The comments on melt-out till and the generation of meltwater for megafloods set up a red herring. We have been at pains to point out that the melt-out till precedes the megaflood and is commonly a remnant within erosional drumlins (e.g. Shaw *et al.*, 2000). To demonstrate this, we present some of the most detailed field sketches of fluting sediment together with structural and fabric data. A stone lag, interpreted to have been produced by flood erosion, lies on an erosional surface, truncating melt-out till and diapiric mélange. It is this erosional surface that defines the fluting. The melt-out till preceded the megaflood that eroded the drumlin and meltwater involved in the formation of the till was probably of little consequence to the flood which originated far to the north. Of course, as pointed out since 1982, the water for melt-out till was released slowly; we present estimates of thousands of years for melt-out till formation (Shaw *et al.*, 2000). Once again, we are misrepresented and the arguments we make for a supraglacial origin for the megaflood discharge are overlooked.

Obviously, there is a major difference in perception between Benn and Evans and us. They consider that the megaflood hypothesis flies in the face of a huge body of mainstream research. In our defence, there is no known observation that contradicts the hypothesis. Nor does this hypothesis violate any fundamental principle in science. It might be incompatible with mainstream research, but the same can be said of any new paradigm. In answer to their assertion that our work is unscientific and unnecessary, our only response is—we do not think so. It would help if Benn and Evans were more specific where they write that our work is inconsistent with the evidence. Again, there are no known observations that *contradict* the megaflood hypothesis.

NINE

Groundwater under ice sheets and glaciers

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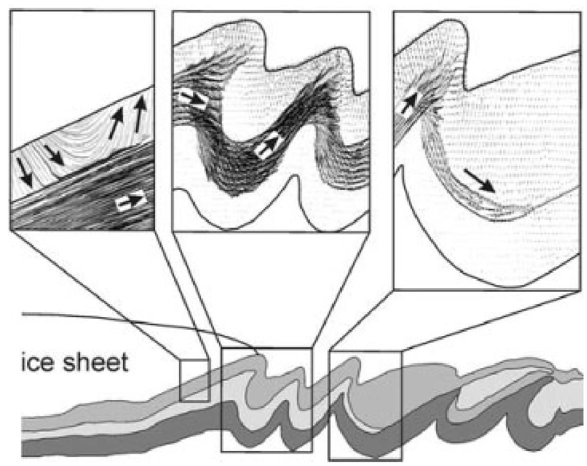


9.1 Introduction

It has been realized only recently that groundwater under ice sheets and glaciers is an important, integral part of the hydrological system in environments affected by glaciation. The late start in research and the resulting scarcity of published work on subglacial groundwater was caused by its position in the no-man's land between glaciology and hydrogeology, despite its relevance for both. It is now recognized that water in permeable rocks and soft sediments overridden by glaciers, through a system of feedbacks, influences glacier stability, movement mechanisms, sediments and landforms. Water discharged from melting ice contributes to the renewal of groundwater resources.

Large-scale groundwater circulation patterns and dynamics experience fundamental changes in glacial–interglacial cycles subjected to repeated loading and relaxation by kilometre-thick ice. Old glacial groundwater trapped in low-permeability areas yields important information about past environmental changes, and modifications of future groundwater flow dynamics in areas likely to be affected by prospective ice sheets must be considered in disposal strategies of toxic waste. It is therefore clear how important the impact of glaciation on groundwater is, which explains the recent interest in this field.

9.2 Water source and drainage systems

Subglacial meltwater originates from a range of sources, mainly from melting of ice by geothermal heat trapped at the glacier sole and by the frictional heat caused by ice movement past the substratum. These two sources yield up to some 100 mm yr^{-1} of water. Close to the ice margin, in the area where englacial conduits extend to the bed, surface ablation water may reach the ice sole with recharge several orders of magnitude greater than the basal meltwater alone. It is difficult to estimate how wide this area is, but as deeper conduits tend to close under cryogenic pressure or they bend horizontally towards the ice margin, it is probable that ablation water would reach the bed only where the ice thickness is less than about 100–200 m (Reynaud, 1987; Fountain & Walder, 1998). Under polythermal conditions, such as those that characterized continental Pleistocene ice sheets, the outermost marginal zone and the ice divide zone may be cold-based and do not contribute to the basal water recharge.

Subglacial water can be evacuated to the ice margin through high-discharge drainage systems consisting of channels incised into the glacier sole (R-channels; Röthlisberger, 1972) and channels carved into the bed (N-channels; Nye, 1973). R-channels typically form on bedrock, whereas N-channels tend to occur on soft beds. This is shown by the distribution of landforms across different substrata overridden by the Fennoscandian and Laurentide ice sheets where on bedrock areas eskers dominate, whereas in areas of sedimentary basins away from glaciation centres meltwater channels and tunnel valleys prevail (Clark & Walder, 1994).

Another type of subglacial drainage is the low-discharge, distributed system operating at the ice–bed interface or in the rocks below. A linked-cavity network consists of broad and shallow water lenses connected by orifices, favoured by rapid sliding and high bed roughness (Nye, 1970), which originally was suggested for hard bedrock areas by Lliboutry (1976) and Kamb (1987). A modification of this drainage mechanism was proposed by Walder & Fowler (1994) and Clark & Walder (1994) for soft, deformable beds. It consists of broad, shallow 'canals' interconnected in a non-arborescent system without orifices, capable of evacuating more water than the linked cavities. Yet another mechanism is the subglacial water film, typically about 1 mm thick, generated by regelation at the ice–rock interface (Weertman, 1972). Under special circumstances, parts of a glacier may be lifted by pressurized water leading to short-lasting subglacial sheet-floods of high magnitude.

9.3 Basic laws of groundwater flow through the subglacial sediment

If a glacier or an ice sheet rests on a permeable bed, be it rock or soft sediment, a part of the subglacial meltwater will enter the bed and be evacuated as groundwater flow, governed by the same physical rules as groundwater flow in confined aquifers outside the glaciated areas. The major differences, however, are that (i) the flow is driven by hydraulic gradient imposed by ice overburden, and (ii) some groundwater may be advected within the sediment if it deforms in response to glacier stress.

The basal meltwater will enter the substratum if the pressure at the ice–bed interface is greater than the pressure within the bed. If the bed is a porous medium such as glacial deposits and most other sedimentary rocks, water flow will be governed by the Darcy law

$$Q = \left(\frac{KA}{\rho_w g}\right)\left(\frac{dh}{dl}\right)$$

where Q is the water flux, K is the hydraulic conductivity, A is flow cross-sectional area, ρ_w is the density of water, g is the acceleration due to gravity, h is the hydraulic head and l is the flow length. The flow is driven by the hydraulic gradient $(1/\rho_w g)(dh/dl)$ determined largely by the ice-surface slope (Shreve, 1972; Fountain & Walder, 1998). The potentiometric surface of groundwater confined by an ice sheet runs approximately parallel to the ice surface at a depth corresponding to the water pressure (Fig. 9.1A). At the ice margin the pressure in the aquifer is atmospheric, thus the groundwater changes from a confined to unconfined flow. Under conditions of high meltwater recharge and low drainage capacity of both the bed and the drainage systems at the ice–bed interface, the groundwater pressure may be elevated to the vicinity of the ice overburden pressure (glacier flotation point). Under such conditions groundwater recharge ceases because of equilibrated pressures and lack of head drop into the sediment.

