Glaciohydrogeology

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CHAPTER

GLACIOHYDROGEOLOGY

12

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12.1 INTRODUCTION

Hydrogeology is defined as the science of the occurrence, distribution, and movement of water below the Earth's surface. This chapter will focus on groundwater occurrence and movement below or at the periphery of ancient and modern ice sheets (e.g., glaciohydrogeology), but also on the interaction of meltwater with the sediments and rocks within glaciated terrain. The growth of ice sheets has a large-scale and long-term impact on groundwater flow as evidenced by modern groundwater flow which locally did not recover from the Late Pleistocene glacial period (Lemieux et al., 2008b). Over the last 20 years, the interest in glacial hydrogeology from glaciologists, hydrologists, and glacial geologists has significantly increased because of the impact of groundwater flow on ice sheet dynamics, water budgets in glacial catchments, solute fluxes, and tunnel valley genesis (Boulton and Hindmarsh, 1987; Piotrowski, 1997, 2006; Wadham et al., 2001; Cooper et al., 2002; Flowers et al., 2005; Le Brocq et al., 2009). The implications of groundwater flow for the management of potable water resources, the viability of nuclear waste repositories, and the exploitation of hydrocarbon resources have also greatly contributed to the development of this subject (Forsberg, 1996; Talbot, 1999; McIntosh et al., 2002; McIntosh and Walter, 2006; Huuse et al., 2012; Iverson and Person, 2012).

In modern and ancient glacial environments, groundwater hydrology is influenced by numerous parameters including the meltwater production, glacier thermal regimes, the overburden ice pressure, the bed lithology and stratigraphy, and permafrost distribution among others. These different parameters have a strong impact on hydraulic heads, groundwater flow patterns and velocities, penetration depths of meltwater, aquifer recharging rates, and porewater pressure in subglacial to marginal subsurface sediments.

After introducing elementary principles of hydrogeology adapted to glacial terrain, this chapter describes the evolution of groundwater flow characteristics during glaciation. Also examined are the impact of groundwater drainage on ice dynamics, glacial landforms, and the hydrogeochemical evolution of meltwater.

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12.2 ELEMENTARY PRINCIPLES OF HYDROGEOLOGY

12.2.1 GROUNDWATER FLOW

In permeable subglacial material (sediments or rocks), basal meltwater flow will enter the substratum and will be evacuated as groundwater flow. In this configuration, meltwater flow is controlled by the same physical laws describing groundwater flows in confined aquifers outside glacial areas. Meltwater flow is governed by Darcy's Law defining the ability of a fluid to flow through a porous media (Fig. 12.1). Darcy's Law is inversely proportional to the hydraulic gradient (dH/dL) and is a function of water discharge and sediment/rock hydraulic conductivity. The hydraulic gradient measures the potential energy that causes groundwater to flow, while hydraulic conductivity defines the aquifer's ability to transmit water under a hydraulic gradient. Hydraulic gradients are calculated as a measure of change in hydraulic head (elevation of a water body above an arbitrary datum plus the height to which water will rise in a piezometer) over a given distance. The potentiometric or piezometric surface defines a hypothetical surface to which water in a confined aquifer would rise in piezometers (Fig. 12.1). The slope of this potentiometric surface defines the hydraulic gradient and the horizontal direction of groundwater flow (Fig. 12.1). Groundwater flow in a subglacial confined aquifer can be considered as a special case since the hydraulic gradient is largely influenced by the ice-surface slope (Fountain and Walder, 1998). As a consequence, the potentiometric surface of groundwater in a subglacial confined aquifer is nearly parallel to the ice surface (Fig. 12.1). In a

