^{-5} – 5×10^{-5}
/Baden and Eggelsmann, 1963/	Brushwood H6–H10	10^{-6} → 10^{-3}
/Romanov, 1968/	Bog peat, highly decomposed	10^{-8}
/Romanov, 1968/	Fen peat slightly decomposed	5×10^{-5}
/Romanov, 1968/	Fen peat, moderately decomp.	8×10^{-6}
/Romanov, 1968/	Fen peat, highly decomp.	10^{-7}
/Letts et al, 2000/*	Fibric peat	2.8×10^{-4}
/Letts et al, 2000/*	Hemic peat	2.0×10^{-6}
/Letts et al, 2000/*	Sapric peat	1.0×10^{-7}

* These values are derived from the median within each group from a study that covered several literature sources.

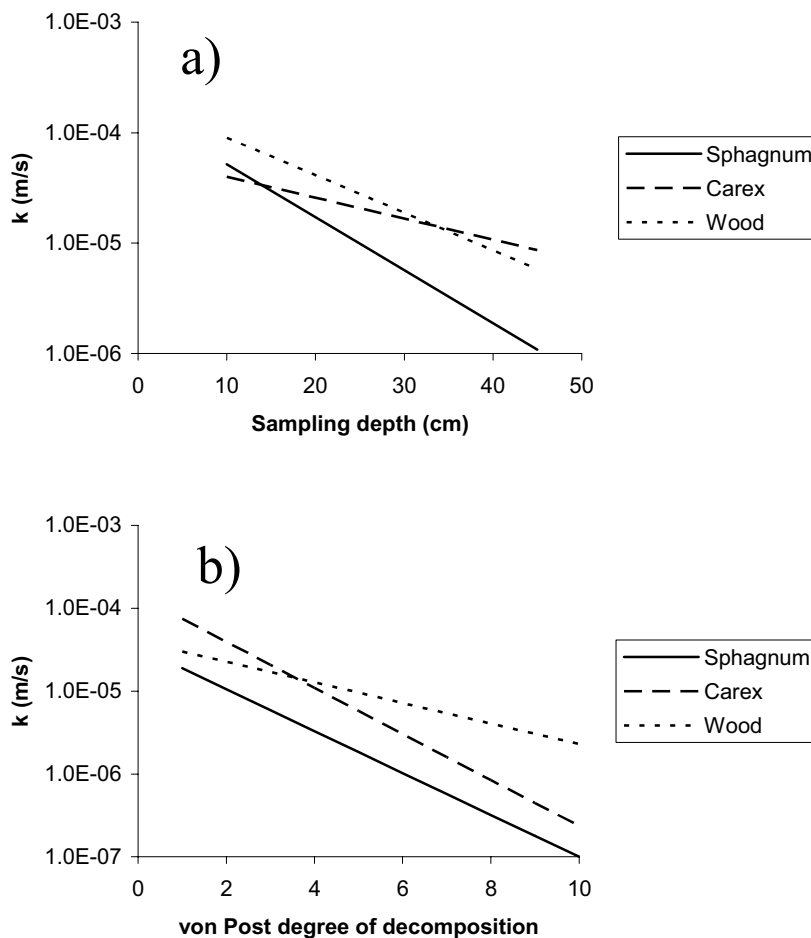


Figure 5-2. Found relationships between hydraulic conductivity and a) sampling depth and b) degree of decomposition (when sampling depth > 25 cm), for some different peat types. After /Päivänen, 1973/.

In the deeper layers beneath, several measurements of k have been done (Table 5-3) showing that values of k vary with degree of decomposition, although conditions of water flows are still relatively unknown. Traditionally, the view has been that the lower layers have a very low effective k and that the water is more or less stagnant /Ingram, 1983/. However, in some recent studies of groundwater flows within wetlands, the results indicate that the peat seldom have such a low k that these layers restrict the vertical flow but it is often the mineral soil layers underneath that determine the size of the flux /Woo and Winter, 1993; Reeve et al, 2000/. Again, the available methods for description of peat do not cover the hydraulic functions satisfactorily, since not only the plant material and degree of decomposition is important. Also the composition of layers, pore structure and presence of pores in connection with branches and other macro-detritus become more important in deep layers where k is generally low. Since there is still a lack of reliable and generally applicable methods to describe these structures in peat, it is both hard to compare different studies and to predict properties for a special fen. Since the elasticity of the peat causes changes in volume, the conducting pores also change size and hence the hydraulic conductivity changes significantly, possibly several orders of magnitude /Chow et al, 1992; Price, 2003/. Apart from the effects of changing porosity with water level/pressure also volume changes in trapped gas bubbles occur as water pressure changes with time /Baird, 1995/, influencing greatly on the hydraulic conductivity. In studies comparing natural peat profiles and sterilised dittos, the formation of methane bubbles during biological active periods has shown to reduce the hydraulic conductivity, from one /Beckwith and Baird, 2001/ to several /Reynolds et al, 1992/ orders of magnitude. The elasticity of peat has also been identified to cause large uncertainties in the measurements of k /Baird, 1995/.

With accordance to the earlier description of peat (at section 4.3) as having 'open' and 'closed' pores, some investigations indicate that the peat can be considered to be a dual porosity medium where water moves preferentially through 'active' macropore spaces, corresponding to open or external pores /Ours et al, 1997; Hoag and Price, 1997/. The distribution of flows between different pore sizes is an important issue for estimation of substance transports, and a lot of studies need still to be done. There are also found indications of 'channel' flows in some peats /e.g. Norrström and Jacks, 1996/, where macropores conduct large flows.

The horizontal and vertical hydraulic conductivity can also have significantly different size (Figure 5-3). Peat deposits often accumulate as layers of plant residues which exhibit lamination and are commonly anisotropic with different hydraulic conductivity in horizontal and vertical direction. A horizontal layer with less decomposed peat than adjacent layers, and/or a having a structure with horizontal main direction of the pores, is not uncommon and can often link off the flow sideways by enhancing a horizontal flow (Figure 5-4). There can also be enhancement of vertical flows induced by vertically oriented stalks etc. The relationship between horizontal k and vertical k is of interest because it largely determines the flow paths and consequently the paths of substance transport. At present, there are no methods which generally can help to determine in a quantitative way the ratio between horizontal and vertical k .

The flow geometries in wetland soils is hence very much depending on the distribution of hydraulic conductivity. In mires, assumptions of stagnant water in deeper peat layers are often motivated by the presence of dense peat layers and an almost impermeable layer at the bottom of the peat. This is plausible, because this layer often consists of clay with elements of gyttja or other fine detritus in overgrown lakes and shorelines, whereas in paludified areas (chapter 3.2.1), the bottom layer often is having the character of dense, well decomposed fen peat, mixed into the top mineral soil layer. The hydraulic conductivity of clayey soils normally range from 10^{-11} to 10^{-8} m s^{-1} . Therefore has the common view been that the groundwater flow in mires is dominated by lateral flows in the surface layers, whereas water in deeper layers has been considered inert. A special terminology based

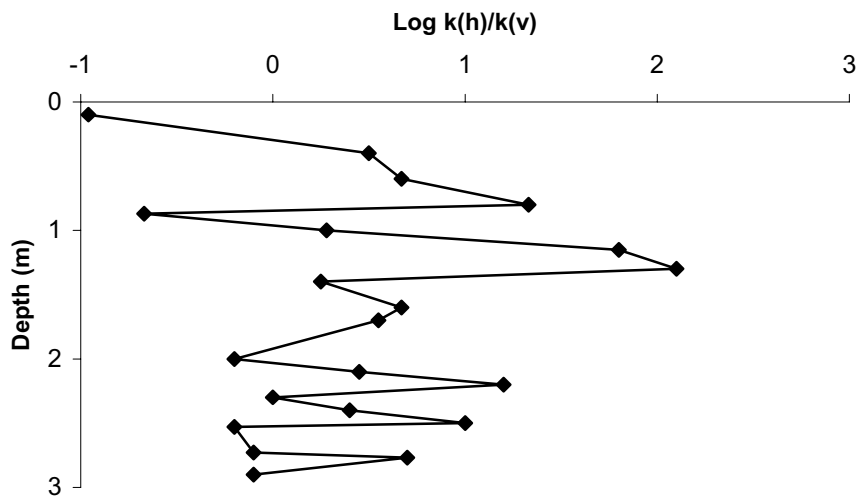


Figure 5-3. 10-Logarithm of the ratio between horizontal (K_h) and vertical (K_v) hydraulic conductivity in a mire. From /Chason and Siegel, 1986/.

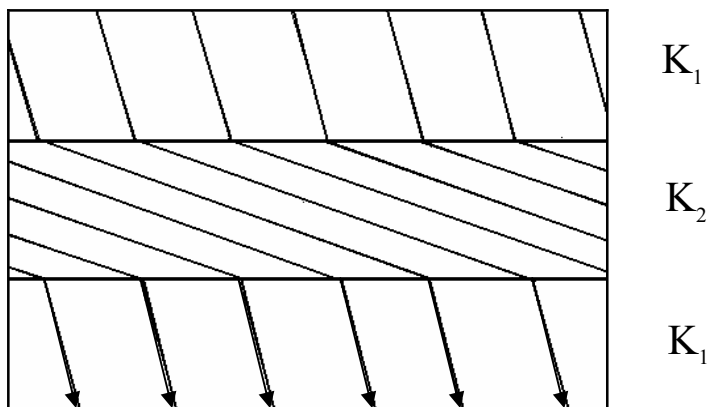


Figure 5-4. Flowlines, expressing the paths of water particles moving through a layered media, with hydraulic conductivities K_1 and K_2 , under a gradient directed downward and to the right. The ratio K_2/K_1 is 7–8. After /Freeze and Cherry, 1979/.

on this view, and in correspondence with what is considered to be biologically active and passive layers respectively, is used widely /see e.g. Ingram, 1983/. The upper zone – the acrotelm, extending to the lowest point of water table fluctuation, is considered to be highly active both hydrologically and biologically whereas the lower zone – the catotelm – holds only low biological activity and very small water flows. That division is found to be helpful when discussing the intensity and localisation of processes and is used by many authors. However, studies of hydraulic heads have shown that considerable vertical groundwater flows often take place in the lower zone (the catotelm), earlier considered to be inactive /Siegel and Glaser, 1987; Devito et al, 1997/. /Reeve et al, 2000/ suggested that two conceptual models would be applicable to describe water flow in different hydrogeologic settings: Lateral flow is dominant in places where the peat overlies sediment of low permeability. The lateral flow is not primarily due to low permeability in the peat itself, but is due to low hydraulic conductivity in the underlying mineral soil. In geological settings where the peat overlies material with a higher permeability, vertical flow will develop under bog mounds as the water seeks the easiest path from high to low hydraulic head, leading it down to the permeable soil and later discharge upwards where the bog dome grades into fen (Figure 5-5). The size of the vertical flow compared to the