Groundwater flow under polythermal glaciers and ice sheets will be affected by the distribution of frozen ground in ice-marginal areas (e.g. Haldorsen *et al.*, 1996; Cutler *et al.*, 2000) and under central parts of ice sheets where basal freezing may occur due to advection of cold ice (Fig. 9.1B). Because frozen soil is orders of magnitude less permeable than the same soil in unfrozen state (Williams & Smith, 1991), permafrost will act as a confining layer. Pressurized groundwater thus will be forced under the permafrost, and the confined drainage system will first terminate outside the permafrost zone or at large discontinuities in the frozen ground such as taliks under rivers, lakes and over salt domes.

Pressurized, often artesian (Haldorsen *et al.*, 1996; Flowers & Clarke, 2002b; Flowers *et al.*, 2003) groundwater at a glacier margin can cause hydrofracturing of sediment and rocks. Boulton & Caban (1995) suggested that such fractures may occur to a depth of about 400 m, and the zone affected by hydrofracturing can extend many tens of kilometres into the ice foreland. They gave examples of small-scale sediment dykes filling hydrofractures in Spitsbergen, England and Sweden (see also Larsen & Mangerud, 1992; van der Meer *et al.*, 1999; Grasby *et al.*, 2000). Plate 9.1 shows a hydrofracture of similar origin in northwest Germany.

Pressurized groundwater may also facilitate formation of push moraines, diapirs and large-scale extrusion moraines (Boulton *et al.*, 1993; Boulton & Caban, 1995), typically by sediment liquefaction. Artesian groundwater behind an ice front is capable of lifting loosened, thick rock slabs (Bluemle & Clayton, 1984; Pusch *et al.*, 1990), which may then be redeposited by ice as allochthonous rafts and megablocks within glacial sediments (Aber *et al.*, 1989). Water pressure increase will also occur in subglacial aquifers that wedge out in the direction of glacier flow. Such groundwater traps were an important cause of glaciotectonism in northwestern Germany during the last glaciation (Piotrowski, 1993), and facilitated glaciotectonism along the Main Stationary Line in Denmark (Piotrowski *et al.*, 2004; Fig. 9.2). Deformation also may be caused by fast ice retreat where a swath of land with pressurized groundwater is exposed, causing sediment blow-ups before pressure equilibrium is established.

Groundwater flow dynamics, especially the flow field, velocity and rate, strongly depend on the substratum porosity and hydraulic conductivity, which act as 'knobs that open and close subglacial drainage valves' (Flowers & Clarke, 2002a). In the case of sorted granular materials such as sand and gravel, these parameters do not differ much from those under non-glacial conditions. More difficult to estimate are hydrogeological parameters of subglacial tills, yet this sediment type is of particular impor-

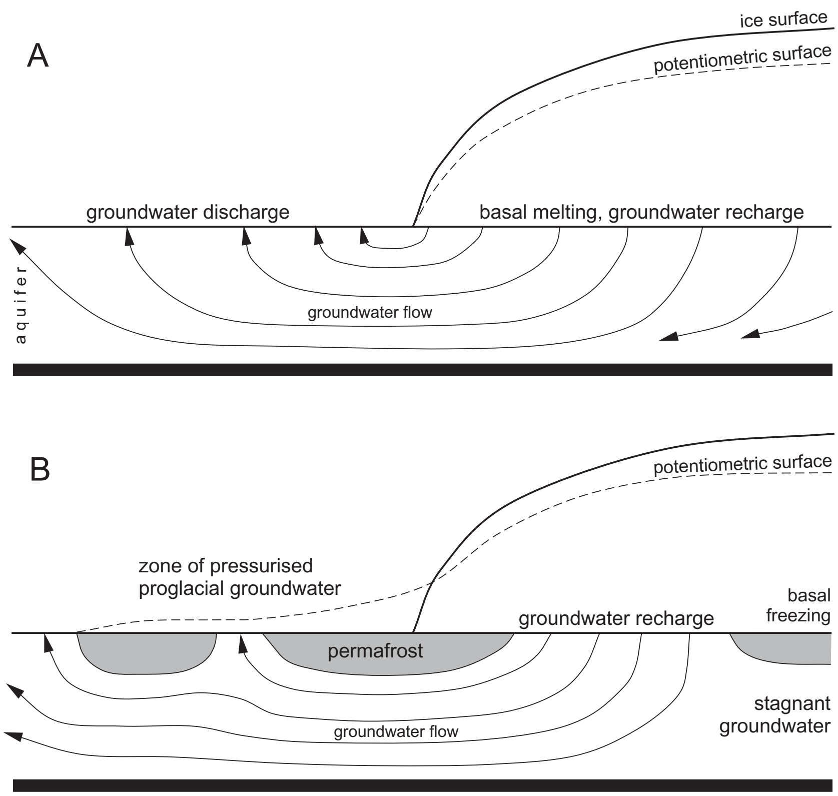


Figure 9.1 Schematic representation of groundwater drainage through a subglacial aquifer under a glacier resting on unfrozen (A) and partly frozen (B) bed.

tance because it occurs immediately under most glaciers and is the first sediment on the pathway of the subglacial groundwater into the bed.

When a basal till is dilated due to shear deformation and high porewater pressure, its porosity is around 0.4 (Blankenship *et al.*, 1986; Engelhardt *et al.*, 1990b), whereas non-dilated tills appear to have porosities around 0.25–0.3 (Fountain & Walder, 1998). The difference corresponds to an increase in flow velocity of up to about 10 times. Hydraulic conductivity of tills from past glaciations varies over several orders of magnitude, typically between 10^{-1} and 10^{-7} m s^{-1} (Freeze & Cherry, 1979), with an extreme effect on groundwater fluxes (see the Darcy equation).

Conductivity of a till under a glacier can be expected to vary accordingly, but its direct determination is complicated owing to limitations in the accessibility of glacier beds. Reliable estimates are, therefore, scarce. Examples of *in situ* estimates using different methods, and some laboratory tests on till samples collected from under modern glaciers and ice sheets, are given in Table 9.1 and show the extreme spread of values between some 10^{-2} and 10^{-12} m s^{-1}. Worth noting is that *in situ* measurements tend to yield higher conductivities than laboratory tests, owing to large-scale non-Darcian flow pathways and large-scale till heterogeneities (e.g. Gerber *et al.*, 2001), such as sand and gravel lenses, known to occur in nature but not accounted for in laboratory tests.

Besides the textural composition, also compression and shear deformation influence the conductivity of subglacial tills. Both result in highly anisotropic distribution of till properties with horizontal conductivity up to two orders of magnitude greater than the vertical one owing to particle alignment, as shown by laboratory experiments (e.g. Murray & Dowdeswell, 1992). Till deformation in response to glacier stress may increase the conductivity (in the case of dilation; Clarke, 1987b) or decrease it (in the case of compaction), as demonstrated experimentally by Hubbard & Maltman (2000). They also showed that hydraulic conductivity in tills is inversely related to the effective glacier pressure because pressurized porewater prevents sediment compaction and closure of drainage pathways.

If till undergoes pervasive deformation, substantial volumes of groundwater may be advected toward the ice margin within the high-porosity deforming layer (Alley *et al.*, 1986a, 1987a). Such deformation may destroy drainage paths at the ice–bed interface (Clarke *et al.*, 1984), but it will enhance drainage through the bed (Murray & Dowdeswell, 1992). Despite our still fragmentary knowledge about hydrogeological properties of tills under glaciers, one can safely conclude that these properties vary extremely both in space and time, influenced by the nature of the source material and stresses applied by the overriding glacier.