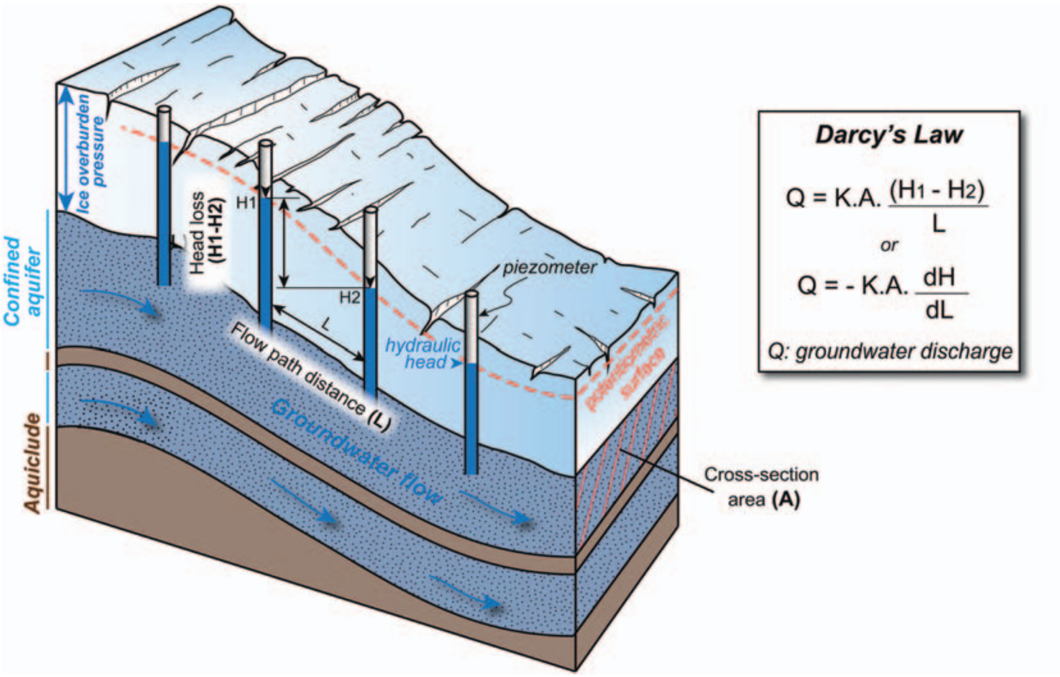


FIGURE 12.1

Illustration of Darcy's Law in an aquifer emplaced beneath an ice sheet. K corresponds to the hydraulic conductivity of the porous medium.

subglacial configuration some groundwater can also be advected within the sediments as it deforms in response to glacier stress (Piotrowski, 2006).

12.2.2 ROCK AND SEDIMENT PROPERTIES AFFECTING GROUNDWATER MOVEMENT

Groundwater can transmit water through a range of geological materials from unconsolidated sediments to sedimentary and crystalline rocks that can either be fractured or not (Fig. 12.2). According to their ability to transmit water, geological formations are subdivided into aquifers, aquicludes, and aquitards. Aquifers are defined as saturated geological formations that contain and transmit significant quantities of water, while aquitards are described as geological formations of lower permeability allowing some water movement through it, but at much reduced rates. Although similar to aquitard, the term aquiclude is commonly used to describe a geological formation that may contain water but does not transmit significant quantities and can act as an impermeable barrier with respect to regional groundwater flow. Aquifers, aquitards, and aquicludes are defined on the basis of porosity, permeability, and hydraulic conductivity of sediments and rocks which are