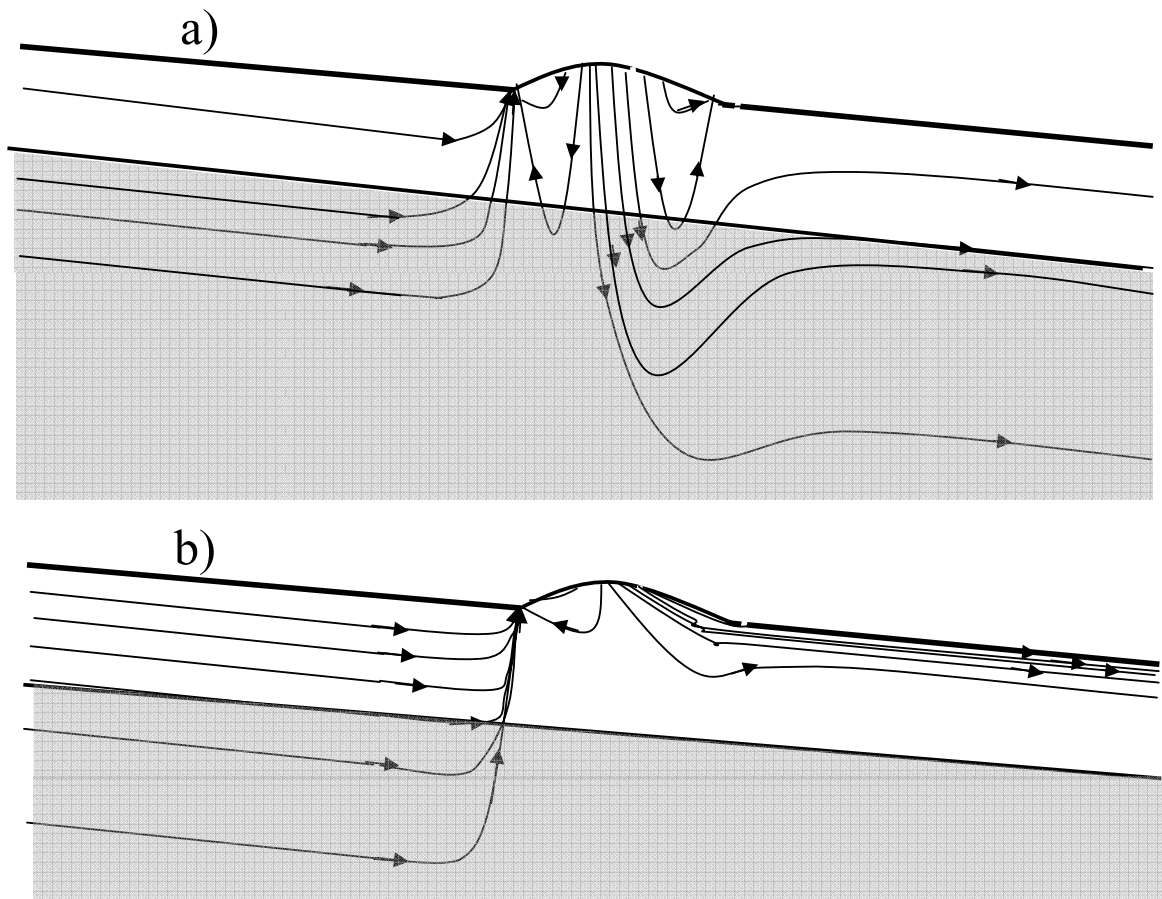


Figure 5-5. Flowlines expressing the paths of water particles moving through peat and underlying mineral soil (shaded area), with a general slope to the right and a (bog) peat mound inducing vertical flows. The mineral soil in a) has a hydraulic conductivity about 1000 times larger than the mineral soil in b), otherwise similar conditions prevail. After /Reeve et al, 2000/.

horizontal is then also depending on the relation between vertical and horizontal hydraulic conductivity and the corresponding components of hydraulic head gradient. If the border of the peat to mineral soil has a low permeability, it is not uncommon that there are breaches through these tight layers creating ‘windows’ where vertical groundwater flow is enhanced. The driving forces behind vertical flows are differences in hydraulic head. The gradients induced by these differences often vary with time and can even shift direction. This occurs in poor mires where a predominant downward flow can occasionally cease and change into an upward flow when water table level lowers at surface /e.g. Siegel et al, 1995/.

5.6 Specific effects of hydrology in respect of systems ecology

The hydrologic conditions affect in many ways on a wetland’s structure and function. The water budget of the wetland sets the variation of water table and wetness in the upper layers of the soil and thus the cycle of aerobic/anaerobic conditions which is crucial for the composition of vegetation. In addition the primary production and other ecosystem functions are often enhanced by flowing conditions and a pulsing water table variation whereas stagnant conditions often have a depressing effect on the biological activity

/Mitsch and Gosselink, 2000/. The water table variation further sets the conditions for the soil chemistry by the effects of the redox potential and other changes with saturated conditions. The nutrient cycling is thereby significantly influenced by the water table variation. The nutrient budget is also dependent on the hydrological dynamics since inflowing water often gives a significant supply of dissolved nutrients and particulate matter. The exports of nutrients and organic matter are likewise dependent on water flows.

Usually, large water flows are judged to give a larger supply of nutrients and stimulate the decomposition and circulation of nutrients by a better oxygen supply to the sediments. However, in submerged wetlands an 'openness' to increased water flow can also be serving as stress for the plants, by increased mechanical force and lower water temperatures. It might also increase the erosion and export nutrients and organic substance from the wetland. A moderate openness is thus the most favourable for productivity. A fluctuating water level often result in a rich ecosystem, giving place for facultative organisms and a pulsing nutrient supply, but also periods of oxygenated substrate, permitting a more efficient decomposition and occasional export of organic matter. Hence the export of organic matter from riparian wetlands can be substantial, not only by export of macrodetritus such as whole plant parts but also by transporting dissolved organic matter and nutrients from upper soil layers.

Further is the location of the wetland within the landscape important for its biogeochemistry and development. Apart from the bog type, which is solely supplied by precipitation water, the hydrochemistry depends largely on the inflow by either groundwater or surface water, which properties depend largely on the sources.

The effect of the water table regime on species composition and diversity in wetlands is widely acknowledged; there are often tight connections between amplitude and frequency of water table level variations, as well as average level, and species composition /e.g. Ivanov, 1981/. However, the development of ecosystems should not be predicted from only hydrological data, since species distribution may be influenced by other variables than water regime and species response to water conditions may be modified by the proximity of other species or by other environmental conditions /Wheeler and Shaw, 1995/. The supply of nutrients and oxygen to the roots also influence on the species composition and plant growth.

6 Geochemistry

The common characteristic for wetland soils is that they are hydric, i.e. they are formed under conditions of saturation, flooding or ponding long enough during the growing season to develop anaerobic conditions. For peat soils the organic material naturally dominates the soil conditions, but the organic content in non-peat soils can also be significant and so abundant that the soils are classified as organic /Mitsch and Gosselink, 2000/.

6.1 Peat chemistry

6.1.1 Peat, constitution and decomposition

Peat decomposition takes place mainly by aerobic and anaerobic microbial metabolism in the acrotelm and in the catotelm respectively. Representing peat with the simplified formula CH_2O , the reaction for aerobic decay can be written: $\text{CH}_2\text{O} + \text{O}_2 \rightarrow \text{CO}_2 + \text{H}_2\text{O}$ whereas the anaerobic decay process resulting in methane production can be written as $2\text{CH}_2\text{O} \rightarrow \text{CH}_4 + \text{CO}_2$ /Lucisano and Bozkurt, 1998/. There are probably other processes of organic matter oxidation in the border zones just below the aerated layers, but as nitrate, metal oxides and sulphur concentrations are generally low in peat their importance are probably low. The most important agents of decay present in peat are fungi and bacteria. Fungal mycelia are abundant in the oxygenated zone but become much rarer in the anaerobic zone, showing the same behaviour as aerobic bacteria.

The chemical composition can be understood in the terms of the processes of death and decomposition of bog plants. The living plants contain mainly proteins, carbohydrates, lipids, and polyphenols such as lignin, along with smaller of other organic and inorganic compounds. Cellulose, which constitutes a large fraction of the plants carbohydrate content, forms the cell walls and gives the structure of the plants. Other carbohydrates, such as simple sugars, are water soluble and more transient, consumed by the plant in making cellulose or oxidised to provide energy alternatively stored as starch to be converted to sugars when needed. When a plant dies its sugars are rapidly released into the surroundings and quickly consumed by the microorganisms. Starches are also quickly digested. Hemicellulose, pectins and gums are attacked more slowly and tend to survive along with cellulose to form the peat. Nitrogen in plant proteins are lost, largely as ammonium salts going into solution, but some of the amino acids or their derivatives appears in the formation of peat humic compounds. The humic compounds are composed by humic and fulvic acids, the latter are transparent whereas the former are brownish and are often constituted by very large molecules. The humic compounds make up the group which gives greatest influence to the properties of the peat. They have a very large specific area giving a large capacity of ion exchange and binding, a great colloidal affinity and a remarkable water absorbing capacity causing a humified peat to become jelly like /Fredriksson, 2002/. The heterogeneous group of lipids within plants encompasses fats and oils, which during decomposition are largely hydrolised into glycerol, readily consumed in anaerobic conditions, and fatty acids which mainly remain unaltered in peat together with other lipids such as waxes and steroids. These substances, which are extractable with organic solvents, are collectively called bitumens /Fuchsman, 1980/. Woody plants, grasses, reeds and sedges contain varying amounts of lignin, a high molecular weight phenolic material that works as a cementing substance for the cellulosic cell structure. During plant

decomposition, lignin is relatively stable. Mosses and lichens do not possess lignin but do have similar chemical substances. In peat analysis, lignin-like elements are often grouped together with humic acids since these substances are similarly polyphenolic in character (these substances are grouped together under the name klasonlignin by /Bergner et al, 1995/). Unlike lignin, humic acids have a high carboxylic acid content and a significant amount of nitrogen.

A precise identification of all peat constituents is hardly possible. Classification and description of peat constituents are often made upon how they dissolve in different solutions. From a technological point of view, /Fuchsman, 1980/ suggested the following classification:

- bitumens,
- water-soluble and easily hydrolyzed materials (sugars), including ‘reducing substances’ (mainly hemicelluloses, distinguished from cellulose by being more readily dissolved and hydrolysed by diluted acids, cellulose demanding concentrated acid),
- humic acids,
- fulvic acids,
- cellulose,
- lignin,
- ash.

6.1.2 Proportions of organic substances in different peats and variations with decomposition

The proportion of these substances is different among different peat types and changes also with degree of decomposition and peat depth. In general our knowledge of the organic composition of peat, and of the changes which occur, is fragmentary. The composition of plant fragments and their degree of decomposition determine most peat properties /Bohlin, 1993/ though the almost infinite number of possible combinations of different botanical, hydrological and chemical properties make it hard to connect causes and effects on the peat characteristic /Clymo, 1983/. In general the carbohydrate fractions decrease with increased decomposition, whereas the proportion of lignin, humic and fulvic acids increases with decomposition /Bergner et al, 1995; Mattson and Koutler-Andersson, 1954; Clymo, 1983/. The fractions of bitumen and inorganic substances also increase with decomposition. The fractions of bitumen and carbohydrates are generally greater and the fraction of lignin, humic and fulvic acids is smaller in Sphagnum peat than in Carex and woody peat /Fuchsman, 1980; Bergner et al, 1995/ (Figure 6-1). Within the Klasonlignin group, the proportion of humic acids generally increase more with degree of decomposition than lignin, whose proportion may even decrease in fen peat /Naucke et al, 1993/.

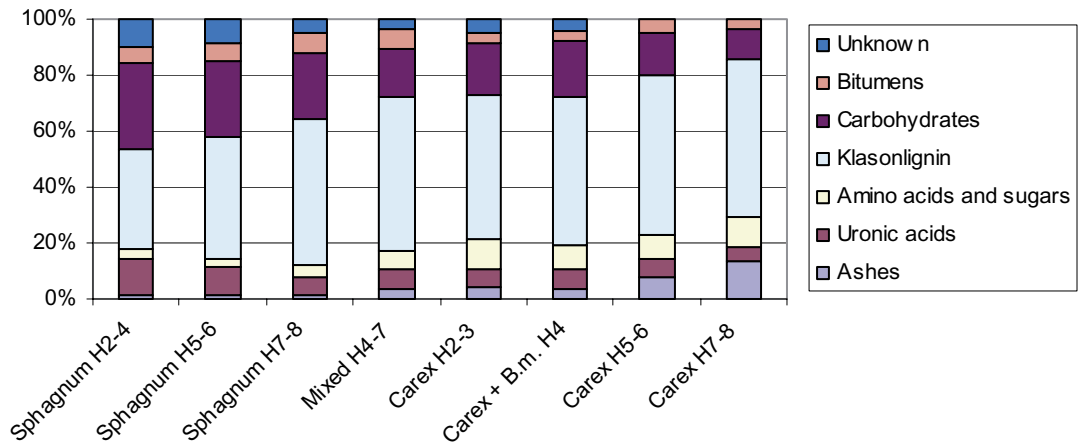


Figure 6-1. Composition of peat with different original plant composition and in different stages of decomposition (von Post number). Uronic acids can be seen as a part of the carbohydrate group, while the amino acids and sugars can be seen as part of both carbohydrate and humic substance. B.m. = brown mosses (*Scorpidium spp.* etc). From /Bergner et al, 1995/.