9.4 Subglacial groundwater in past and modern environments

As ice sheets grow and expand over permeable rocks, groundwater flow evolves from a subaerial, precipitation-fed system con-

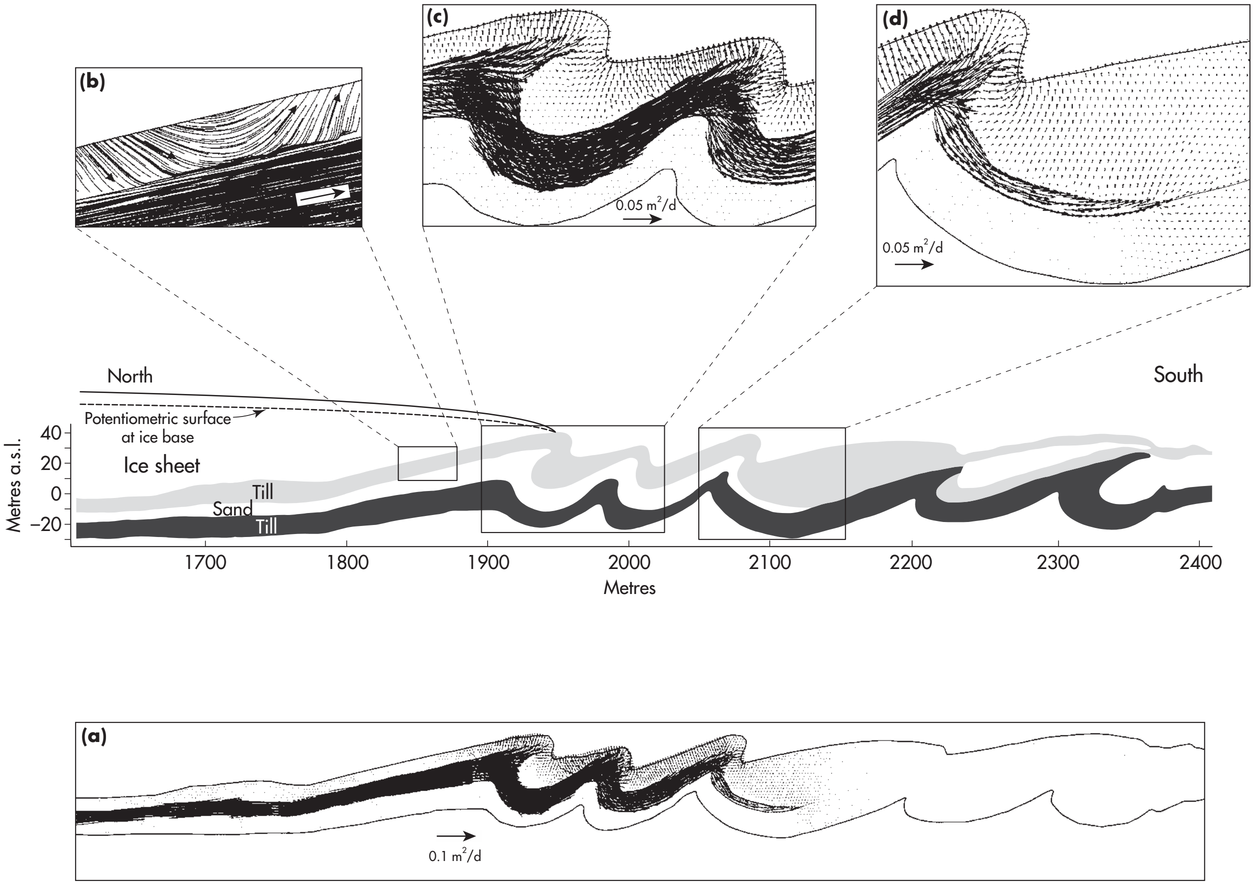


Figure 9.2 Modelled groundwater flow pattern under the margin of the Weichselian ice sheet at its maximum extent at the Main Stationary Line in Denmark (Bovbjerg). Large glaciotectonic folding was facilitated by high porewater pressure in the low-transmissivity bed, partly due to thin aquifers wedging out in the direction of groundwater flow. Note that the transition from groundwater recharge to discharge still occurs under the ice sheet (b), caused by the low hydraulic conductance of the bed not capable of evacuating all meltwater to the ice foreground (Piotrowski *et al.*, 2004).

Table 9.1 Hydraulic conductivities of some subglacial tills derived from *in situ* and laboratory measurements

Location	Hydraulic conductivity (ms^{-1})	Reference
Bakaninbreen	3×10^{-7} to 8×10^{-8}	Porter & Murray (2001)
Bakaninbreen	8×10^{-3}	Kulessa & Murray (2003)
Midre Lovénbreen	2×10^{-5}	Kulessa & Murray (2003)
Trapridge Glacier	5×10^{-4}	Stone *et al.* (1997)
Trapridge Glacier	$1–2 \times 10^{-9}$	Waddington & Clarke (1995)
Trapridge Glacier	2×10^{-8}	Murray & Clarke (1995)
Trapridge Glacier	1×10^{-8}	Flowers & Clarke (2002a)
South Cascade Glacier	10^{-7} to 10^{-4}	Fountain (1994)
Haut Glacier d'Arolla	10^{-7} to 10^{-4}	Hubbard *et al.* (1995)
Haut Glacier d'Arolla	8×10^{-12} to 9×10^{-9}	Hubbard & Maltman (2000)
Breidamerkurjökull	$1–2 \times 10^{-6}$	Boulton *et al.* (1974)
Breidamerkurjökull	6×10^{-7} to 4×10^{-4}	Boulton & Dent (1974)
Storglaciären	10^{-7} to 10^{-6}	Iverson *et al.* (1994)
Storglaciären	10^{-9} to 10^{-8}	Fischer *et al.* (1998)
Storglaciären	10^{-7} to 10^{-6}	Baker & Hooyer (1996)
Gornegletscher	2×10^{-2}	Iken *et al.* (1996)
Ice Stream B	2×10^{-9}	Engelhardt *et al.* (1990b)
Ice Stream B	10^{-9}	Tulaczyk *et al.* (2001a)

trolled by landscape topography, to a system pressurized by and recharged from the overlying ice. Studies reconstructing water flow in subglacial sediments are very scarce, but they all show that flow pattern and velocity differed substantially from the interglacial situation in the same areas. Owing to the complexity of the glacial and groundwater systems, numerical methods have been applied following the pioneering study of Boulton & Dobbie (1993). They have demonstrated that under glaciers resting on aquifers with high hydraulic transmissivity, all basal meltwater can be evacuated through the bed and no other drainage systems at the ice base need to form. If, on the other hand, a glacier is underlain by fine-grained sediment, its conductance may be insufficient triggering formation of more efficient drainage pathways, such as channels. This conclusion is important for inferences on ice movement mechanisms and for interpreting the origin of eskers and tunnel valleys, which, accordingly, should indicate areas with excess of basal meltwater.