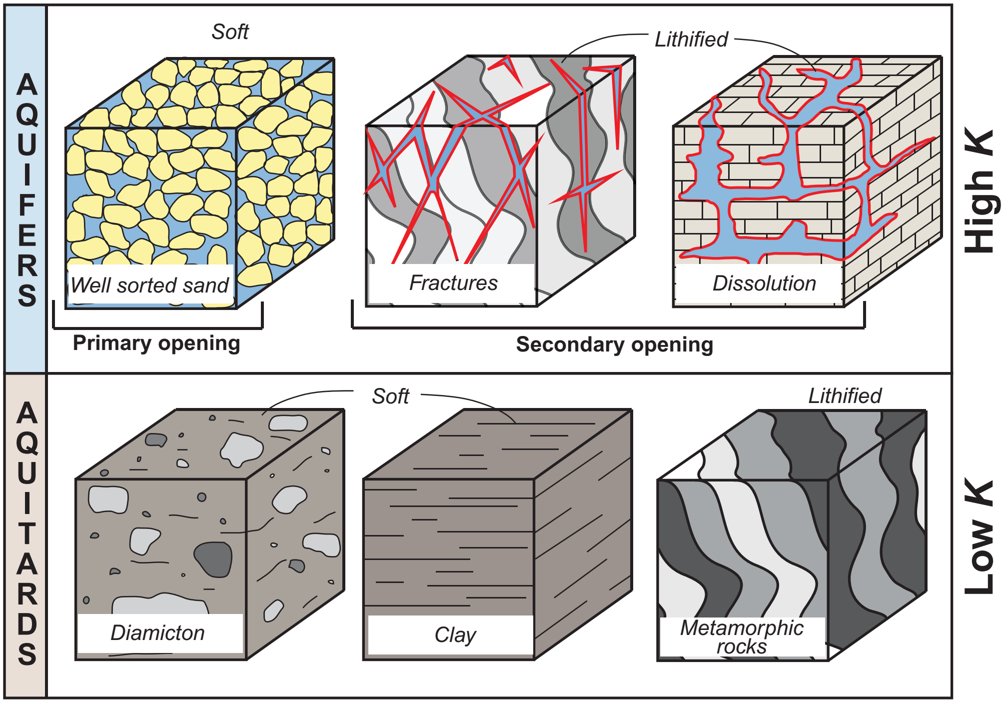


FIGURE 12.2

Examples of aquifers (high hydraulic conductivity (K)) and aquitards (low hydraulic conductivity (K)) encountered in sedimentary basins affected by glacial episodes. The primary porosity and permeability (e.g., primary opening) correspond to the voids, interstices, pores, or pore space contained in a geological material and created by geologic processes. Secondary opening can develop after the rock was formed. Joints or fractures opening, solution openings lead to the development of secondary porosity and permeability.

controlled by grain size, grain sorting, grain shape, clay and organic content, cement percentage, and fractures concentration.

The substratum beneath or at the periphery of ice sheets and glaciers is often composed of stacked sediment or rock layers that either behave as aquifers, aquitards, or aquicludes. When an aquifer is entrapped between two aquitards, the aquifer is said to be confined (Fig. 12.1). Confined aquifers, also known as artesian aquifers, occur where groundwater is confined by overlying impermeable strata or ice under pressures greater than atmospheric. Conversely, a groundwater *aquifer* is said to be *unconfined* when the watertable is open to the atmosphere. Subsurface sediments are generally partitioned into unconfined or confined aquifers controlling large-scale groundwater movement. It is therefore necessary to study the stratigraphic and architectural relationships of layered geological materials (i.e., hydrostratigraphy) to understand and predict groundwater flow patterns (Maxey 1964; Seaber 1982; Atkinson et al., 2014). Aquifers and aquitards are often referred to as hydrostratigraphic units which may be further partitioned into individual hydrofacies. Hydrostratigraphic units are defined as laterally continuous (kilometre-scale) layers of sediments whose range and spatial distribution of hydraulic conductivity contrasts with that of adjacent layers (Atkinson et al., 2014). Hydrostratigraphic units can be further divided into individual hydrofacies characterized by a relatively homogeneous hydraulic conductivity and recognized on the basis of sediment texture (grain size and sorting), fabric, and packing (Klingbeil et al., 1999; Kostic et al., 2005) (see chapter 11: Glacial Lithofacies and Stratigraphy).