6.1.3 Sorption properties

Peat materials can very easily remove cations from solution and retain them. The processes governing this high sorption capacity of peats can be attributed to different processes and there is not yet any consensus to which process is prevalent in which occasion. Different studies have reached different conclusions, probably depending on differences in type and preparation of peat, metals studied and the used methods /Brown et al, 2000/. Theories include ion-exchange, surface adsorption and complexation. All these types of sorption depend on pH as the cation sorption capacity increases with an increasing pH (increased amount of negatively charged sites). The most common belief is that ion-exchange is the most prevalent mechanism /Kadlec and Keoleian, 1986/. The humic substances are then considered to determine the binding capacity. The dominating exchange sites of peat compounds are the carboxyl (-COOH) and phenolic hydroxyl (-OH) groups. Metals react with these groups to release protons or, more seldom, to displace an existing metal. The protons and metal ions can then be considered to form a system of reversible interchange, whose equilibrium constant is given by

$$K_{eq} = (\bar{a}_B^{|z_A|})(a_A^{|z_B|}) / (a_B^{|z_A|})(\bar{a}_A^{|z_B|}) \quad (6-1)$$

where \bar{a}_k and a_k refer to the activities of the ions (k = A, B) in the exchanger and solution phases respectively, while Z is the valence number. The pH and the pK_a (K_a is the ionisation constant of the humic acid) are then important factors for the binding capacity which would vary significantly within the pH ranges found in mires. The carboxyl groups start to dissociate already at low pH (pH<3) whereas phenolic hydroxyl groups are dissociated markedly only at pH higher than neutral /Tan, 1993/. Complex and variable composition of the humic and fulvic acids make it hard to predict the reactions of sorption without analysis on the properties. There are also different affinities to ion sorption among different ions. This is probably caused by the molecular structure of the humic compounds and the difference in ability among the ions to fit in these structures. There are then many opportunities for multivalent ions to couple to different groups within the same molecule. Some ions form very insoluble compounds by coupling their double or triple charges to negative groups of one organic molecule, forming ring-form structures, chelates. Others form complexes with several humic compounds as ligands. Depending then on the ligand structure these complexes could precipitate or remain

in solution. The variability of ion valence with redox and pH (causing varying formation of inorganic, ionic complexes) also complicate the description. Although the selectivity of organic matter for various cations does not conform to any absolute rules, some trends have been observed /Kadlec and Keoleian, 1986/. The peat generally prefers ions of higher valence. Stability complexes of bivalent ions were found by /Irving and Williams, 1948/, cited by /Kadlec and Keoleian, 1986/, to follow the order $Pb^{2+} > Cu^{2+} > Ni^{2+} > Co^{2+} > Zn^{2+} > Cd^{2+} > Fe^{2+} > Mn^{2+} > Mg^{2+}$, while /Bunzl et al, 1976/ found that the order of stability for humic complexes was $Pb^{2+} > Cu^{2+} > Cd^{2+} > Zn^{2+} > Ca^{2+}$. The measured stabilities of the complexes depend on the ambient conditions, i.e. the peat chemistry and the pH /Kadlec and Keoleian, 1986/. There are some uncertainties on how different concurrently present cations influence on each other. Presence of monovalent cations such as Na^+ has been found to reduce the sorption of metals /Brown et al, 2000/ as well as a high concentration of Ca^{2+} in solution. On the other hand, peat with exchangeable Ca^{2+} (and Mg^{2+} and Fe^{2+}) ions are found to enhance the sorption of heavy metals, since Ca^{2+} ions are less tightly bound to peat than H^+ ions /Kadlec and Keoleian, 1986/. The sorption of metals on peat has gained increased interest during latest years and several studies have been made on the sorption capacity of peat, mainly for treating wastewater /e.g. Brown et al, 2000; Ringqvist et al, 2002/, but also on substance mobility and sorption in natural systems, some of which are treating potentially radioactive substances.

The cation exchange capacity (CEC) of a medium is a measure of its sorption ability and is often used for mineral soils /Wiklander, 1976/. In mineral-soil wetlands, the CEC is relatively independent of pH. The mechanisms for both cation exchange and anion adsorption are quite different from the peat environment /Ross, 1995/. The CEC of mineral soils is to a large degree dependent on the content and composition of clay minerals. The clay particles present a large specific surface but the major cause to their large CEC is the large amount of unsatisfied clay-surface negative charges to which aqueous cations in solutions are attracted /e.g. Tan, 1993; Brady, 1990/. These are generated through morphologic changes in constituent clay minerals and are independent of pH. The CEC of soil clay minerals and oxides vary between 2 and 100 mEq $100g^{-1}$ with the higher values for highly swelling minerals as smectites and vermiculite and lower for sesquioxides (for details, see e.g. /Brady, 1990/). In using this term for organic soils, care must be taken since the CEC covers all different processes of sorption and the value depends on extraction methods and varies both with different adsorbed substances and with pH. /Clymo, 1983/ discusses this subject more. The CEC of mires is found to be of the order of 150–250 mEquivalents (mEq) per 100 g dry soil /Naucke et al, 1993/. The organic peat polymers of the peat with their abundant carboxyl and hydroxyl groups are responsible for this high value. The dissociation of these groups is pH dependent and so is consequently the CEC. Apart from the variation with pH, the structure differ between different peat types and the CEC is consequently found to differ, e.g. *Sphagnum* peat having a higher CEC than *Eriophorum* grass peat and “moss peat” with decomposition H4–H5 having CEC of 50 mEq dm^{-3} whereas “fen peat” had 400 mEq dm^{-3} /Naucke et al, 1993/. /Ross, 1995/ summarised some studies on wetland soils and found CEC was 3–6 times higher in acid peats than in mineral gley soils.

6.1.4 Dissolved organic matter

An important part of the carbon budget for peatlands is the production and transport of the dissolved organic carbon (DOC) fraction. The DOC is determined by analysis of the dissolved organic matter (DOM) content. This fraction is composed by the same type of components which make up the solid fraction of the peat. Hence is the DOM mainly composed by fulvic acids, long-chain fatty acids and esters. The proportions of these are connected to the properties of the parent material. These elements are hard to distinguish from the “solid” peat in highly decomposed peats. In fact is the definition of DOM that it

has a particle size less than $0.45 \mu\text{m}$ /Droppo, 2000/ by which it differs from particulate organic matter. The metal binding by the dissolved fractions is consequently hard to distinguish from the non-dissolved matter. Hence is the net effect by dissolved organic matter on metals included within the total peat matter, with the exception that the dissolved fraction is more mobile. There are a few publications on measurements of concentration profiles and fluxes of DOC. The chemical composition of DOC found in discharge water from wetlands were in many ways depending on the water pathways. /Schiff et al, 1998/ found that most DOC in water originating from wetlands was produced by fresh organic matter. They explained the large fraction of recent carbon by the shallow flow paths through the highly active surface layers in high-flow situations. /Scott et al, 1998/ found a considerable seasonal variation of DOC concentration in mire pools at a North-British mire, with an minimum concentration in winter time and a peak concentration reached in late summer, often in connection with an increase in water flow after a dry summer period. It was concluded that the DOC variation was not just a phenomena of dilution but reflected a seasonal variation of microbial activity. The fraction of hydrophilic DOC was also larger in the summer and it was concluded that this fraction was more easily washed out of the peat matrix whereas the hydrophobic fraction was stronger adsorbed to the peat and leached more slowly through the fall and winter. The upper layers were then intrinsically assumed to be the main source of the pool DOC, although they did not try to determine the origin and age of the DOC. /Fraser et al, 2001/ found that the quality as well as the total concentration of the DOC changed significantly with depth in a bog. They found that the main DOC flowing out from the mire through the borders was coming with water flowing in the top layers and was hence of recent origin. During most time, the vertical component of the groundwater flow was directed downward and the decreased DOC content with depth was attributed to in situ microbial consumption of the more readily processed substances. The dissolved inorganic content (mainly CH_4) increased correspondingly with depth. It can thus be concluded that the production of DOM in deeper layers is small. The ability of different types of DOM to bind metals by soluble complexes is however unpredictable with current knowledge, although we can expect a certain transport effect by complexes between metals and dissolved organic matter. /Gorham et al, 1998/ examined the principal components of DOC in stream water from wetlands and found that dissolved Fe, and (although to a much less extent) dissolved Al, showed a strong positive correlation with DOC.

6.2 Inorganic chemistry

6.2.1 Dominating processes

The water chemistry is not only a result of the local geology, the origin (sea, mineral, or organic soils) and the flow paths of the laterally inflowing water and its mixture with precipitation. It is also highly dependent on the biological activity and the often varying conditions of pH and redox potential. Although the mineral-soil wetlands and lake bottoms lack peat layers the soil chemistry is most often dominated by the presence of organic matter /Mitsch and Gosselink, 2000/. Since the diffusion of oxygen from the surface down into the soil is very slow, it is often readily consumed by the bacterial demineralisation of the organic matter. The soil bacteria then proceeds through a sequence of reactions in which the available oxidant that yields the greatest free energy determines the dominant process at a particular level. At the same time, these processes will greatly influence on the redox potential of the system. If O_2 is present aerobic decomposition is the major pathway. When O_2 is depleted, organic matter degradation shifts to nitrate (NO_3^-) reduction (denitrification). Once nitrate is fully utilised, metal oxides (e.g. MnO_2 and Fe_2O_3) may serve as oxidants. When metal oxides are no longer available, the next step would be sulphate reduction. The final step in the series is when the organic matter itself or CO_2

become the electron acceptor (oxidant), whereby methane is produced as a final product. As the decomposition processes proceed through this series, the redox potential (expressing the electron availability) falls as the electron donors successively turn into their reduced forms. The dominating oxidants and the corresponding redox potentials for their transformations are given in Table 6-1. These redox potentials are not precise thresholds as temperature and especially pH are important factors. The pH dependence of redox potentials, where chemicals are stable in either oxidised or reduced states, can be shown on redox-pH stability diagrams (see, for example, /Brookings, 1988/).

The pH of wetland soils and water vary widely between different wetland types. Organic soils are often acidic, with exception of rich or extremely rich fens, whereas mineral soil wetlands often are close to neutral. Some of the dominant elements which get reduced while they take part in the oxidation of organic matter also affect the pH depending on their function as acids or bases in the transformation between oxidised or reduced forms. The nitrate reduction influence is probably small in all peatlands, where only small amounts of dissolved N occur, whereas it can be significant in eutrophic environments, e.g. within farmland catchments and along coastal marshes /Mitsch and Gosselink, 1993/. The most common result of organic N mineralization is that it transforms into ammonium, which is readily assimilated by organisms again or retained by the peat matrix; it could be mobilised by transforming into nitrate, nitrification, a process which however demands oxygen and near neutral pH. Both iron and manganese occur in two oxidation states (Table 6-1). In wetlands they are primarily found in their reduced forms, which also are more soluble and mobile than the oxidised forms /Mitsch and Gosselink, 1993/. In conditions with abundant sulphide, FeS can be formed and in neutral- basic conditions ferric carbonate FeCO_3 can develop. Influences from transformations between sulphate and sulphide on the redox-pH system can be expected primarily within minerotrophic wetlands and especially in marine environments /Mitsch and Gosselink, 1993/. In such sulphur-rich areas, periodical variation of water level may result in relatively high concentrations of sulphuric acid which could rend pH to fall below 2.5 /Bloomfield, 1972/. In peatlands, especially in bogs and poor fens, the small amount of inorganic substances in comparison with the amount of carbon compounds makes the CO_2 - CH_4 transformations to dominate the redox potentials in the layers that are situated below the water table, which however not excludes the CH_4 transformations in mineral-soil wetlands and, in fact, the high productive mineral-soil wetlands are the greatest CH_4 producers /Mitsch and Gosselink, 1993/. In addition to the effects on the redox potential, the organic compounds influence the pH (as decomposition generates organic acids) and, greatly, the ion exchange capacity. There is hence most often a vertical stratification within the wetland soil with a decreasing redox potential with depth, ranging from often aerated conditions at the surface to strongly reduced conditions at some point deeper in the profile. However the occurrence of larger pores (from channels of roots or worms and soil structure with inter-aggregate cracks) which are readily aerated could cause a horizontal variation of redox conditions as well. This can be manifested in mineral-soil wetlands where gley forms with ferric hydroxide and manganic oxide mottles in coarser pores in the otherwise greyish ferrous (Fe^{2+}) environment /Brady, 1990/. Further, plants in wetlands are often equipped with aerenchyma, tissues incorporating large, gas-filled spaces interspersed with the cells in a characteristic pattern, which allow them to transport oxygen down to the roots. Some of the oxygen diffuses outward from the roots, aerating the adjacent soil. /Engler and Patrick, 1975/ demonstrated that sulphur bound in reduced soil metal sulphides could be oxidised by penetrating rice roots and taken up. This is also of potentially great importance for processes of nitrogen transformations as ammonium can be oxidised to the more mobile nitrate which then can be transported to other areas or be reduced by denitrification in the surrounding reduced environment /Ross, 1995/. /Ross, 1995/ also found the nitrogen transformation processes to be particularly important in peats with seasonally saturated surface horizons since the peat may host considerable amounts of organic nitrogen.