9.4.1 Did all basal meltwater drain through the bed?

Boulton & Dobbie (1993) suggested that areas in Holland, during the Saalian glaciation, had sufficient transmissivity to drain all meltwater from the ice sheet base. This inference was based on calculations with very low basal melting rates of 2–3 mm yr^{-1}, but claimed valid also in subsequent simulations with melting rates of 20 mm yr^{-1} (Boulton *et al.*, 1993). In a numerical model of groundwater flow under ice sheets of the last two glaciations along a transect from the ice divide in Scandinavia to the ice periphery in Holland, Boulton *et al.* (1995) used a melting rate of 25 mm yr^{-1} with a similar conclusion. In their model, however, the entire Quaternary sequence was lumped together into one aquifer with conductivity of 3×10^{-4} m s^{-1}, corresponding to the conductivity of sand. This value is clearly too high, neglecting the true nature of Quaternary strata characterized by interlayered sediments of much lower bulk transmissivity. The hypothesis of basal water drainage entirely through the bed is also inconsistent with widespread sediment deformation postulated by Boulton (1996a), because such deformation will occur only if porewater pressure is at least 90–95% of the overburden pressure (Paterson, 1994, p. 169), that is when the bed capacity to absorb meltwater is reached. As pointed out by Arnold & Sharp (2002), the modelling of Boulton and co-workers is seemingly contrary to most geological evidence. More recently, Boulton *et al.* (2001b) proposed a revised model of basal hydraulic regimes that considers a zone where the meltwater flux is too large to be discharged by groundwater flow alone, and in which tunnels form.

In a three-dimensional model of groundwater flow under the Weichselian ice sheet in northwestern Germany that accounts for the heterogeneity of Quaternary sediments, located close to Boulton's transect, Piotrowski (1997a) showed that only about 25% of meltwater could have been evacuated through the bed, and the rest was drained in spontaneous outburst episodes through subglacial channels. This is supported by the occurrence of tunnel valleys, some up to ca. 80 m deep (Piotrowski, 1994), abundantly found in this area. Modelling the adjacent area around the Eckernförde Bay showed that about 30% of basal meltwater drained as groundwater flow, and the rest through channels (Marczinek, 2002; Marczinek & Piotrowski, this volume, Chapter 10). It should be noted that these simulations were made with conservative basal melting rates of 36 mm yr^{-1} not considering surface ablation water, which probably recharged the bed close to the ice margin. Accounting for this additional water source would further substantiate the conclusion about the insufficient drainage capacity of the bed, consistent with the formation of tunnel valleys as high-discharge drainage pathways. It also should be stressed that if a permafrost wedge under the ice-sheet margin is considered, the bed conductivity will be yet lower (Fig. 9.3), further increasing the likelihood of tunnel valley formation.

It is tempting to suggest that similar hydraulic deficiency of the bed could have initiated tunnel valleys elsewhere across the Central European Lowland and at the bottom of the North Sea, where they represent the largest glacial features of the Elsterian and Weichselian glaciations, some over 500 m deep and hundreds of kilometres long (e.g. Huuse & Lykke-Andersen, 2000). If tunnel drainage is predicted to have extended up to 150 km from the ice sheet margin (Arnold & Sharp, 2002), then some of these tunnel valleys possibly formed time-transgressively during deglaciation.

Because subglacial channels form in response to the excess of water at the ice sole, a succession of drainage mechanisms operating in cycles can be envisaged. Each cycle starts with low water pressure in the bed and groundwater recharge, followed by gradual increase of water pressure, formation of a basal water layer or a lake at the ice flotation pressure, succeeded by spontaneous drainage through channels, lowering the water pressure and ending with basal recoupling (Piotrowski, 1997a). Formation of the channels would thus act as a stabilizing feedback preventing widespread basal decoupling and catastrophic collapse of an ice lobe. Origin of subglacial lakes owing to meltwater ponding in low-transmissivity areas of northern Germany was also considered by van Weert *et al.* (1997), and in marginal zones of the Laurentide Ice Sheet by Cutler *et al.* (2000), in accordance with the considerations of Shoemaker (1991) who gave a physical framework for the formation of such lakes.

Most reconstructions of groundwater flow under past ice sheets in North America and Europe indicate a deficiency of bed materials to evacuate all the incoming meltwater, which is interesting, bearing in mind the diversity of beds overridden by ice sheets on the two continents. Brown *et al.* (1987) showed that only a fraction of basal meltwater could have been evacuated through a soft substratum under the last glacial Puget Lobe of the Cordilleran Ice Sheet. They envisage a widespread basal decoupling by water pressurized to the ice flotation level, indicated among other things by lenses and layers of sand in till, and low overconsolidation of subglacial clays. The same line of sedimentological and geotechnical evidence was used by Piotrowski & Kraus (1997) and Piotrowski & Tulaczyk (1999) to suggest bed drainage deficiency in northern Germany. Shoemaker (1986) demonstrated that even under an ice sheet resting on a metre-thick gravel bed, channels must develop to keep porewater pressures below the ice overburden pressure, which was later confirmed by Ng (2000a). Two recent studies along transects from the Hudson Bay to the southern margin of the Wisconsinan ice sheet independently show that subglacial aquifers were not capable of evacuating the basal meltwater, possibly due to a combination of low bed-permeability and permafrost, compensated by channelized drainage (Breemer *et al.*, 2002) or subglacial water storage (Cutler *et al.*, 2000). Under