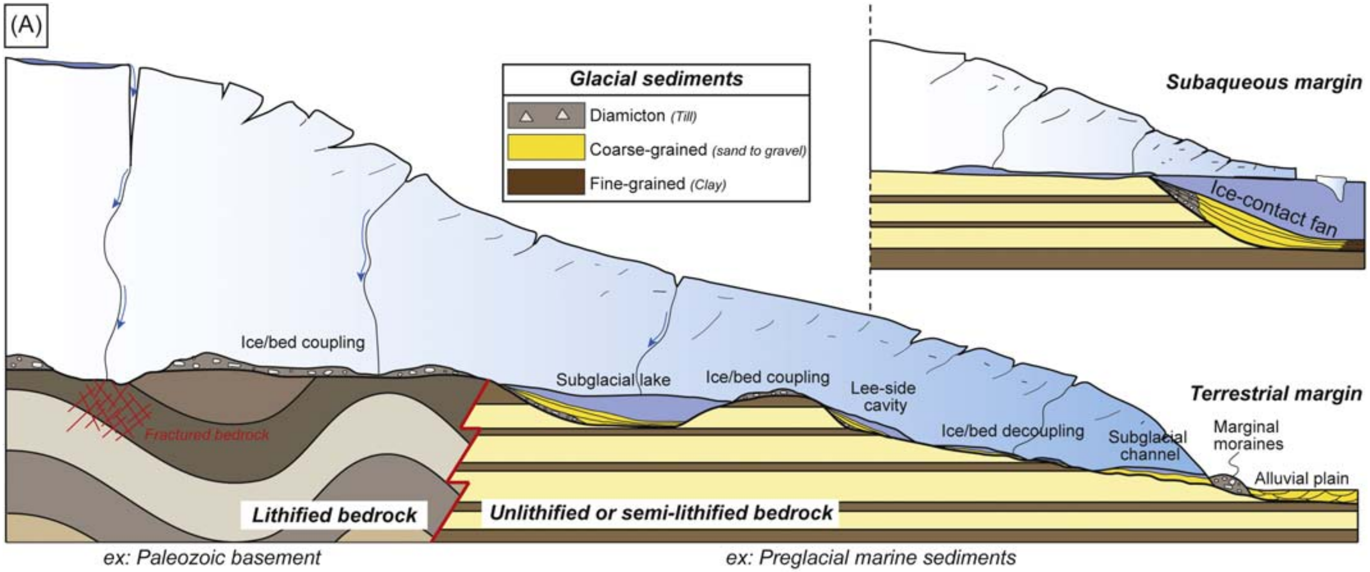
12.2.3 AQUIFERS, AQUITARDS, AND AQUICLUDES ALONG GLACIATED TERRAINS

Ice sheets can overlie many types of geological materials including unconsolidated or consolidated bedrock types. Hydraulic properties of these different materials can be altered by weathering, fracturing, diagenesis, or by permafrost development. Weathering and fracturing processes commonly increase the hydraulic conductivities of bedrock while pore-filling crystals during sediment diagenesis (cementation) and permafrost development tend to significantly decrease the ability of sediments to transmit water. Sediment and rock types have therefore a major influence on groundwater movement beneath ice sheets.

12.2.3.1 Unconsolidated glacial sediments

Ice sheets may develop on unconsolidated material of glacial or nonglacial origin displaying primary porosity inferred from depositional processes (Fig. 12.2). The uppermost portion of glaciated terrain is usually composed of glaciogenic deposits displaying a wide range of hydraulic properties (Freeze and Cherry, 1979; Stephenson et al., 1988) (Fig. 12.3A and B). In subglacial to marginal environments, deposition of unconsolidated material occurs in many settings including channels, cavities, proglacial lakes, alluvial plains, with various depositional processes leading to a variety of lithologies, grain size, and grain sorting (Domenico and Schwartz, 1998; Heinz and Aigner, 2003) (Fig. 12.3A). Indeed, the range of depositional environments associated with the different glacial landsystems leads to a complex and heterogeneous distribution of hydraulic conductivity values (three orders of magnitudes) beneath and at the periphery of ice sheets (Robinson et al., 2008).

Most aquifers in glaciated regions are composed of glaciofluvial sands and gravels occurring as blanket bodies or as channel deposits in alluvial plains and buried bedrock valleys (Freeze and Cherry, 1979; Maizels, 1993; Domenico and Schwartz, 1998). Tills are also a major component of glacial terrain and can either behave as aquifers or aquicludes since they display highly variable



(B)

Unconsolidated sediments

Types	*Conductivity* (m.s^{-1})
Clay	1.10^{-11} to $4.7.10^{-9}$
Silt	1.10^{-9} to 2.10^{-5}
Fine sand	2.10^{-7} to 2.10^{-4}
Medium sand	9.10^{-7} to 5.10^{-4}
Coarse sand	9.10^{-7} to 6.10^{-3}
Gravel	3.10^{-4} to 3.10^{-2}
Till	8.10^{-12} to 2.10^{-2}

Sedimentary rocks & crystalline rocks

Types	*Conductivity* (m.s^{-1})
Limestones	1.10^{-9} to 6.10^{-6}
Karst and Reef limestones	1.10^{-6} to 2.10^{-2}
Shale	1.10^{-13} to 2.10^{-9}
Siltstone	1.10^{-11} to $1.4.10^{-8}$
Sandstone	3.10^{-10} to 6.10^{-6}
Salt	1.10^{-12} to 1.10^{-10}
Basalt	2.10^{-11} to $4.2.10^{-7}$
Unfractured igneous/metamorphic rocks	3.10^{-14} to 2.10^{-10}
Fractured igneous/metemorphic rocks	8.10^{-9} to 3.10^{-4}

Conductivity values from Domenico and Schwartz, 1998

FIGURE 12.3

(A) Conceptual models showing different types of substratum which either predate glaciations (e.g., lithified bedrock; preglacial marine semilithified sediments) or which are deposited during glaciation beneath or at the margin of ice sheets. (B) The hydraulic conductivities of some representative unconsolidated and consolidated geological material encountered along formerly glaciated terrains are compiled in two tables.

hydraulic conductivities ($2 \cdot 10^{-2}$–$8 \cdot 10^{-12}$ m $\cdot$ s^{-1}). Estimating hydrological parameters of subglacial tills is therefore fundamental in the description of groundwater flow during glacial episodes as it often represents the first layer in which meltwater penetrates. In subglacial environments, deformational processes can modify till hydrogeologic parameters (Alley et al., 1987; Domenico and Schwartz, 1998; Engelhardt et al., 1990; Piotrowski, 2006) (Fig. 12.3). Deformation can increase till conductivity by dilation, fracturing, faulting, and shear-banding, while stress imposed by glacier weight can decrease till conductivity by compaction (Hubbard and Maltman, 2000). Dessication or freeze–thaw processes can also lead to till fracturing (Klint and Tsakiroglou, 2000).