Table 6-1. Oxidised and reduced forms of several elements and approximate redox potentials for their transformation /after Ross, 1995/.

Element	Oxidised form		Reduced form	Approximate redox potential for transformation (mV)	
				pH=7	pH=5
Oxygen	$O_2 + 4H^+$	\leftrightarrow	$2 H_2O$	820	930
Nitrogen	$NO_3^- + 2H^+$ (Nitrate)	\leftrightarrow	$NO_2^- + H_2O$	420	530
Manganese	$Mn^{4+} + 4H^+$ (Manganic)	\leftrightarrow	$Mn_{2+} + 2 H_2O$ (Manganous)	410	640
Iron	$Fe^{3+} + 2H^+$ (Ferric)	\leftrightarrow	$Fe_{2+} + 3 H_2O$ (Ferrous)	-180	170
Sulphur	$SO_4^{2-} + 10 H^+$ (Sulfate)	\leftrightarrow	$H_2S + 4 H_2O$ (Sulphide)	-220	-70
Carbon	$CO_2 + 8H^+$ (Carbon dioxide)	\leftrightarrow	$CH_4 + 2 H_2O$ (Methane)	-240	-120

7 Transport processes

7.1.1 Advection, Diffusion, Dispersion

The distribution of concentration and transport of solutes are dependent on the processes of advection, dispersion and diffusion. These are briefly described here. Descriptions of these processes and how they can be treated mathematically and estimated in a quantitative manner can be found in e.g. /Freeze and Cherry, 1979; Hillel, 1998; Hughes, 1999/.

Advection implies the mass flow of solute following the flow paths of water at the average velocity of the local water flow. This flux f_{xa} [$M L^{-2}T^{-1}$] can thus be described as

$$f_{xa} = Qc/A \quad (7-1)$$

where Q is the water flow from eq. (5-2) and c is the concentration [$M L^{-3}$]. The simplest estimation of particle velocity v through a medium can then be described by

$$v = Q/(A\theta_{eff}) \quad (7-2)$$

where A is cross-section area and θ_{eff} is effective porosity i.e. the volume of the pores that are active in conducting the transported solute. Complications of this simple estimation are caused by dispersion, i.e. heterogeneities in the medium create variations in flow velocities and flow paths such that the solute is transported slower in some parts and quicker than average velocity in other parts. The dispersion hence mixes the water and evens out the concentration differences along the transport. The dispersion also occurs normal to the direction of flow by diverging flow paths that cause the solute to spread laterally from the main flow direction.

Diffusion is a molecular-scale process, which causes spreading due to concentration gradients and random motion and can occur in the absence of velocity. A solute moves from an area of higher concentration to an area of lower concentration. Diffusion is a factor to account for in the case of very low velocities or in the case of mass transport involving very long time periods. Mass transport in one dimension can be described by Fick's law of diffusion.

$$f_{xd} = -D_d \frac{\partial c}{\partial x} \quad (7-3)$$

where f_{xd} = mass flux by diffusion
 D_d = diffusion coefficient [$L^2 T^{-1}$]
 $\delta c/\delta x$ = concentration gradient

Dispersion can be formulated mathematically in the same way as diffusion in eq. (7-3) except that a dispersion coefficient D_h is used instead of D_d . They can be lumped together in one coefficient $D_{dh} = D_d + D_h$, although one has to remember the difference in their physical meaning. The diffusion coefficient D_d depends then on the porosity and pore structure whereas the dispersion coefficient D_h also depends on the (advective) flow velocity v . It is defined as $D_h = \alpha v$, where α is dispersivity of the medium [m]. /Hoag and Price, 1995/ found values of α in a blanket bog peat to be between 1.8×10^{-3} m and 4.9×10^{-3} m at depths between 0.20 m to 0.60 m. /Ours et al, 1997/ found an accelerated dispersivity for NaCl solution in peat in laboratory. During continued experimentation, chloride dispersion

increased, making it impossible to determine precise values of dispersivity although they were found to be in the order of cm up to tens of cm. At the same time they found a slow total chloride transport, indicating that the peat functions as a dual porosity medium, where solutes move preferentially through active macropore spaces and diffuse into passive pore space in the matrix.

The mentioned three transport processes can be added together in one expression describing the balance of the solute within the soil, i.e the net storage within a finite volume is equal to the net sum of fluxes plus possible sinks or sources. Using this principle of conservation of mass, a one-dimensional description is /Hillel, 1998/:

$$\frac{\partial(\theta c + \sigma_s)}{\partial t} = \frac{\partial}{\partial x} \left(D_{dh} \frac{\partial c}{\partial x} \right) - \frac{\partial(v\theta c)}{\partial x} + S \quad (7-4)$$

where t is time, S is source (+) by e.g. production or sink (-) by e.g. consumption and σ_s is temporary storage of substance in soil matrix but outside the liquid phase, such as in precipitated form or in the soil's exchange complex. The latter process is often described by the term sorption. Typical values of D_d are of the order of magnitude $10^{-9} \text{ m}^2 \text{ s}^{-1}$ /Reeve et al, 2001/. /Giles, 1977/, cited by /Clymo, 1983/, found the diffusion coefficient of phosphate in peat to be $10^{-10} \text{ m}^2 \text{ s}^{-1}$.

7.1.2 Sorption

The term sorption encompasses the processes of adsorption and absorption (association of the solute on the surface of a particle and within the particle, respectively). In reality, the processes behind sorption are of different physical and chemical phenomena, including ordinary ionic exchange within the diffusive double layer, van-der-Waal forces, forming and precipitation of ionic compounds, complex building and forming of hardly soluble chelates. Often it is difficult to distinguish between these processes and to quantify their effects, since they may occur simultaneously. Hence the term sorption may be used to describe the overall phenomena. The most common approach for evaluating a solutes distribution between the solid and aqueous phase is to assume a linear relation through a distribution coefficient (or partition coefficient /Hughes, 1999/) k_d :

$$k_d = \frac{[A]_{solid}}{[A]_{aqueous}} \quad (7-5)$$

The use of the distribution coefficient is a convenient way of describing the behaviour of the peat and the interchange between the peat and the solution. However the complex composition of peat and the different connections between pH, Eh, peat type and abundance of different inorganic substances and the various processes affecting the sorption make the use of k_d uncertain. There are numerous possibilities of combinations of these processes, which will affect k_d in different ways and estimations of sorption of substances in peat should incorporate some estimation of uncertainty. More realistic would be to estimate the effects through a range of possible or probable situations.

7.1.3 Relocation of minerals by biological uptake, secretion, circulation, enrichment

The relocation of inorganic constituents by biological processes are in principle the absorption of plant roots and relocation within plants up into the leaves, rhizomes and possibly also downwards by transport in deeper roots. In peat-covered mires, particularly the content of K and P in vascular plants is very large compared to the surrounding peat

/Malmer, 1958/. The role of mosses is unclear, since they lack roots and internal transport systems and are hence dependent on external supply, but there are evidences that relocation between different parts of Sphagnum plants may occur /Clymo, 1983/. Especially, there is an enrichment of essential elements such as N, P and K within the upper parts of the moss plants /Damman, 1978; Malmer and Sjörs, 1955/. The process behind is unclear. According to /Shotyk, 1988/, the surface mineral enrichment depends on deposition and less to bioaccumulation. On the other hand is the enrichment of Na much less than of K, despite their similar deposition rates (inland conditions) should give a similar supply. Probably, the biological need for K causes a quick incorporation of the K^+ ions within the cells as soon as they are supplied from external sources or released from other adjacent biological processes.

An important feature of Sphagnum mosses is their production of organic acids and hence their great cation exchange capacity /Clymo and Hayward, 1982/. Cations in supplied water are exchanged for H^+ ions on the cell walls of mosses. This is also considered to be a major cause to the low pH of Sphagnum dominated mire surfaces /Clymo and Hayward, 1982; Shotyk, 1988/. However, /Shotyk, 1988/ pointed out that the main reason for the bogs to be acidic is that the supply of mineral bases is too small to catch the H^+ production. Though the plant growth continuously produces organic acids, there seems to be an accumulation of inorganic substances in the top layer. Especially is the enrichment of K and P conspicuous. Results and discussions on element contents in mire vegetation and peat are thoroughly presented by /e.g. Malmer and Sjörs, 1955; Malmer, 1958, 1962a,b/.

8 Resulting contents and transports in peat

The content of minerals in peat and peat water reflect in many ways the result of water origin, water movements and diffusion. Organic matter chemical or biochemical processes may influence on the inorganic chemistry as well. Peat chemistry is reviewed by e.g. /Clymo, 1983; Shotyk, 1988; Naucke et al, 1993/. Shortly, peat water pH is the result from the acid production rendered by biological activity and the supply of bases from inflowing waters. In bog surface, the pH is normally approximately 4. In fens the pH normally range between 4 and 8.5. /Shotyk, 1988/ suggested that the range is connected to buffering carboxyl groups of organic matter (at pH = 4) and calcium carbonate in equilibrium with atmospheric concentration of CO₂ (at pH = 8.4), respectively. The approximate range of pH in different mire types is given in (Table 8-1). Observe that this is approximate since although the pH, amount of base cations and vegetation type is well correlated, the pH and soil chemistry can vary much within the same type of mire /Sjörs, 1956; Gorham, 1955/. It is not unusual that the pH of bogs may increase with depth as deeper layers are more influenced by groundwater coming from mineral-soil /Shotyk, 1988/ and it is probable that also fen pH can vary both horizontally and vertically depending on the influence from inflowing water. Steep pH gradients have been found in peat where Sphagnum hummocks are established above otherwise minerotrophic fens /Kivinen, 1933; Bellamy and Riley, 1967/, consistent with observations that bog peat may accumulate concurrently with rich fen peat /McNamara et al, 1992/.

The redox potential of mire waters have been measured with inert metallic electrodes in many studies /e.g. Malmer, 1962a/ but /Shotyk, 1988/, referring to /Morris and Stumm, 1967/, considered that these could not be used for quantitative determination and are hence useless for comparison with redox species of interest. /Clymo, 1983/ deals largely this view but still finds it interesting to discuss redox variations qualitatively. Apart from areas with open water and possibly within dense aerating root zones, dissolved O₂ is absent in peat waters. Abundant methane production in all layers below the zone of water table indicates genuine low redox potentials. However there are some indications that the lowest layers could be somewhat more oxidised /Shotyk, 1988/, possibly by influence from inflowing mineral-soil groundwater.

In comparison with mineral soils, the waters of mires have a much lower salt content. The content of inorganic substances in peat water is largely determined by the same factors that determine pH. The mixing ratio between precipitation water and inflowing minerotrophic water and its origin is hence decisive for the peat mineral content. Most measurements of peat water chemistry have traditionally been made in connection with mire ecology studies

Table 8-1. The range of pH in different mire types /from Sjörs, 1950/.

Type of mire	pH range
Bog	3.7–4.2
Extreme poor fen	3.8–5.0
Transitional poor fen	4.8–5.7
Intermediate fen	5.2–6.4
Transitional rich fen	5.8–7
Extreme rich fen	7–8.4

and consequently they are concentrated at the surface layers /e.g. Witting, 1947, 1948; Sjörs, 1948, 1950; Malmer, 1962a, 1963/. The electrical conductivity corrected for pH increases from sometimes less than $10 \mu\text{S cm}^{-1}$ in bogs and poor fens to about 100 for rich fens and about 300 for extreme rich fens /Sjörs, 1950/. However, in comparison of different mires the total salt content is not, unlike pH, a good indicator of mineral richness or composition of vegetation /Sjörs, 1950/. The composition of salt content in mire waters also varies with local factors such as vicinity to sea, impact of inblowing dust particles etc /Witting, 1948; Gorham et al, 1985/. /Witting, 1948/ found a large difference between poor fens along a transect from the coastline of western Sweden (400 and $1100 \mu\text{S cm}^{-1}$) to inland (11 and $54 \mu\text{S cm}^{-1}$). As with pH, the mineral content can also vary within each mire, both in space and time (Table 8-2) /e.g. Witting, 1948; Malmer, 1962a/.