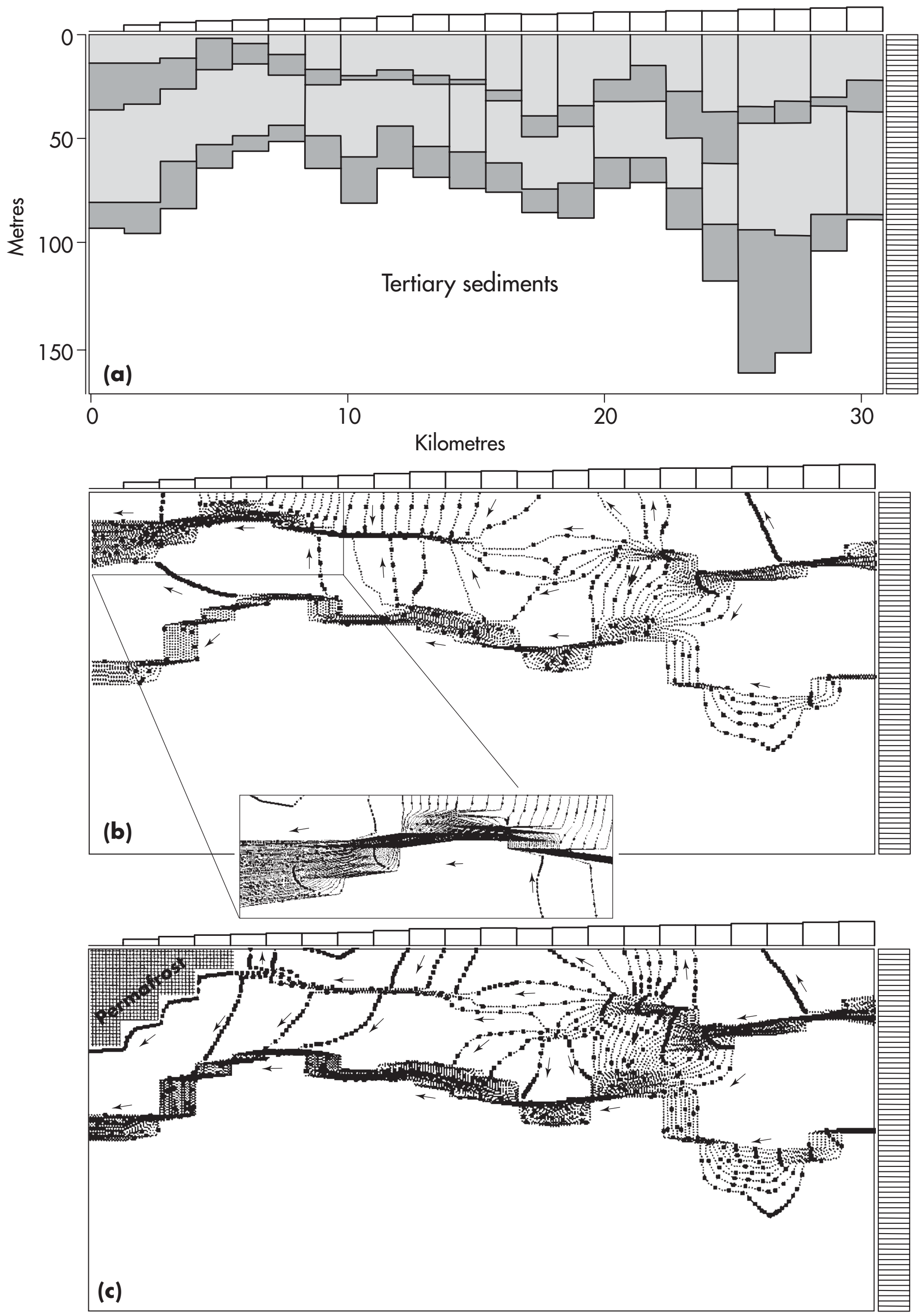


Figure 9.3 Numerical model of groundwater flow under the maximum extent of the Weichselian ice sheet in northwestern Germany along an ice-flow parallel transect between the ice margin (0 km) and the present Baltic Sea coast (30 km). Geology (a) is generalized into two aquifers (dark grey) and two aquitards (light grey) resting on impermeable Tertiary sediments. Groundwater flow lines are shown with time markers and direction arrows. Scenario in (b) assumes non-permafrost conditions, and scenario in (c) considers a permafrost wedge under the ice margin. Note that in (c) the whole upper aquifer is frozen, so that (i) more groundwater drains through the lower aquifer, (ii) less water recharges the bed from the ice sole, and (iii) more water is forced upward toward the ice sole from the bed. (From Piotrowski, 1997b.)

the Green Bay Lobe, only raising the conductivity values by three orders of magnitude would have allowed all basal melt to be evacuated as groundwater (Cutler *et al.*, 2000). A comprehensive numerical study of the hydrology under the largest European ice cap, Vatnajökull, showed that buried aquifers may only evacuate up to ca. 30% of subglacial water (Flowers *et al.*, 2003). Under Trapridge Glacier, Yukon, the aerial extent of aquifer saturation is 70–90%, with the groundwater drainage system at the edge of its capacity beneath the ablation zone (Flowers & Clarke, 2002b).

Also worth noting is that data from beneath the Whillan's Ice Stream (formerly Ice Stream B), Antarctica, indicate the substratum inefficiency in capturing basal meltwater. Lingle & Brown (1987) suggested that the subglacial water discharge mechanism is advection in the deforming layer, but in the light of a much thinner zone of deformation than originally assumed (Engelhardt & Kamb, 1998) this mechanism is not likely to be efficient. Subglacial ponding and localized ice decoupling are likely, among other effects, owing to the postulated upwards-directed groundwater flow in parts of the bed (Tulaczyk *et al.*, 2000b). Likewise, marine-based ice sheets terminating on shelf sediments that tend to be fine-grained could not have discharged all the melt through the bed (Hindmarsh, 1999).

9.4.2 Interaction between groundwater and water in subglacial channels

Where glaciers rest directly on aquifers or are separated from them only by a thin layer of low-permeability sediment, the groundwater is in contact with the water in channels at the ice–bed interface. Channel formation reduces the water pressure and generates a hydraulic gradient, which will drive groundwater from the surrounding sediment into the channel and create a catchment area along the channel. This was suggested independently by Shoemaker & Leung (1987) and Boulton & Hindmarsh (1987), and recently measured in piezometers nested around a channel under Breidamerkurjøkull in Iceland (Boulton *et al.*, 2001b).

Radial groundwater flow to the channel can fluidize the sediment, followed by its injection into the channel and subsequent removal by water flow. In a steady-state situation, this mechanism could produce large incisions resembling tunnel valleys by erosion in relatively narrow R-type channels serving as sediment sinks for the surrounding aquifers (Boulton & Hindmarsh, 1987). Possible capturing of groundwater by tunnel valleys was also suggested for one prominent tunnel valley tract in northern Germany (Piotrowski, 1994). However, it was postulated that an increase of effective pressure at the channel flanks would strengthen the sediment there and, in particular, prevent pervasive deformation of the glacier bed.

Modelling by Boulton *et al.* (2001b) and Fleming & Clark (2000) shows a progressively damped pressure wave propagating away from a channel into the sediment for tens of metres, consistent with some field measurements (e.g., Fountain, 1994; Hubbard *et al.*, 1995; Murray & Clarke, 1995). Boulton *et al.* (2001b) predicted a bulb of low pressure in a shallow zone beneath the channel. Occurrence of this low-pressure zone is confirmed by Piotrowski *et al.* (1999), who documented a soft-sediment diapir under a subglacial channel of Saalian age in eastern Germany. The diapir demonstrates sediment creep into the channel from below, and yields support for theoretical considerations of Boulton & Hindmarsh (1987) and Shoemaker & Leung (1987) suggesting a sediment mobilization into a channel and erosion rates higher than could be estimated from channel dimensions alone.

Coupling of water flow in channels and in the bed also implies that, under specific circumstances such as a wave of ablation water reaching a subglacial channel or channel blockage by sediment injection or roof collapse, a water pressure increase will propagate into the aquifer. This will happen with delay, caused by hydraulic resistance of the sediment, particularly in low-conductivity, fine-grained aquifers. The result will be a reversal of hydraulic gradient, now driving the water from the channel into the aquifer. Frequent shifts in groundwater flow directions close to channels would be expected where the subglacial hydrology is dominated by diurnal or seasonal ablation cycles, i.e. close to a glacier margin.

9.4.3 Groundwater flow dynamics under past ice sheets and preservation of old glacial groundwater

Numerical modelling shows that groundwater flow velocities and hydraulic heads in northwestern Europe, especially in the relatively shallow aquifers, were significantly higher under ice sheets than they are at present. For an area bordering the Baltic Sea in Germany, Piotrowski (1997b) estimated flow velocities in the upper aquifer under the last ice sheet as about 30 times higher than at present. Furthermore, a reversal of flow direction occurred. At present, the groundwater drains to the Baltic Sea; during the glaciation it was forced in the opposite direction by the ice sheet advancing out of the Baltic Sea basin. A corresponding flow reversal was also determined by Boulton *et al.* (1996) and van Weert *et al.* (1997) in a large-scale, vertically integrated palaeoflow model (Fig. 9.4), by Glynn *et al.* (1999) in a transect between the ice-sheet centre in Scandinavia and its periphery in Poland, by Grasby *et al.* (2000) in a transect from South Dakota to Manitoba during the last glaciation, and by Marczinek & Piotrowski (this volume, Chapter 10) in a coastal area of northwest Germany. Maximum flow velocities in upper subglacial aquifers in northwestern Europe were estimated at 20 m yr^{-1} (Boulton *et al.*, 1995), 200 m yr^{-1} (Boulton *et al.*, 1996) and over 100 m yr^{-1} (van Weert *et al.*, 1997).