Ice sheets can also overlie soft sediments of nonglacial origin (Fig. 12.3). In the North Sea Basin, Pleistocene ice sheets developed over Cenozoic shallow marine to deltaic deposits (Janszen et al., 2012). Similarly, the Gondwana Ice Sheet developed over unlithified preglacial marine sediments in North Africa during the Late Ordovician glaciation (Loi et al., 2010; Clerc et al., 2013). For preglacial successions, hydrogeologic properties of sediments are also a function of grain size, grain shape, and grain sorting, and therefore linked to the depositional processes and environments. Indeed, studies on Ordovician sedimentary deposits in Morocco showed that sandstone deposited in shoreface environments behaved as aquifers during the Late Ordovician glaciation, while claystones deposited in offshore environments behaved as aquicludes, therefore influencing groundwater movement beneath the palaeo-ice sheet (Ravier et al., 2014).

12.2.3.2 Lithified substratum

Glaciers may also cover sedimentary bedrock of varying hydraulic characteristics. Sedimentary rocks can display a wide range of hydraulic conductivity values according to their diagenetic histories. Indeed, during burial and exhumation sedimentary rocks can experience cementation and/or dissolution phases. Cementation phases typically decrease the hydraulic conductivities of rocks due to pore-infilling by crystals. Conversely, the circulation of fluid in lithified sedimentary rocks can dissolve crystals and therefore increase permeability values (Fig. 12.2). For example, meltwater circulation into carbonate or evaporate may increase the hydraulic conductivities of the subglacial substratum through dissolution and development of karst-like features beneath the ice (McIntosh and Walter, 2006). Cemented sedimentary rocks, igneous, and metamorphic rocks have typically little hydraulic conductivities that can be later modified by fracturing during active tectonic phases, glaciations (i.e., ice loading), and postglacial rebound.

12.2.3.3 Impacts of glaciations on rock and sediment properties

During glaciation, sediments and rocks can experience modifications altering their original hydraulic properties caused by porewater freezing (i.e., permafrost development) and/or fracturing. Permafrost development drastically reduces the hydraulic conductivity of geological materials (French, 1996; Lemieux et al., 2008b; Stothoff, 2012). A permafrost zone acts as an aquiclude, with the frozen medium exhibiting effective hydraulic conductivity several orders of magnitude lower than the unfrozen medium.

Bedrock with little primary porosity will form secondary porosity as the rock is fractured due to shear and normal stresses respectively imposed by ice flow and ice weight (Passchier et al., 1998; McCarroll and Rijsdijk 2003) (Fig. 12.2). Ice sheets can effect rapid changes in mechanical stress in the subsurface, leading to deformation and fracturing of sediments and rocks, thus altering fluid transport properties of the subglacial substratum (King-Clayton et al., 1997; Neuzil, 2012).

If fracturing leads to increasing hydraulic conductivities in the subglacial bed, sediment or rock compaction induced by overburden ice pressure can also reduce the hydraulic conductivity and increase pore pressure (Neuzil, 2003; Lemieux et al., 2008a; Walsh and Avis, 2010).