The content and distribution of minerals in peat often follow the concentration in the water, especially for the total ash content, since the major elements Na, Ca, Mg and K are very soluble in relation to other cations and contents of Si and Al in large reflect the amount of supply of these elements from the mineral soil /Shotyk, 1988/. Iron (Fe) and manganese (Mn) have some more complicated distribution depending on pH and redox patterns in the past, found to be soluble in acidic anaerobic peats /Shotyk, 1988/. /Shotyk, 1988/ has a detailed discussion on the reactions between peat Fe^{3+} , Fe^{2+} ions and other ions, how they precipitate and transform and how the complexes may interact. The solubility for Al is highly dependent on pH as well. Aluminium has also shown to very easily form complexes with organic matter. In waters with low pH and high content of DOC, as much as 80–90% of dissolved Al could be complexed with the organic matter and transported away /Helmer et al, 1990/. Some trace metals as Ni, V and Cr are found to form insoluble compounds and can be expected to behave conservatively in peatlands /Shotyk, 1988/. Others, like Cu, Pb and Zn have shown to form stable complexes with humic compounds, often leading to enrichment of these metals. Total mineral contents for bog and fen peats are given in Table 8-3.

There are also studies made over content and movements of radioactive substances. Studies with the objective to examine the behaviour of radioactive isotopes for assessing the potential transports from leaking nuclear fuel waste disposals are also made /Reid et al, 1994/.

There are several studies indicating that uranium (U) is easily adsorbed by peat /Fredriksson et al, 1984; Halbach et al, 1980; Idiz et al, 1986; Sheppard and Thibault, 1988; Zielinski and Meier, 1988/. /Idiz et al, 1986/ found that U was enriched greatly compared with other trace metals in a spring-fed peatland. The high content of U was explained by complexation by humic acids' carboxylate functional groups at a pH which maximize these groups sorption of uranyl (UO_2^{2+}) ions. These studies of metal contents in natural wetlands are probably

Table 8-2. pH and concentration [mg l^{-1}] of dissolved cations in surface water from different parts of the mire Ryggmossen, Uppland. The range within each item arises from seasonal variations. From /Witting, 1948/.

Entity	Bog hollow (<i>Eriophorum</i> v.)	Bog hollow (<i>Scheuchzeria</i> p.)	Poor fen hollow	Rich fen hollow
pH	3.8	3.4–3.9	4.2	5.8–6.2
Na	0.26–0.65	0.78–1.57	1.01–2.75	2.93–3.40
K	0.12–0.26	0.16–0.32	0.13	0.16–0.52
Mg	0.23	0.17–0.40	0.96–1.44	0.45–1.44
Ca	0.22–0.4	0.18–0.36	0.79–1.30	4.9–6.9
Total cations	1.0–1.36	1.29–2.67	2.9–5.6	10.4–11–7

Table 8-3. Summarised data on 14 bog and 25 fen peat soils in Germany /after Naucke et al, 1993/. The numbers are on weight basis.

	Ash (%)	pH	Al (%)	Fe (%)	Mn (ppm)	Ni (ppm)	Co (ppm)	Cr (ppm)	Zn (ppm)	Pb (ppm)	Cu (ppm)	Sr (ppm)	Ca (%)	Mg (%)
Bog														
Mean	3.0	4	0.15	0.24	28	15	8	16	23	23	8	8	0.3	0.16
Max	5.1	4.5	0.36	0.47	61	40	13	28	50	60	20	14	0.6	0.27
Min	1.0	3.5	0.02	0.03	4	2	3	1	3	2	2	4	0.2	0.08
Fen														
Mean	35	6	1.0	1.7	280	43	25	80	66	42	18	40	3.0	0.39
Max	56	7	1.7	2.6	420	66	40	130	106	83	28	55	4.0	0.68
Min	22	5	0.3	0.1	140	20	10	30	26	2	8	25	2.0	0.10

needed to get the picture of the long-time effects of influx of substances into the peat, sorption of substance to plant tissues, the subsequent degradation of the organic material and formation of new organic products and the related destruction and formation of complexes /Idiz et al, 1986/. /Halbach et al, 1980/ found that almost all U in peat was associated with humic and fulvic acids by complex bindings with the uranyl ion but also that a large part of the uranyl was associated with dissolved humic and fulvic acids (i.e. DOM) and transported away without precipitating into the peat or any of the following lake sediments. /Zielinski and Meier, 1988/ found that U could be remobilized from peat during certain conditions following either human impact or natural changes. Effective leaching of peat uranium was reached when it was subjected to solutions with anions that could form complex with the uranyl ions. These solutions included both H_2SO_4 and $NaHCO_3$ and Na_2CO_3 , meaning that dissolving of U could be reached in a wide range of pH, and suggesting that events of peat liming and periods of sulphate enriched acid rain could set on leaching. /Fredriksson et al, 1984/ concluded that of the ^{238}U and ^{232}Th radioisotope series, only the uranium was adsorbed by the peat in large amounts compared with the other substances. This depended partly on the higher mobility of the uranyl ions in comparison with the other cations which made the U to reach the peat layers whereas the other ions were caught earlier by adsorption in the mineral soil layers. They suggested also that the uranium was reduced to U^{4+} , which has much less solubility, in the peat. If the peat then was subjected to more oxidizing conditions, e.g. by drainage, the uranium could easily be mobilized again. This was also concluded by /Kochenov et al, 1965/ as they found reduced peat could retain 2–3 orders of magnitude more uranium than the same peat exposed to air. They also pointed out the importance that k_d values for uranium and peat should be determined under the same redox conditions as they would be applied for in sorption estimations. /Landström and Sundblad, 1986/ found high enrichment of thorium together with uranium in some peat soils, although the Th/U ratio decreased with depth in peat. /Sheppard and Thibault, 1988/ found that U, Np and Tc were quickly immobilized in both bog and fen peat, while I was much more mobile. The k_d values for I ranged between 0.2 and 64 $L\ kg^{-1}$, whereas U had k_d values between 2600 and 19000 $L\ kg^{-1}$, Np 31–2600 $L\ kg^{-1}$ and Tc 2–180 $L\ kg^{-1}$. These wide ranges of k_d also for each substance indicate that the conditions, i.e. peat type, pH and redox must be known for estimation of k_d values for each situation. /Echevarra et al, 2001/ suggested that pH should be the main variable for estimation of the sorption of uranium in soils. Method testing for calculation of radionuclides in soils and sediments are given by e.g. /Puigdomènech and Bergström, 1994/. /Fredriksson et al, 1984/ studied peat samples from 146 Swedish mires. The ash concentration of U was less than 100 ppm in 80% and < 200 ppm in 90% of the samples, but very high contents of natural U were found in some fen peats situated in areas with uranium-rich bedrocks, with a top value of 7500 ppm U in the ash of the bottom peat close to a bedrock fracture zone (Figure 8-1).

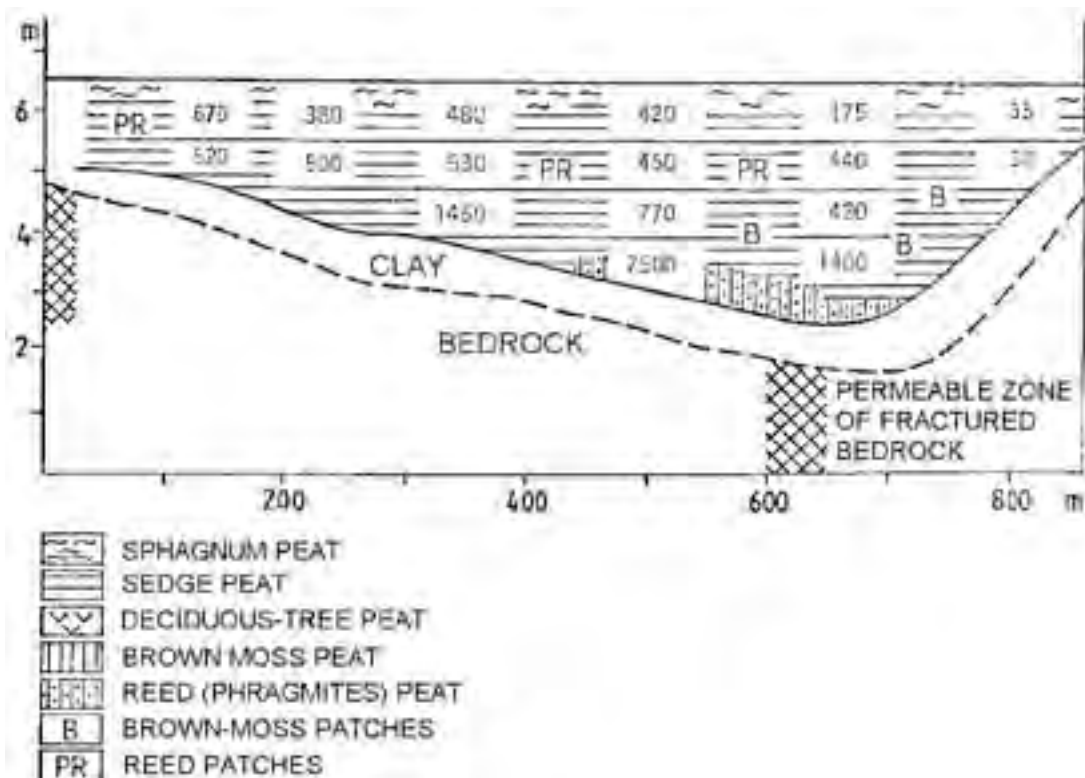


Figure 8-1. Cross section of a fen in a moderately uranium-rich bedrock area. The numbers represent uranium content (ppm U in the ash fraction). From /Fredriksson et al, 1984/.

The distribution of inorganic elements in peat is the result of the complex combinations of amount and form of supply of the elements, pH and redox environment together with the peat chemical constitution and physical properties and the concurrent water flows. Relocations of elements are occurring by the processes of advection, dispersion and diffusion and by biological processes. Recognised plant effects on inorganic constituents are absorption in plant roots with subsequent relocation both upwards into stem and leaves and downwards by root growth. Relocation of elements occur likewise in Sphagnum plants /Clymo, 1983/. In bogs and very poor fens with deep peat layers, the surface layers have a relative high content of minerals compared to the layers underneath. The mineral content increases then again in the deep layers close to the mineral soil causing a “C-shape” form in the mineral distribution through the profile (Figure 8-1, Table 8-4) /Damman, 1978; Clymo, 1983/. The biological relocation processes and the shifts between aerobic and anaerobic conditions are claimed to be responsible for these ‘enrichments’ together with deposition of trace elements from air at the surface /Clymo, 1983; Shotyk, 1988/.

There is still very little known about the transport processes within peatlands. The use of models coupling groundwater flow and chemical transport processes has shown to be a useful tool to understand and quantitatively describe the processes and to estimate parameters for predictive purpose.

/Price and Woo, 1988/ could show with a simple model that the content of minerals at coastal marshes of James Bay, Canada, was determined by supply from deeper sediments rather than from tidal salt water inflow. The main contribution to this supply was brought from the deep layers by diffusion, acting over large parts of the wetland, but also advection effects in local discharge areas contributed significantly amounts. Later, by using a more elaborate model, /Price and Woo, 1990/ showed that local recharge-discharge processes have to be considered to get concentrations corresponding to measured values whereas only

Table 8-4. Amount and distribution of elements in 4 m of ombrotrophic peat from a bog in southern Sweden. From /Damman, 1978/.