Given the high flow velocities and hydraulic heads, groundwater can penetrate deep into the subglacial sediments and rocks. Because of the typically layered structure of sedimentary beds, the flow pattern is highly anisotropic with preferential flow approximately horizontal in the transmissible, coarse-grained sediments and vertical in aquitards. Most studies indicate that upper aquifers with residence times in the order of several thousand years were completely flushed out during glaciations and the groundwater derived from precipitation during the interglacials was replaced by glacial meltwater. In Holland the entire upper (Quaternary) aquifer system, about 300 m thick, and a large part of the lower (Mesozoic) aquifer, up to 1500 m thick, are believed to have been flushed under the Saalian ice sheet, whereas the Tertiary aquitard in between was partly penetrated by meltwater (Boulton *et al.*, 1993). Deep meltwater circulation is also suggested for northwestern Germany, where during the Weichselian glaciation the flow field in the whole Quaternary sequence consisting of two major aquifers and two aquitards up to ca. 200 m thick was reorganized by ice overriding (Piotrowski, 1997a,b). Similar conclusions were drawn for the Illinois basin, USA where the several-kilometres-thick drainage system of superposed aquifers and aquitards was significantly modified by the last ice sheet, as modelling (Breemer *et al.*, 2002) and hydrochemical analyses (McIntosh *et al.*, 2002) indicate. Under the Des Moines lobe in Iowa, subglacial meltwater penetrated over 300 m down into the

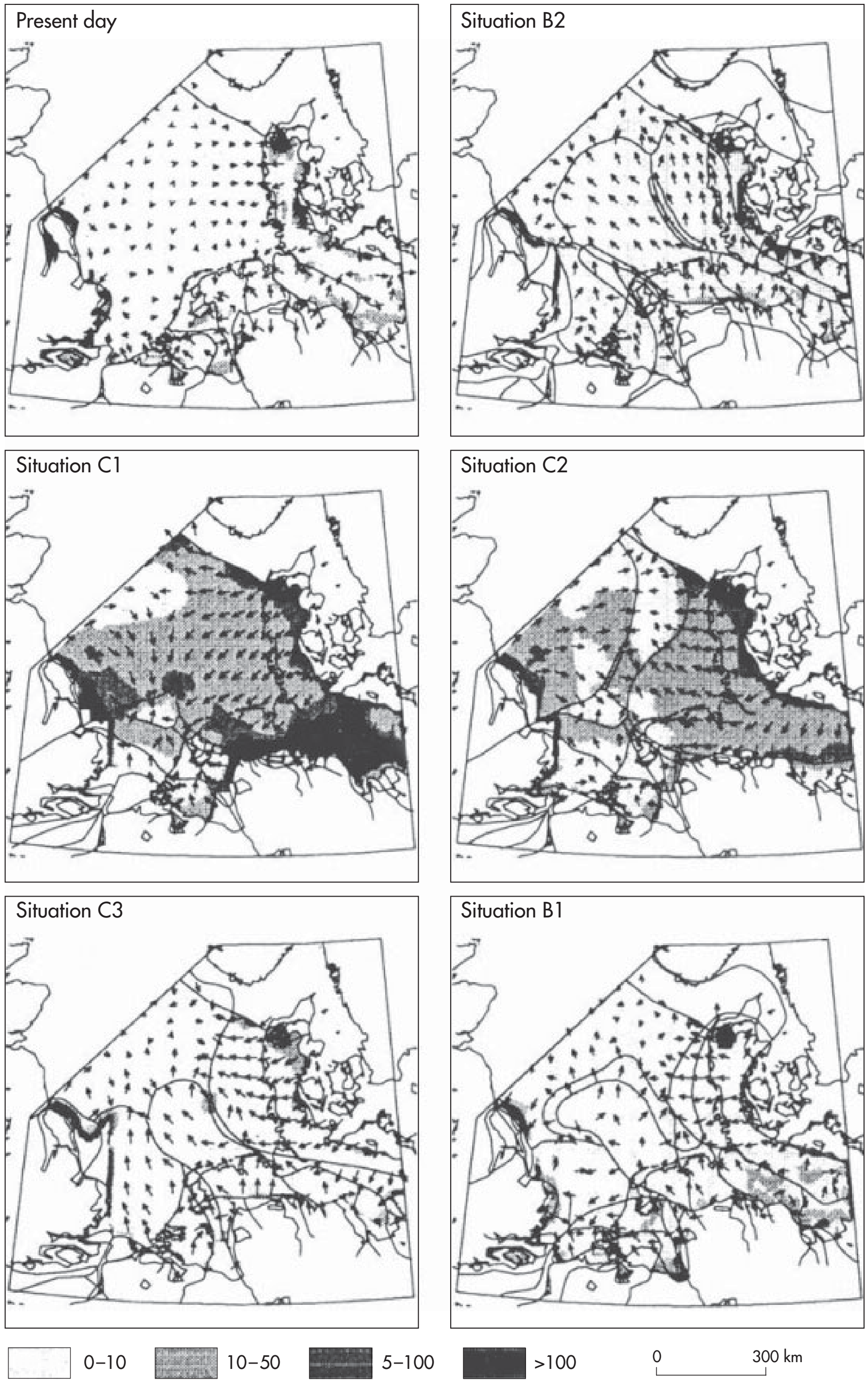


Figure 9.4 Horizontal groundwater velocity vectors and the velocity field (ranges in $m\,yr^{-1}$) in the major regional aquifer in northwest Europe at present and during the expansion of Fennoscandian and Scottish ice sheets (different palaeogeographical scenarios with maximum ice extent in C1). Note the reorganization of flow patterns under glacial conditions as compared with the present time. (From van Weert *et al.*, 1997.)

bed (Siegel, 1991), and the hydrogeology of the Atlantic continental shelf off New England experienced short-lived but dramatic reorganization during the Last Glacial Maximum, including large-scale disturbances in freshwater/seawater equilibrium (Person *et al.*, 2003). Pressurized glacial meltwater penetrated the bedrock in Northwest Territories, Canada, to depths over 1.6 km (Clark *et al.*, 2000). Under some circumstances, however, a glacier may seal the substratum preventing groundwater recharge, as was the case in one Alpine valley covered by the last ice sheet (Beyerle *et al.*, 1998).

Admittedly, numerical models of past groundwater flow are notoriously difficult to validate, and the range of possible solutions vary with the conceivable range of input parameters and boundary conditions, such as the ice sheet thickness and glacier persistence in a certain area, lithospheric response to loading, and spatial variability of hydraulic conductivities. Heuristic studies are promising but scarce (e.g. Fleming & Clark, 2000). A semi-quantitative test of model results may be provided by the isotope composition of deep groundwater, preferably in aquitards where the preservation potential of old water is high. In most cases direct dating with radiocarbon is not possible due to its relatively fast decay, but unstable isotopes with half-lives comparable to glacial cycles may be used (see below).