12.3 MELTWATER PRODUCTION AND DRAINAGE SYSTEMS TO THE BED

During glacial periods, meltwater is likely to have been the only source of groundwater recharge in high-latitude areas covered by ice sheets. Ice sheets melt on their surfaces (the supraglacial zone) and at their bases (the subglacial zone). The subglacial meltwater production is linked to friction due to ice flow, ice melt by water flow, and ice melt due to geothermal heat from the Earth's crust. Beneath the Antarctica ice sheet, the rate of subglacial meltwater production is about 3.5 mm a^{-1}, representing 1% of the total mass balance (Llubes et al., 2006). Supraglacial meltwater production represents the main source of meltwater production in most glacial systems. Supraglacial meltwater production rates vary according to the seasons, climate changes, and local weather conditions and result from snow, firn, and ice melting. Supraglacial melting rates are notably controlled by the amount of solar radiation and albedo changes at the surface of the glacier. Supraglacial meltwater drains through moulins toward the base of the glacier following englacial networks composed of fractures and crevasses (Fig. 12.4C). Supraglacial meltwater transfer to the subglacial network is controlled by the permeability of snow, firn, and ice, and by ice mass velocity, the presence and geometry of moulins and the basal subglacial topography. The high permeabilities of snow and firn allow meltwater to be infiltrated into the englacial system. At the same time, with a permeability of 10^{-18} m^2, ice can be considered almost impermeable, except where fractures and crevasses locally occur. Above the equilibrium line altitude (ELA), a deep layer of firn can store huge volumes of meltwater and constitute a meltwater reservoir during winter (Forster et al., 2014). Each summer, melting of snow and glacier ice leads to the development of a diffuse runoff at the ice surface through arborescent supraglacial stream networks (Fig. 12.4A and C). Supraglacial channelized stream network patterns are controlled by crevasse orientations and glacier surface slope. Meltwater can be temporally stored in depressions at the glacier surface forming supraglacial lakes (Fig. 12.4A and C). These lakes typically form annually and are generated in high ablation rate areas. On the Greenland Ice Sheet, small lakes form at about 1000 m above sea level (a.s.l.) and larger lakes can develop at higher altitudes later during the melting season (Fitzpatrick et al., 2014). Supraglacial lakes store a huge volume of meltwater and play a crucial role on their transfer to subglacial hydraulic networks using the englacial network. When supraglacial incision by stream channels is faster than surface ablation in uncrevassed areas, englacial conduits can form by incision and closure of supraglacial streams, this model of formation is defined as the 'cut and closure' model (Gulley et al., 2009). This method of meltwater transfer is limited and meltwater will be generally quickly transferred through englacial fractures, often propagating by hydrofracturing in areas of accelerating ice flow, or slowly transferred using narrow englacial conduits or 'moulins' (Fountain et al., 2005; Das et al., 2008; Benn et al., 2009) (Fig. 12.4B). The exploration of these englacial conduits suggests karstic style morphologies developed along a fracture network in crevassed areas. Studies of 'moulins' on the Greenland Ice Sheet revealed their persistence for years, thus developing an efficient way to deliver large volumes of supraglacial meltwater to the subglacial network (Gulley et al., 2009;