Element	N	Na	K	Ca	Mg	P	Al	Fe	Mn	Pb	Zn
Amounts (kg ha ⁻¹)	10000	349	259	99	1259	382	358	470	9.9	31	20
0–5 cm (%)	0.8	1.4	11.0	0.8	0.3	1.8	1.0	1.3	8.5	8.1	2.8
5–35 cm (%)	6.3	4.4	21.8	10.2	3.3	9.5	19.3	26.3	59.7	42.1	46.8

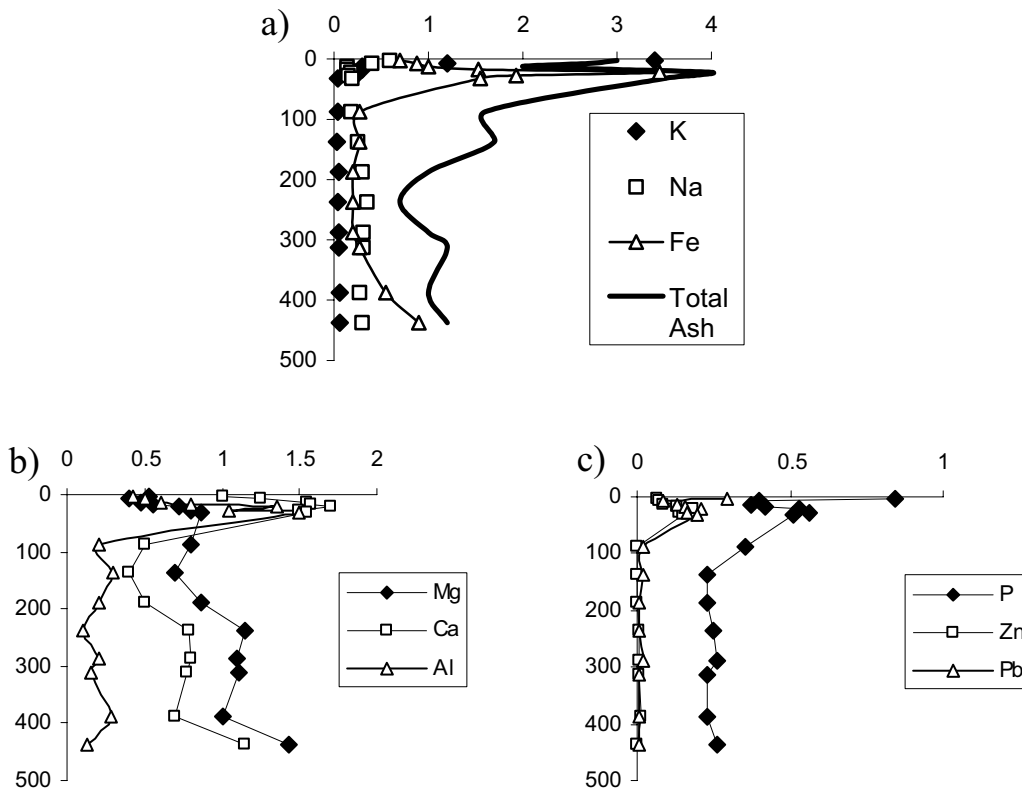


Figure 8-2. Distribution of a) K, Na, Fe ions and ash content; b) Mg, Ca and Al ions; c) P, Zn and Pb ions with depth (cm) in peat beneath a hummock at a raised bog in southern Sweden. Units are in mg/g dry weight except total ash content which is in % weight basis. Water table varied between 22 and 37 cm depth. From /Damman, 1978/.

diffusion could not explain the concentration patterns. /Reeve et al, 2001/ showed how mechanical dispersion could be responsible for measured concentration distributions in large bog-fen systems where the dominating flow is horizontal. However, /McKenzie et al, 2002/ showed that vertical flow influences pore-water chemistry in domed bogs by advective transport. It seems hence to be very important to be able to describe the groundwater flow pattern correctly and consequently must the hydraulic conductivity distribution be accurately assessed. These results show the possibility to use models to reveal knowledge of transport processes although more measurements on both physical and chemical properties are clearly needed.

9 Use and disturbance of wetlands

9.1 Impact by man

The impact of man on wetlands can not be overstated. In many areas, human activity has led to not only changes of wetland type and function but even to their total disappearance. /Löfroth, 2001/ regards that today, only ¼ of the Swedish wetlands are undisturbed considering the hydrology and at least ¼ of the original area covered by wetlands has disappeared. Further impacts from air pollution etc, leave very few current wetlands unaffected by human activity.

Traditionally, wetlands have been used as harvest land for cattle feed. In forest areas this was done at fens dominated by tall sedges, such as *Carex lasiocarpa* and *C. rostrata*. Shoreline marshes are still used extensively for grazing and have been used for harvesting, species like *Carex nigra*, *Agrostis stolonifera*, and *A. canina*, *Phalaris arundinacea*, *Scirpus lacustris* /Jacobson, 1997/. In the 19th century, drainage of wetlands and lowering of lakes to conquer new arable land started to take very large proportions and in the beginning of the 20th century, the cultivated peatland area was estimated to be 600 000 hectares /Löfroth, 2001/. The total area of wetlands drained for agriculture is probably much larger. In 1970 the total tile-drained agricultural area in Sweden was estimated to be about 1 million hectares, of which a large part probably has been considered to be too wet for agriculture, which should be close to the definition of wetlands /Löfroth, 2001/.

The Swedish forestry has likewise caused large impacts on wetlands. Up till now has about 1.5 million hectares of Swedish wetlands been drained for forestry purposes /Löfroth, 2001/. Another impact from forestry activity is the disturbance from establishment of forest roads.

Peat cutting is another type of disturbing or destroying exploitation of wetlands. Peat has been used for long as fuel and stable bedding in some areas and the consumption reached 4 millions m³ peat per year in the 1920:s, before other materials took over. Since 1950, use of peat as a substrate for growing plants and for soil improvement has been increasing. Except for occasional periods, the use of peat for fuel in Sweden has previously been of limited character, but it is growing and many are tempted to use peat as an alternative heat source to biomass or petroleum products. Peat is by tradition an important fuel in central Europe and on the British isles. Since the end of 1980:s, the peat harvesting for fuel is dominating also in Sweden with an average value (1990–2000) of 2.6 millions m³ piled peat per year while the peat cutting for horticultural use was 1 millions m³ piled peat per year for the same period, of which about 25% went to export. The current concessions granted for peat harvesting currently cover about 50 000 ha. All data in this paragraph are attained from /Statistiska centralbyrån, 2001/.

The use of undrained wetlands for cattle feed should not give any significant impact on their hydrology or hydrochemistry. There could be some transport of substances away from the wetland in connection with taking biomass which has grown there. However, clearcutting treed wetlands have shown to give some effects on the groundwater table /e.g. Dubé et al, 1995/.

Draining wetlands leads to changes in hydraulic head, which could lead to changes in water flow directions and size of flow. An increased component of upward flow and thus a more buffered water with an increased concentration of ions is often experienced after

drainage /Lundin, 1993; Stenbeck, 1996/. A change in hydraulic head also leads to a change in effective stress in the soil matrix /see e.g. Freeze and Cherry, 1979/. The increased soil stress may lead to compaction especially in peat soils which are highly compressive /Hobbs, 1986/. Compaction causes lower hydraulic conductivity, although very little is still known on how the conductivity changes. Compaction of the soil also causes subsidence of the ground surface.

As the soil gets drained a larger part will be aerated and the redox potential increases. The mobility of substances will get changed as new conditions establish. This is indeed found for uranium in oxidized peat /Kochenov et al, 1965/. In sulphide-rich soils this can lead to large drops in pH, as sulphur is oxidised into sulphate, with following large leaks of heavy metals. Organic-rich soils and especially peat soils get increased biological activity and decomposition. The decomposition and changed chemical environment generally increase the mobility of inorganic constituents /Heathwaite, 1991/, but also change the structure of the organic soil and eventually whole soil layers disappear as the organic material is decomposed. An extensive study on the drainage (for agricultural use) of the mire complex Bälänge mossar in Uppland, and its effects on the peat subsidence, is presented by /McAfee, 1985/, together with a study of theories on subsidence rates and their applicability.

Peat cutting will require drainage and drying of peat before it is transported away to the peat consumers. All steps in the exploitation are in some way influencing on the environment. Not only the local flora and fauna is wiped out but the peat per se and the quality and quantity of the outflowing water are affected. Both to its extent and pattern, the influence on the environment depends on climate, wetland type, peat type, topography and the localisation of peat cutting area and on the type of harvesting method /Stenbeck, 1996/. Intensive drainage can lead to large increases of peak runoff. A bare peat surface will have a lower infiltration capacity and thus contribute to a larger peak runoff. The elimination of vegetation cover will also decrease evaporation and lead to more total runoff.

The effect on water quality can vary greatly between areas regarding dissolved mineralization elements as well as erosion materials. A deeper unsaturated zone will lead to chemical changes in the peat as mentioned in the paragraph above. The increased surface runoff on bare peat will facilitate increased erosion. The size of particle transports then depends on the field slope and design of ditches and if sediment traps are present. In average, the total organic content from cutover peatlands is found to be 2–5 times the original concentration in unaffected outflows /Stenbeck, 1996/. Depending on the cutting method, the peat can be re-colonised by vegetation when abandoned. This has happened at strip-cutting areas, whereas flat milled-and-cut peat areas have been found to stay infertile for very long periods unless measures are taken to increase the surface moisture during growing seasons.

Using peat as fuel can have influence on air quality. The dust content in air during harvesting may have an impact although the radiation doses for peat-handling workers in current plots have been estimated to be small /Möre and Hubbard, 2003/. However, the release of substances while burning peat has been identified as a major concern and it sets the limits of how high concentration of radioactive substances should be allowed in peat fuel. In general, enrichment on the small flying particles leads to a higher concentration of more volatile substances in these fractions than in the average sized ash /Möre and Hubbard, 2003/. Within a heat plant, the enrichment of ^{210}Pb was found to be a factor 8 on small particles while corresponding factor for ^{137}Cs was found to be 2 /Mustonen and Jantunen, 1985, cited by Möre and Hubbard, 2003/. The transport of substances with smoke has been hard to estimate since the vapourisation of substances, their enrichment on flying particles and particle size distributions depend on several different factors such as type of peat, burning process and temperatures. /Fredriksson et al, 1984/ estimated that a concentration

of 200 ppm U in ash should be a suitable limit given a 10 MW heat plant with a 30% filtration of smoke. Small household fires without filters can then generate much larger local impact by particles in the smoke.

9.2 Animal impacts

The amount of animals in mires is usually small, mostly because of the soil oxygen deficiency, but also because of low pH and relative low nutrient concentrations. However, insects and larvae can be abundant especially in open water, (mosquitoes, diving beetles, dragon flies etc) and some amphibians (mainly frogs) occur. Enrichment of some substances can occur by the ecological nutrition chains, and animals having large parts of their food intake based on wetland species may be affected. This could be the case for birds and some rodents, e.g. voles, that inhabit mires and may accumulate some substances and also for their predators (snakes, owls and other raptorial birds). However, most mammals and birds living in connection to wetlands get their food from wide areas spanning over more than specific wetlands. Some herbivores (moose, roe deer) have wetland grasses and herbs as part of their food intake. Sphagnum is not eaten by any herbivore at all /Rydin et al, 1999/.

Very little effect on the wetland functions and processes can be expected by any animal with the exception of beavers, which can drastically change the hydrological conditions by dam building along watercourses and in places with occurring surface water /Mitsch and Gosselink, 2000/.