Indeed, several studies have shown that glacial meltwater flushed through and remained trapped in fine-grained sediments since the last glaciation in lowland areas of North America and Europe. Remenda *et al.* (1994) reported old groundwater in shallow tills and glaciolacustrine deposits from sites spanning a distance of 2000 km along the margin of the Last Glacial Maximum ice sheet in North America, identified by $\delta^{18}O$ values around −25‰, much lower than the modern precipitation of around −13 to −14‰ (see also Stueber & Walter, 1994; Clark *et al.*, 2000). Similarly, Marlin *et al.* (1997) have found old glacial porewater depleted in ^{18}O and ^{2}H in thick clay sediment inside the extent of the last glaciation in northern Germany. In another area in northern Germany at Gorleben, deep, saline water on top of a salt dome was formed in a cold Pleistocene climate, very likely under glacial conditions (Schelkes *et al.*, 1999). Glacial meltwater has also been documented in several places, notably down to depths of at least 500 m, in fractured rocks of the Fennoscandian and Canadian shields (Wallin, 1995; Smellie & Frape, 1997; Tullborg, 1997a; Glynn *et al.*, 1999; Laaksoharju & Rhén, 1999) and in deep Cambrian aquifers in Estonia (Vaikmäe *et al.*, 2001). These case studies show that, under favourable hydrogeological conditions, glacial meltwater injected into low-conductivity rocks under high pressure can remain there for thousands of years after deglaciation, and can be used as a proxy of global climatic changes of the past. In many other places groundwater recharged during cold periods of the Pleistocene also has been documented, but it may not necessarily be water derived from melting of glacier ice (e.g. Zuber *et al.*, 2004).

9.5 Hydrochemical and environmental aspects

In a cycle comprising a glaciation and two bracketing interglacial periods, the predicted changes in groundwater chemistry will be as profound as the changes in groundwater-flow dynamics (Fig. 9.5). Worth noting is isostatic rebound that may cause fracturing and matrix expansion, thereby increasing the rock permeability on a regional scale and enabling migration of new fluids into the substratum after deglaciation (e.g. Weaver *et al.*, 1995). Glacial meltwater recharging the subsurface has certain characteristics that distinguish it from waters derived from other sources and under warm climate conditions, which help its identification and facilitates reconstruction of groundwater circulation under ice sheets (Smellie & Frape, 1997; Glynn *et al.*, 1999). Such features are:

1 Low ^{18}O and ^{2}H contents. Glacial groundwater may have $\delta^{18}O$ values as low as ca. −30‰. Most old deep glacial groundwaters, however, show slightly higher signatures, probably due to mixing with heavier, pre-glacial water. Low $\delta^{18}O$ is also found in minerals precipitated from subglacial meltwater, e.g. in some calcites and iron oxyhydroxides on bedrock surfaces in Scandinavia.

2 High dissolved oxygen content. During ice formation, oxygen is trapped in air bubbles that will later dissolve under the pressure of the overlying ice and enrich the groundwater. Using O_2 concentrations measured by Stauffer *et al.* (1985) in the Greenland ice sheet, Glynn *et al.* (1999) concluded that O_2 content in subglacial meltwater may be three to five times greater than the normal concentration at equilibrium with the

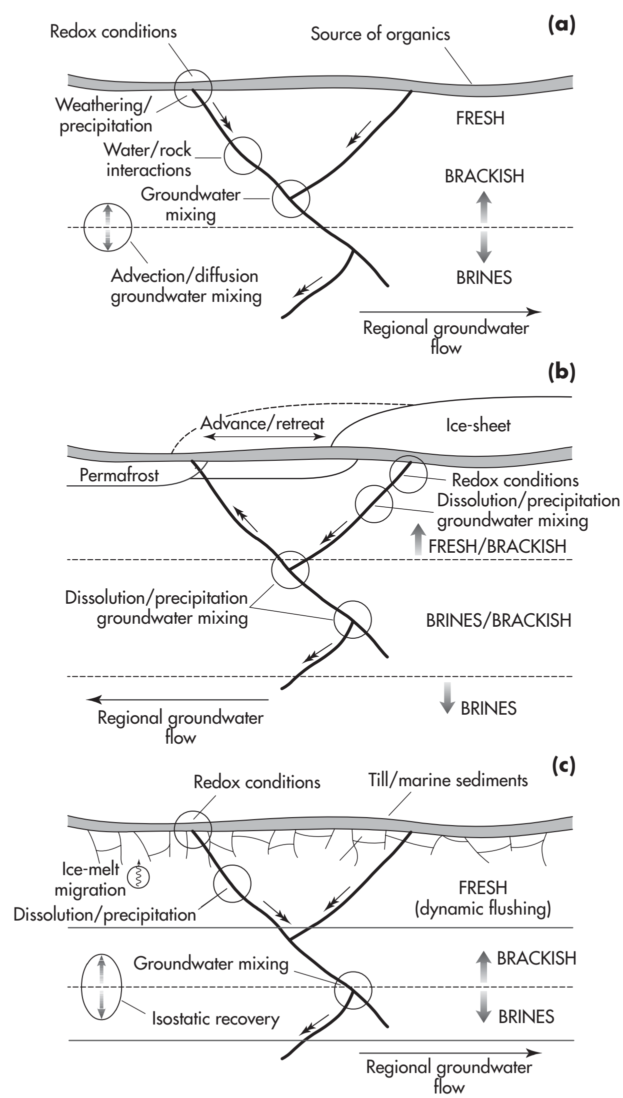


Figure 9.5 Hydrogeochemical conditions related to pre-glacial (a), glacial (b) and post-glacial (c) periods in fractured bedrock areas. Circled are aspects sensitive to surface conditions that may provide signatures of glaciations or interglacials. (Adapted from Smellie & Frape, 1997.)

atmosphere at 0°C. This can lead to the formation of non-hydrothermal iron oxyhydroxides such as those coating fractures in bedrock at a depth of several hundred metres in Sweden (e.g. at 900 m at Äspö in southeast Sweden; Glynn *et al.*, 1999). Up to a depth of 120 m, occurrence of iron oxyhydroxides coincides with the absence of pyrite and calcite (Tullborg, 1997b). Meltwater rich in oxygen would also lead to sulphide oxidation and characteristic $\delta^{13}C$ values (Wallin, 1992).

3 Lack of pyrite along the conductive fractures. Borehole cores from crystalline bedrock around Äspö reveal absence of pyrite in hydraulically active fractures, whereas this mineral occurs in closed fractures and in the rock matrix (Tullborg, 1989; Smellie & Laaksoharju, 1992). This possibly is due to the penetration of oxidizing glacial groundwater during the last glaciation, which did not affect the low-conductivity matrix and fractures outside the active groundwater system.

4 High values of $\delta^{13}C$ in calcites. Values around −5‰ and higher (together with low $\delta^{18}O$ contents) commonly occur in fracture fills in the Fennoscandian shield, e.g. at 600 m depth at Laxemar, possibly indicating glacial meltwater as the precipitant source (Wallin, 1995).

5 Compositional zoning of fracture minerals. Smellie & Frape (1997) pointed out that minerals may reflect more than one glacial cycle as discrete zones, whereas water composition will be a hybrid resulting from mixing, dominated by the most recent events. They give examples of zoned calcites from Sellafield, UK with depleted, low temperature fluid inclusions often interpreted as a glacial signature, dated to around the end of the last glaciation. Also at several sites in Finland and Sweden multiple generations of mineral precipitation were documented, interpreted as reflection of repeated cold and warm climatic events (Smellie & Frape, 1997).