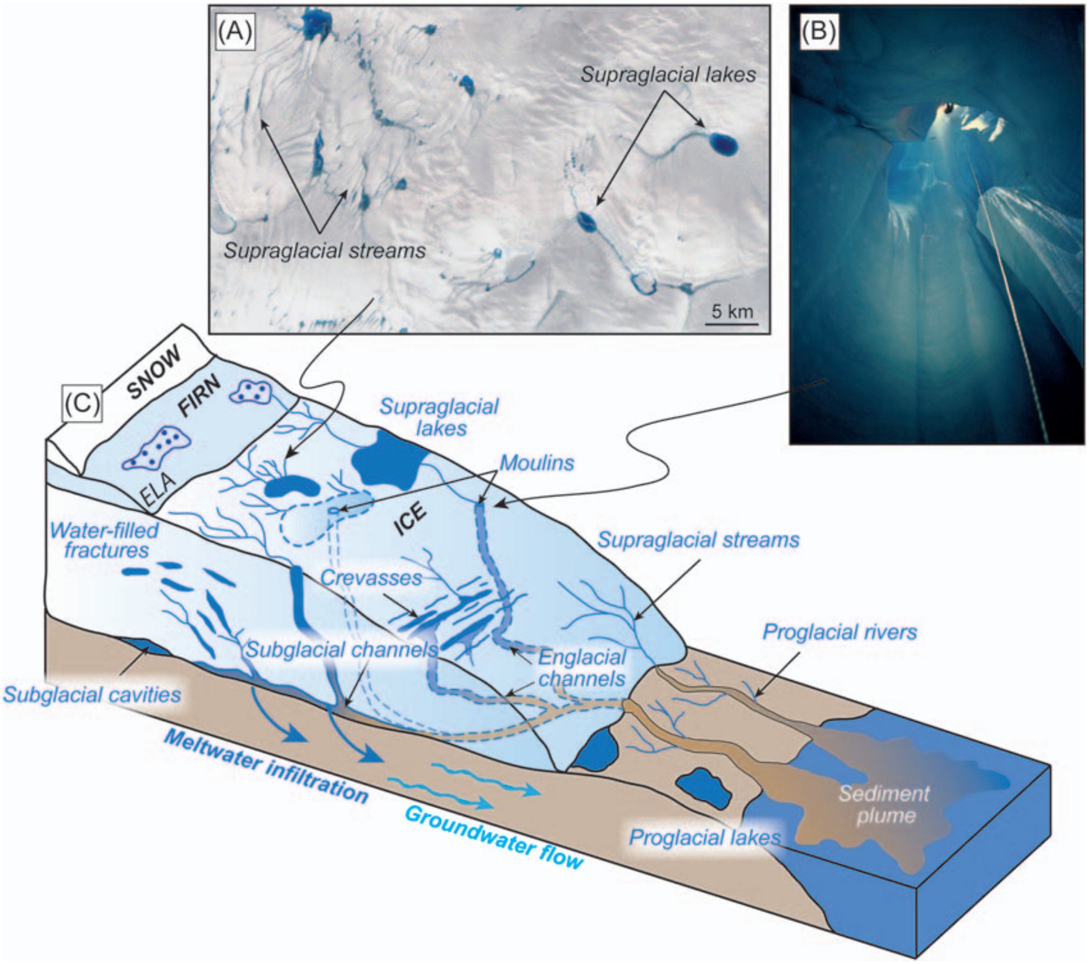


FIGURE 12.4

(A) Aerial view of the ablation area on the surface of Greenland showing supraglacial stream networks and supraglacial lakes 1 NASA. (B) Exploration of a moulin in the east of the Greenland 1 Luc Moreau. (C) Elements of an ice sheet hydrologic system. In the accumulation zone above the equilibrium line altitude (ELA), water percolating through the snow/firn can pool into slush regions and channelize into supraglacial streams. In the ablation zone beneath the ELA, meltwater pools in supraglacial lakes and flows through streams into crevasses and moulins, entering englacial networks, and supplying the subglacial hydrological system.

Modified from Cuffey, K., Paterson, W.S.B., 2010. The Physics of Glaciers, fourth ed. Elsevier Inc., Burlington and Oxford (Cuffey and Paterson, 2010).

Catania and Neumann, 2010). Injections of meltwater from supraglacial lakes or stream channels via crevasses or moulins may trigger water-driven hydrofracturing and transport of large volumes of supraglacial meltwater toward the base of an ice sheet (van der Veen 2007; Das et al., 2008; Benn et al., 2009). Once meltwater reaches the bed, water can be discharged by three main mechanisms:

flow in a thin layer (film) at the ice–bed interface, flow in tunnels, channels, and conduits, and flow as groundwater within sediment and/or bedrock.

12.4 GROUNDWATER FLOW CHARACTERISTICS UNDER ICE SHEETS

Glaciations represent large-scale global climate changes drastically impacting groundwater flow systems and potentially affecting basin-scale groundwater cycles (Piotrowski, 2006). Under nonglacial conditions, groundwater is recharged in topographically high areas and discharged from topographic lows. Glacial periods are responsible for the reorganization of relatively small topographically controlled groundwater catchments to produce continental-scale integration of pressurized groundwater flow recharged by meltwater (Boulton and Caban, 1995). To investigate the characteristics of groundwater flow during glaciations, the absence of modern analogues or the limited access to the subglacial bed of the Antarctica or Greenland ice sheets have necessitated the use of numerical models and the geochemical analyses of palaeo-water (Boulton and Caban, 1995; Glynn et al., 1997; Piotrowski, 1997; Person et al., 2007; McIntosh and Walter, 2005; Lemieux et al., 2008b). Some field measurements have also been conducted in Iceland, United States, Switzerland, and Canada using boreholes, piezometers, and electric down-hole pressure transducers (Fountain, 1994; Hubbard et al., 1995; Boulton and Hindmarsh, 1987; Boulton et al., 2001, 2007a; Robinson et al., 2008). Meltwater flow occurrence in the subsurface has also been detected for ancient glaciations using the sedimentary record and the analyses of soft-sediment deformation structures (Meer et al., 1999; Denis et al., 2010; Ravier et al., 2014, 2015). Indeed, meltwater flow within subglacial confined aquifers can be responsible for the local increase in porewater pressure, thus leading to different sediment deformation mechanisms including liquefaction, fluidization, and hydrofracturing (Maltman, 1994). Because of the complexity and number of processes involved during glacial periods, groundwater flow models have become the most powerful tools in characterizing groundwater flow and predicting flow pattern development through glacial periods (van Weert et al., 1997; Breemer et al., 2002; Person et al., 2003; Hoaglund et al., 2004; Carlson et al., 2007).