10 References

- Almquist-Jacobson H, Foster D R, 1995.** "Toward an Integrated Model for Raised-Bog Development – Theory and Field Evidence." *Ecology* 76(8): 2503–2516.
- Backéus I, 1985.** Aboveground production and growth dynamics of vascular bog plants in Central Sweden. *Acta Phytogeogr. Suec.* 74:1–98.
- Baden W, Eggelsmann R, 1963.** Zur Durchlässigkeit der Moorböden. *Z. Kulturtechnik und Flurberein* 4: 226–254.
- Baird A J, 1995.** Hydrological investigations of soil water and groundwater processes in wetlands. In: Hughes, J.M.R. Heathwaite, A.L. (eds). *Hydrology and hydrochemistry of British wetlands.* 111–129.
- Beckwith C W, Baird A J, 2001.** Effect of biogenic gas bubbles on water flow through poorly decomposed blanket peat. *Water resources research* 37: 551–558.
- Bellamy D J, Rieley J, 1967.** Some ecological statistics of a "miniature bog". *Oikos* 18: 33–40.
- Berglund K, 1996.** Properties of Cultivated Gyttja Soils. *International Peat Journal* 6: 5–23.
- Bergner K, Bohlin E, Albano Å, 1995.** Vad innehåller torv? : en sammanställning av botaniska, fysikaliska och kemiska data. 3e uppl. SLU, Umeå.
- Bergström S, 1994.** Sveriges Hydrologi. SMHI/Svenska Hydrologiska Rådet.
- Bergström U, Nordlinder S, Aggeryd I, 1999.** Models for dose assessments – Modules for various biosphere types. SKB TR 99-14, Svensk Kärnbränslehantering AB,
- Bloomfield C, 1972.** The oxidation of iron sulphides in soils in relation to the formation of acid sulphate soils, and of ochre deposits in field drains. *Journal of soil science* 23: 1–16.
- Boelter D H, 1968.** Important physical properties of peat materials. *Proceedings from the 4th international peat conference.* 150–154.
- Boelter D H, 1969.** Physical properties of peat as related to degree of decomposition. *Soil Science Society of America Journal*, 33: 606–609.
- Bohlin E, 1993.** Botanical composition of peat : implications on chemical, microbiological and physical properties. *Dissertations in forest vegetation ecology*, Swedish agricultural University, Umea. ISBN: 91-576-4709-7
- Brady N C, 1990.** The nature and properties of soils. 10th edition. Macmillan Press, New York.
- Bromley J, Robinson M, 1995.** Groundwater in raised mire systems: models, mounds and myths. In: Hughes, J.M.R. Heathwaite, A.L. (eds). *Hydrology and hydrochemistry of British wetlands.* 95–109.
- Brookings D G, 1988.** Eh-pH diagrams for geochemistry. Springer-Verlag Berlin, Germany.

- Brown P A, Gill S A, Allen S J, 2000.** “Metal removal from wastewater using peat.” Review paper, *Water Research* 34: 3907–3916.
- Brunberg A-K, Blomqvist P, 2000.** Post-glacial, land rise-induced formation and development of lakes in the Forsmark area, central Sweden. SKB TR-00-02, Svensk Kärnbränslehantering AB.
- Brydsten L, 2003.** A mathematiceal model for lake ontogeny in terms of filling with sediments and macrophyte vegetation. SKB technical report in press.
- Bunzl K, Schmidt W, Sansoni B, 1976.** Kinetics of ion exchange in soil organic matter. IV. Adsorption and desorption of Pb^{2+} , Cu^{2+} , Cd^{2+} , Zn^{2+} , and Ca^{2+} by peat. *Journal of soil science* 27: 32–41.
- Chason D B, Siegel D I, 1986.** Hydraulic conductivity and related physical properties of peat, Lost River Peatland, northern Minnesota. *Soil science* 142: 91–99.
- Chow T L, Rees H W, Ghanem I, Cormier R, 1992.** Compactibility of cultivated Sphagnum peat material and its influence on hydrologic characteristics. *Soil science* 153: 300–306.
- Clymo R S, Hayward P M, 1982.** The ecology of Sphagnum. In: Smith, A.J.E. (ed) *Bryophyte ecology*, Chapman and Hall: London; 229–289.
- Clymo R S, 1983.** Peat, In: Gore, A.J.P. (Ed.), *Ecosystems of the World 4A Mires: Swamp, Bog, Fen and Moor*, Elsevier Scientific Publishing Company, Amsterdam, pp. 159–224.
- Clymo R S, 1984.** The limits to peat bog growth. *Philosophical Transactions of the Royal Society of London B303*, 605–654.
- Clymo R S, Turunen J, Tolonen K, 1998.** Carbon accumulation in peatland. *Oikos* 81: 368–388.
- Damman A W H, 1978.** Distribution and movment of elements in ombrotrophic peat bogs. *Oikos* 30: 480–495.
- Devito K J, Waddington J M, Branfireun B A, 1997.** Flow reversals in peatlands influenced by local groundwater systems. *Hydrological processes* 11:103–110.
- Droppo I G, 2000.** Filtration in particle analysis. In: Meyers, R.A. (Ed) *Encyclopedia of analytical chemistry*. pp. 5397–5413.
- Dubé S, Plamondon A P, Rothwell R L, 1995.** Watering up after clear-cutting on forested wetlands of the St. Lawrence lowland. *Water resources research* 31: 1741–1750.
- Echevarra G, Sheppard M I, Morel J, 2001.** Effect of pH on the sorption of uranium in soils. *Journal of environmental radioactivity* 53: 257–264.
- Eggelsmann R, Heathwaite A L, Grosse-Brauckmann G, Küster E, Naucke W, Schuch M, Schweickle V, 1993.** Physical Processes and Properties of Mires, In: Heathwaite, A.L. and Göttlich, K.H. *Mires: Process, Exploitation and Conservation*. John Wiley and Sons, Chichester, UK, pp. 171–262.
- Engler R M, Patrick W H, 1975.** Stability of sulfides of manganese, iron, zinc, copper, and mercury in flooded and nonflooded soil. *Soil science* 119: 217–221.

- EPA (US Environmental Protection Agency), 2002.** Types of wetlands. Fact Sheet EPA 843-F-01-002b. Also at internet (July 2002): <http://www.epa.gov/OWOW/wetlands>
- Farnham R S, Finney H R, 1965.** Classification and properties of organic soils. *Advances in agronomy* 17: 115–162.
- Fraser C J D, Roulet N T, Moore T R, 2001.** Hydrology and dissolved organic carbon biogeochemistry in an ombrotrophic bog. *Hydrological processes* 15: 3151–3166.
- Fredriksson D, Ek J, Homberg B-T, 1984.** Uran och radioaktivitet i svenska torvmarker; förekomst, miljöaspekter, gränsvärden, analysmetoder. STEV-projekt 216023-1. BRAP 84017, Uppsala.
- Fredriksson D, 2002.** SGU, Uppsala. Pers. Communication. See also at internet (May 2004): http://www.sgu.se/geologi/jord_index.htm
- Freeze R A, Cherry J A, 1979.** *Groundwater*. Prentice-Hall Inc. New Jersey. 1–604.
- Frenzel B, 1983.** Mires – Repositories of climatic information or self-perpetuating ecosystems? In: Gore, A.J.P. (ed.) *Mires: Swamp, Bog, Fen and Moor. Ecosystems of the World 4A*. Amsterdam: Elsevier. 35–65.
- Frolking S, Roulet N T, Moore T M, Richard P J H, Lavoie M, Muller S D, 2001.** Modeling northern peatland decomposition and peat accumulation. *Ecosystems* 4: 479–498.
- Fuchsman C H, 1980.** *Peat: industrial chemistry and technology*. Academic press, New York.
- Fölster J, 2001.** Significance of processes in the near-stream zone on stream water acidity in a small acidified forested catchment. *Hydrological Processes* 15: 201–217.
- Giles B R, 1977.** Root Function of *Eriophorum angustifolium*. Thesis, University of London, London, 182 pp.
- Gorham E, 1955.** On some factors affecting the chemical composition of Swedish fresh waters. *Geochimica et cosmochimica acta* 7 : 129–150.
- Gorham E, Eisenreich S J, Ford J, Santelmann M V, 1985.** The chemistry of bog waters. In: Stumm, W. (Ed). *Chemical processes in lakes*. John Wiley and Sons, New York. 339–363.
- Gorham E, Underwood J K, Janssens J A, Freedman B, Maass W, Waller D H, Ogden J G, 1998.** The chemistry of streams in southwestern and central Nova Scotia, with particular reference to catchment vegetation and the influence of dissolved organic carbon primarily from wetlands. *Wetlands* 18: (1) 115–132
- Granlund E, 1932.** De svenska högmossarnas geologi : deras bildningsbetingelser, utvecklinghistoria och utbredning jämte sambandet mellan högmossbildning och försumpning. Sveriges geologiska undersökning. Stockholm. Serie C, 373.
- Grelle A, Lundberg A, Lindroth A, Moren A S, Cienciala E, 1997.** Evaporation components of a boreal forest: Variations during the growing season. *Journal of Hydrology* 197: 70–87.
- Grip H, Rodhe A, 1988.** *Vattnets väg från regn till bäck*. 2:a uppl. Hallgren & Fallgren, Uppsala. 1–156.

- Halbach P, von Borstel D, Gundermann K-D, 1980.** The uptake of uranium by organic substances in a peat bog environment on a granitic bedrock. *Chemical Geology* 29: 117–138.
- Heathwaite A L, 1991.** Solute transfer from drained fen peat. *Water, air and soil pollution* 55: 379–395.
- Helmer E H, Urban N R, Eisenreich S J, 1990.** Aluminum geochemistry in peatland waters. *Biogeochemistry* 9: 247–276.
- Hemond H F, Fifield J L, 1982.** Subsurface flow in salt marsh peat: a model and field study. *Limnology and oceanography* 27: 126–136.
- Hilbert D W, Roulet N, et al, 2000.** “Modelling and analysis of peatlands as dynamical systems.” *Journal of Ecology* 88(2): 230–242.
- Hillel D, 1998.** Environmental soil physics. Academic Press. 771pp.
- Hoag R S, Price J S, 1995.** A field-scale, natural gradient solute transport experiment in peat at a Newfoundland blanket bog. *Journal of hydrology* 172: 171–184.
- Hoag R S, Price J S, 1997.** The effects of matrix diffusion on solute transport and retardation in undisturbed peat in laboratory columns. *Journal of contaminant hydrology* 28: 193–205.
- 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.
- Hughes J H, 1999.** Contaminant fate processes. In: Bedient P. B. Rifai H.S. Newell C.H.: *Groundwater contamination*. Prentice Hall PTR.
- Idiz E F, Carlisle D, Kaplan I R, 1986.** Interaction between organic matter and trace metals in a uranium rich bog, Kern County, California, U.S.A. *Applied Geochemistry* 1: 573–590.
- Ingram H A P, 1983.** Hydrology. In: Gore A.J.P. (ed). *Ecosystems of the World 4A Mires: Swamp, Bog, Fen and Moor*, Elsevier: Amsterdam; 67–158.
- Irving M, Williams R J P, 1948.** Order of stability of metal complexes. *Nature* 162: 746–747.
- Ivanov K E, 1981.** Vodoobmen v bolotnykh landshaftakh. *Gidrometeoizdat; Leningrad*, 280. Translation by Thomson A, Ingram H.A.P. 1981. *Water movements in mirelands*. Academic Press: London; 277.
- Jacobson R, 1997.** Våtmarkernas värden i Uppsala län. Rapport nr 7, 1997. *Upplandsstiftelsen, Uppsala*.
- Kadlec R H, Keoleian G A, 1986.** Metal ion exchange on peat. In: Fuchsman, C.H. (ed.) *Peat and water*. Elsevier Applied Science Publishers Ltd. pp. 61–93.
- Kellner E, 2001.** Surface energy fluxes and control of evapotranspiration from a Swedish Sphagnum mire. *Agricultural and forest meteorology* 110: 101–123.
- Kellner E, Halldin S, 2002.** Water budget and surface-layer water storage in a Sphagnum bog in central Sweden. *Hydrological Processes* 16 (1): 87–103.

Kim J, Verma S B, 1996. Surface exchange of water vapour between an open Sphagnum fen and the atmosphere. *Boundary-Layer Meteorology* 79: 243–264.

Kivinen E, 1933. Über die Reaktionsverhältnisse in den Sphagnum fuscum- Bülden. *Soil Res.* 3:196–201.

Kneale P E, 1987. Sensitivity of the groundwater mound model for predicting mound topography. *Nordic hydrology* 18: 193–202.

Knott J F, Nuttle W K, Hemond H F, 1987. Hydrologic parameters of salt marsh peat. *Hydrological processes* 1: 211–220.

Kochenov A V, Zinevyev V V, Lovaleva S A, 1965. Some features of the accumulation of uranium in peat bogs. *Geochemistry International* 2:65–70.

Korhola A, Turunen J, 1996. The natural history of mires in Finland and the rate of peat accumulation. In: *Peatlands in Finland*. H. Vasander. Helsinki, Finnish Peatland Society: 20–26.

Lafleur P M, 1990. Evapotranspiration from sedge-dominated wetland surfaces. *Aquatic Botany* 37: 341–353.

Lafleur P M, Roulet N T, 1992. A comparison of evaporation rates from 2 fens of the Hudson-Bay lowland. *Aquatic Botany*, 44 : 59–69.