6 Low dissolved organic carbon concentration. Water released from the melting glacier base would typically infiltrate the bed without passing through the organic-rich soil zone, and thus have bicarbonic acid concentrations two orders of magnitude lower than found in interglacial meteoric water. A low ^{14}C content (<12 ppm C) is a strong indication of glacially derived groundwater (Tullborg, 1997a).

7 Low total dissolved solids contents. Studies in Iceland and Spitsbergen show that modern glacial groundwater is characterized by a non-equilibrium (undersaturation) with respect to most constituents otherwise derived from host rocks, indicating efficient flushing, short residence time and low degree of water–rock interactions (Sigurdsson, 1990; Haldorsen *et al.*, 1996). Dissolved solids are expected to rise after ice retreat when the groundwater flow slows significantly.

8 Uranium-series disequilibrium (USD). Penetration of oxic meltwater should affect the U/Th ratio in minerals along groundwater flow passages because uranium is much more mobile than thorium under oxidizing conditions. This can be useful in assessing the role of glacial meltwater in the substratum during the late Pleistocene glaciations. Indeed, USD determined in bedrock in Palmottu, southeastern Finland indicates clear periodicity in uranium mobilization several times during the past 300,000 yr, correlated with major glaciation phases (Suksi *et al.*, 2001; Rasilainen *et al.*, 2003). At Kamlunge, northern Sweden the USD indicating depletion of uranium in the past 500,000 yr was determined to a depth of about 500 m in high-conductivity zones of bedrock (Smellie, 1985).

Remnants of glacial meltwater with these characteristics are found in numerous localities in crystalline shield of Scandinavia and North America as well as in some places in soft sediments south of the crystalline basement. Documentation of such remnants at depths of at least 400 m is convincing. Bearing in mind the likelihood of the next glacial maxima at ca. 5, 20, 60 and 100 ka from now as predicted by the Central Climate Change Scenario (King-Clayton *et al.*, 1997), the old glacial groundwater found at such depths and numerical models suggesting penetration of subglacial water down to several kilometres must be of concern in designing radioactive waste disposal strategies (Talbot, 1999). Repositories are considered, for example, in Sweden and northern Germany, well within the range of prospective glaciations. The major problem is the high dissolved oxygen content of glacial groundwater, which would increase the solubility and mobility of many radionuclides (U, Pu, Tc, Np) by several orders of magnitude and may put the waste canisters stability at risk due to oxidization (Glynn *et al.*, 1999). A special hard rock laboratory on the island of Äspö, southeast Sweden serves as a test and research facility to study the suitability of crystalline bedrock for hosting radioactive waste repositories (SKI, 1997). The laboratory, situated 450 m under the ground surface has served, among others, to evaluate the impact of groundwater flow under a perspective ice sheet on the repository's safety. The majority of the results show that the site would be affected by enhanced, oxidizing groundwater penetration under a future ice sheet. Because the stability of nuclear waste repositories must be ensured for tens to hundreds of thousands of years (e.g. in USA for 10,000 yr, in Switzerland and France for 1,000,000 yr), groundwater stress periods imposed by glaciations must be taken into account.

9.6 Summary

Aquifers overridden by ice sheets and groundwater recharged into permeable beds from the melting ice base are important components of the hydrological cycle in regions affected by glaciations. Research on subglacial groundwater is very scarce and often disputable, but the available data allow the following generalizations:

1 Glaciers pressurize groundwater and impose regional head gradients driving groundwater toward the ice margin where it is released into subaerial drainage systems. If the glacier margin rests on permafrost, the zone of pressurized groundwater may extend for tens of kilometres into the ice forefield.

2 Groundwater systems are significantly if not entirely reorganized during glaciations as compared with the non-glacial conditions. Local topographic catchments are replaced by large, ice-shape controlled regional catchments that extend from glacier termini to the ice divides or to cold-based inner parts of ice sheets. Groundwater flow velocities increase several times, and flow depth may reach several kilometres flushing deep aquifers and aquitards. Non-glacial meteoric groundwater may be replaced completely by glacially fed groundwater.

3 Drainage capacity of the bed influences the ice movement mechanism by controlling the sediment strength and the extent of basal coupling. A substratum of low hydraulic conductivity facilitates fast ice flow by bed deformation and/or basal sliding on a thin water sheet. Indirectly, the drainage capacity of the bed influences glacier transport processes ranging between the subglacial transport in the deforming bed and englacial transport when the deforming bed is absent.

4 Large Pleistocene ice sheets overrode beds whose transmissivity was typically insufficient to evacuate all the subglacial meltwater as groundwater flow. The system responded by formation of channels, the remnants of which, in the form of tunnel valleys (some over 500 m deep and hundreds of kilometres long) and eskers, occur abundantly across Europe and North America.

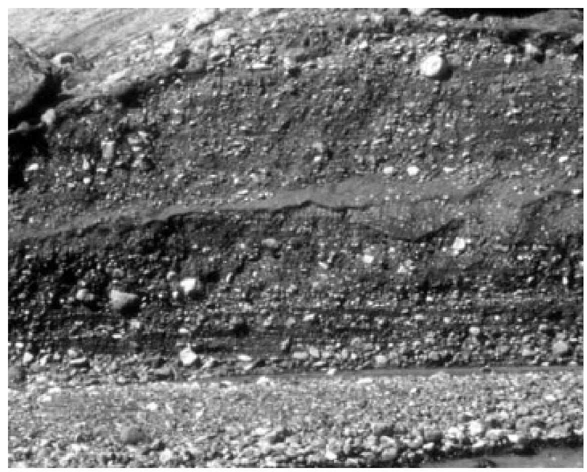
5 Pressurized groundwater may trigger profound disturbances in sediments and rocks such as hydrofracturing and glaciotectonic thrusting and folding, and it may create specific landscapes of soft-sediment extrusion.

6 Glacial groundwater is characterized by a distinct chemical composition, notably low ^{18}O and ^{2}H content and high content of dissolved oxygen, and rocks affected by glacial meltwater may show specific mineralogical and chemical signatures. Old glacial groundwater trapped in aquitards has been documented from several bedrock and sediment areas in Europe and North America, down to depths of several hundreds of metres and more.

7 The predicted reorganization of groundwater flow in areas affected by future glaciations, especially its increased flow dynamics, deep penetration and highly oxidizing chemistry, is of major concern for planning repositories of radioactive waste because many radionuclides have half-lives well outside the next glacial cycle.

TEN

Groundwater flow under the margin of the last Scandinavian ice sheet around the Eckernförde Bay, northwest Germany



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Northwestern Germany experienced several ice advances and retreats during the Last Glacial Maximum. This part of the Scandinavian Ice Sheet, located at the margin of the Baltic Sea basin, was dominated by a land-based but highly dynamic Baltic Ice Stream that terminated about 30 km southwest of the study area at its maximum extent. Abundant tunnel valleys, drumlins, and low preconsolidation ratios of tills suggest fast ice flow caused by some combination of enhanced basal sliding and deformation of soft sediment in this and adjacent areas (Piotrowski & Kraus, 1997; Piotrowski & Tulaczyk, 1999), both indicative of subglacial water pressure elevated to the vicinity of ice flotation point. This study was conducted to evaluate the capacity of the glacier bed