Landström O, Sundblad B, 1986. Migration of thorium, uranium, radium and Cs-137 in till soils and their uptake in organic matter and peat. SKB TR 86-24. Svensk Kärnbränslehantering AB.

Letts M G, Roulet N T, Comer N T, Skarupa M R, Versegby D L, 2000. Parameterization of Peatland Hydraulic Properties for the Canadian Land surface Scheme. *Atmosphere-Ocean* 38: (1): 141–160.

Lonnstad J, Löfroth M, 1994. Myrskyddsplan för Sverige. Statens naturvårdsverk, Stockholm. 416 pp.

Loxham M, 1980. Theoretical considerations of transport of pollutants in peats. In: *Proceedings of the 6th International Peat Congress, Duluth*. Pp. 600–606.

Lucisano M, Bozkurt S, 1998. Peat and peat decomposition, a literature survey. AFR-Report 227. AFN, Swedish environmental protection agency, Stockholm.

Lundberg A, 1996. Interception evaporation; Processes and measurement techniques. Doctoral thesis 1996:196D. Tekniska Högskolan Luleå.

Lundberg A, Halldin S, 2001. Snow interception evaporation. Review of measurement techniques, processes and models. *Theoretical and applied meteorology* 70:11–133.

Lundin L, 1993. Hydrologiska och hydrokemiska effekter av torvtäkt. Rapport Torv-94/1. NUTEK, Stockholm

Löfroth M, 1991. Våtmarkerna och deras betydelse. Solna, Naturvårdsverket.

Löfroth M, 2001. Våtmarkernas situation och långsiktiga förändringar. *KSLA tidskrift* 140: 5, 47–60.

- Malmer N, Sjörs H, 1955.** Some determinations of elementary constituents in mire plants and peat. *Botaniska notiser* 108: 46–80.
- Malmer N, 1958.** Notes on the relation between the chemical composition of mire plants and peat. *Botaniska notiser* 111: 275–288.
- Malmer N, 1962a.** Studies on mire vegetation in the Archaean area of southwestern Götaland (South Sweden) I. Vegetation and habitat conditions on the Akhult mire. *Opera botanica* 7(1): 5–322.
- Malmer N, 1962b.** Studies on mire vegetation in the Archaean area of southwestern Götaland (South Sweden) II. Distribution and seasonal variation in elementary constituents on some mire sites. *Opera botanica* 7(2): 1–67.
- Malmer N, 1963.** Studies on mire vegetation in the Archaean area of southwestern Götaland (South Sweden) III. On the relation between specific conductivity and concentration of ions in mire water. *Botaniska notiser* 116: 249–256.
- Mattson S, Koutler-Andersson E, 1954.** Geochemistry of a raised bog. *Kungliga Lantbrukshögskolans Annaler* 12: 101–118.
- McAfee M, 1985.** The rise and fall of Bälunge mossar; Bälunge mossars bildning och brukningshistoria. Rapport 147. Institutionen för markvetenskap, SLU, Uppsala.
- McKenzie J M, Siegel D I, Shotyk W, Steinmann P, Pfunder G, 2002.** Heuristic Numerical and Analytical Models of the Hydrologic Controls over Vertical Solute Transport in a Domed Peat Bog, Jura Mountains, Switzerland, *Hydrological processes*, 16: 1047–1064.
- McNamara J P, Siegel D I, Glaser P H, Beck R M, 1992.** Hydrogeologic controls on peatland development in the Malloryville Wetland, New York (USA). *Journal of hydrology* 140: 279–296.
- Mitsch W J, Gosselink J G, 1993.** *Wetlands*. 2nd ed. John Wiley & Sons Inc.
- Mitsch W J, Gosselink J G, 2000.** *Wetlands*. 3rd ed. John Wiley & Sons Inc.
- Moore K E, Fitzjarrald D R, Wofsy S C, Daube B C, Munger J W, Bakwin P S, Crill P, 1994.** A season of heat, water-vapor, total hydrocarbon, and ozone fluxes at a sub-arctic fen. *Journal of Geophysical Research-Atmospheres* 99: 1937–1952.
- Morris J C, Stumm W, 1967.** Redox equilibria and measurements of potentials in the aquatic environment. In: Stumm, W. (ed) *Equilibrium concepts in natural water systems*. *Adv. Chem.* 67. American chemical society, Washington. 270–285.
- Mustonen R, Jantunen M, 1985.** Radioactivity of size fractionated fly-ash emissions from a peat- and oil-fired power plant. *Health Physics* 49: 1251–1260.
- Möre H, Hubbard L-M, 2003.** Stråldoser vid användning av torvbränsle I stora anläggningar. SSI report 2003:02. Swedish radiation protection authority.
- National Wetlands Working Group, 1988.** *Wetlands of Canada. Ecological Land Classification Series No. 24.* Sustainable Development Branch, Environment Canada, Ottawa, Ontario and Polyscience Publications Inc, Montreal, Quebec.

- Naucke W, Heathwaite A L, Eggelsmann R, Schuc M, 1993.** Mire chemistry. In: Heathwaite, A.L. Göttlich, Kh. (eds) Mires. Processes, exploitation and conservation. John Wiley and Sons, Chichester. 263–309.
- Norrström A C, Jacks G, 1996.** Water pathways and chemistry at the groundwater/surface water interface to Lake Skjervatjern, Norway. *Water resources research* 32: 2221–2229.
- Nuttle W K, Hemond H F, Stolzenbach K D, 1990.** Mechanisms of water storage in salt-marsh sediments – the importance of dilation. *Hydrological Processes* 4: 1–13.
- Ours D P, Siegel D I, Glaser P H, 1997.** Chemical dilation and the dual porosity of humified bog peat. *Journal of hydrology* 196: 348–360.
- Penman H L, 1953.** The physical bases of irrigation control. In: Synge, P.M. Report of the Thirteenth International Horticultural Congress 1952, vol II. The Horticultural Society, London; 913–924.
- von Post L, Granlund E, 1926.** Södra sveriges torvtillgångar. Sveriges geologiska undersökning. Serie C, Avhandlingar och uppsatser 335 : 127 pp.
- Price J S, Woo M-K, 1988.** Studies of a subarctic coastal marsh, II. Salinity. *Journal of hydrology* 103: 293–307.
- Price J S, Woo M-K, 1990.** Studies of a subarctic coastal marsh, III. Modelling the subsurface water fluxes and chloride distribution. *Journal of hydrology* 120: 1–13.
- Price J S, 1994.** Evapotranspiration from a lakeshore typha marsh on Lake-Ontario. *Aquatic botany* 48: 261–272.
- Price J S, 2003.** The role and character of seasonal peat soil deformation on the hydrology of undisturbed and cutover peatlands. *Water resources research*, 19(9) 1241
doi: 10.1029/2002WR001302.
- Puigdomènech I, Bergström U, 1994.** Calculated distribution of radionuclides in soils and sediments. SKB Technical report 94-32. SKB AB Stockholm.
- Päivänen J, 1973.** Hydraulic conductivity and water retention in peat soils, *Acta For. Fenn.* Vol. 129, pp.1–70.
- Reeve A S, Siegel D I, Glaser P H, 2000.** Simulating vertical flow in large peatlands. *Journal of hydrology* 227: 207–217.
- Reeve A S, Siegel D I, Glaser P H, 2001.** Simulating dispersive mixing in large peatlands. *Journal of hydrology* 242: 103–114.
- Reid J A K, Stephenson M, Sheppard M I, Milley C A, 1994.** The wetland model for the assessment of Canada's nuclear fuel waste disposal concept. TR-569, COG-92-51. AECL Research, Canada.
- Reynolds W D, Brown D A, Mathhur S P, Overend R P, 1992.** Effect of in-situ gas accumulation on the hydraulic conductivity of peat. *Soil science* 153: 397–408.
- Ringqvist L, Holmgren A, Oborn I, 2002.** Poorly humified peat as an adsorbent for metals in wastewater. *Water research*, 36: 2394–2404.

- Romanov V V, 1968.** Hydrophysics of bogs. Kaner, N. (Translator); Heimann (ed), Israel program for scientific translations Ltd, Jerusalem; 1–299.
- Ross S M, 1995.** Overview of the hydrochemistry and solute processes in British wetlands. In: Hughes, J.M.R. Heathwaite, A.L. (eds). Hydrology and hydrochemistry of British wetlands. 133–181.
- Rudquist L, Fiskesjö A-L, 2000.** Den spännande sumpskogen : om Sveriges sumpskogar och dess själ. Skogsstyrelsen, Jönköping.
- Rydin H, Sjörs H, Löfroth M, 1999.** Mires. Acta Phytogeogr. Suec. 84: 91–112.
- Schiff S, Aravena R, Mewhinney E, Elgood R, Warner B, Dillon P, Trumbore S, 1998.** Precambrian shield wetlands: Hydrologic control of the sources and export of dissolved organic matter. Climatic change 40:167–188.
- Schlotzhauer S M, Price J S, 1999.** Soil water flow dynamics in a managed cutover peat field, Quebec: field and laboratory investigations. Water Resources Research 35: 3675–3683.
- Scott M J, Jones M N, Woof C, Tipping E, 1998.** Concentrations and fluxes of dissolved organic carbon in drainage water from an upland peat system. Environment international 24: 537–546.
- Seppä H, 1996.** The morphological features of the Finnish peatlands. In: Vasander, H. (ed.) Peatlands in Finland. Finnish peat society, Helsinki.
- Sheppard M I, Thibault D H, 1988.** Migration of technetium, iodine, neptunium, and uranium in the peat of two minerotrophic mires. Journal of environmental quality 17: 644–653.
- Shotyk W, 1988.** Review of the inorganic geochemistry of peats and peatland waters. Earth-science reviews 25: 95–176.
- Siegel D I, Glaser P H, 1987.** Groundwater flow in a bog-fen complex, Lost River peatland, northern Minnesota. Journal of ecology 75: 743–754.
- Siegel D I, Reeve A S, Glaser P H, Romanowicz E A, 1995.** Climate-driven flushing of pore water in peatlands. Nature 374: 531–533.
- Sjörs H, 1948.** Myrvegetation i Bergslagen. Acta Phytogeographica Suecia 21, Uppsala; 1–299.
- Sjörs H, 1950.** On the relation between vegetation and electrolytes in north Swedish mire waters. Oikos 2: 241–258.
- Sjörs H, 1956.** *Nordisk växtgeografi*. Scandinavian University books/Svenska Bokförlaget, Stockholm; 1–229.
- SKB, 1999.** Deep repository for spent fuel SR 97 – Post closure safety. SKB TR-99-06 Main Report volume I and II. Svensk Kärnbränslehantering AB.
- Souch C, Wolfe C P, Grimmond C S B, 1996.** Wetland evaporation and energy partitioning: Indiana Dunes National Lakeshore. Journal of hydrology 184:189–208.

- Statistiska centralbyrån, 2001.** Torv 2000. Tillgångar, användning, miljöeffekter. Statistiska meddelnaden Mi 25 SM 0101. ISSN:1403–8978.
- Stenbeck G, 1996.** Torvbruk-Miljö; Effekter och åtgärder. Rapport 4596, Naturvårdsverkets förlag. Stockholm. 94 pp.
- Tan K H, 1993.** Principles of soil chemistry. Marcel Dekker Inc. New York, USA. 362 pp.
- Wheeler B D, Shaw S C, 1995.** Plants as hydrologists? An assessment of the value of plants as indicators of water conditions in fens. In: Hughes, J.M.R. Heathwaite, A.L. (eds). Hydrology and hydrochemistry of British wetlands. 63–82.
- Wiklander L, 1976.** Marklära. Inst. för markvetenskap, Lantbrukshögskolan, Uppsala. 223 pp.
- Witting M, 1947.** Katjonsbestämningar i myrvatten. Botaniska notiser 4: 287–304.
- Witting M, 1948.** Preliminärt meddelande om fortsatta katjonsbestämningar i myrvatten sommaren 1947. Svensk botanisk tidskrift. 42: 116–135.
- Woo M-K, Winter T C, 1993.** The role of permafrost and seasonal frost in the hydrology of northern wetlands in North America. Journal of hydrology 141: 5–31.
- Zielinski R A, Meier A L, 1988.** The association of uranium with organic matter in Holocene peat: an experimental leaching study. Applied Geochemistry 3: 631–643.