12.4.1 MELTWATER TRANSFER INTO THE SUBGLACIAL BED

When ice sheets overlie permeable materials, a part of the subglacial meltwater infiltrates the bed and circulates as groundwater flow beneath and at the periphery of ice sheets. Many studies suggest that significant subglacial meltwater may have infiltrated into the subsurface under the ice pressure and therefore becomes stored in the groundwater flow system (Lemieux et al., 2008a). Meltwater infiltration and storage into the ground during glacial periods have been demonstrated using the geochemical signature of meltwater produced during glacial episodes (Section 12.8) (Edmunds, 2001; McIntosh and Walter, 2006; Person et al., 2007). It has been argued that permeable beds beneath an ice sheet can transmit all or various percentages of the meltwater produced depending on the bed lithology and stratigraphy. Boulton et al. (1995) and Boulton and Dobbie (1993) argued that during Pleistocene glaciations, some aquifers in Holland or Scandinavia would have been able to transmit all the basal meltwater by groundwater flow alone as the ice sheet advanced across sedimentary basins composed of high-transmissivity sedimentary layers. However, most models agree on subglacial bed hydraulic inefficiency to drain all meltwaters entering the system and demonstrate that excess meltwater circulates at the ice–bed interface (Breemer et al., 2002; Moeller et al., 2007). Indeed, Flowers et al. (2003) estimate that the groundwater system beneath western Vatnajökull (Iceland) evacuates up to 30% of glacier meltwater discharge annually, while Piotrowski (1997) suggested that during the late Pleistocene aquifers with low transmissivity in northwestern Germany could only have evacuated about 25% of the produced meltwater during the last glaciation. Lemieux et al. (2008a) suggested that an average of about 40% of basal meltwater generated beneath the Laurentide Ice Sheet (LIS) recharged sedimentary and crystalline rocks across North America during the Last Glacial Maximum. Using the salinity value and the isotopic signature of groundwater in North America (Section 12.8), it has been suggested that only 1% of the total water volume of the LIS has been emplaced in confined aquifers overrun by the LIS whereas numerical models conducted in the same areas display higher percentages (5–10% of meltwater generated) (Person et al., 2007). The differences between these two estimations are attributed to differences in the assumptions on which they are based. In their geochemical study, Person et al. (2007) only consider meltwater recharge at the margins of the LIS for a relatively short period of time (Late Wisconsinan) while the numerical continental-scale hydrologic model performed by Lemieux et al. (2006) model results are based on integrated recharge across all of Canada over the whole Wisconsinan period. Both estimations agree on the relatively small percentage of meltwater transferred as groundwater flow during glaciations implying that, in many cases, meltwater excess is mainly evacuated as films or within conduits at the ice–bed interface but also through processes of hydrofracturing (Shoemaker, 1986; Boulton and Caban, 1995; Ng, 2000; Meer et al., 2009) (Sections 12.6 and 12.7).

12.4.2 PARAMETERS CONTROLLING GROUNDWATER FLOW DURING GLACIATION

During glacial cycles, the land surface and the subsurface experience changing conditions as the result of ice sheet loading, increasing meltwater production rates, ice sheet growth and decay, permafrost development, and glacial isostatic adjustments. All these changing conditions, combined with the hydrogeologic characteristics of the glaciogenic sediments and preexisting bedrock control regional groundwater flow during cold climate periods (Provost et al., 1998; Piotrowski, 2006; McIntosh et al., 2011).

12.4.2.1 Glacial mechanical loading

The mechanical ice loading is responsible for major modifications of hydraulic properties of the subsurface sediments and bedrock. Depending on the elastic properties of these materials, which are a function of lithology, grain-size and grain-size distribution, compaction can reduce the hydraulic conductivity by compression of pores and closure of fractures (Neuzil, 2003; Lemieux et al., 2008a). Pore closures are often associated with meltwater expulsion from the sediments. Conversely, fracturing induced by mechanical ice loading may increase the hydraulic conductivities of confining units, thereby providing a pathway for groundwater (Person et al., 2007). In addition, the deformation of the crust by the weight of the ice sheet is such that the surface elevation will be depressed underneath the ice sheet and raised beyond its margins, therefore modifying the characteristics and topography of the subglacial to marginal catchment area (King-Clayton et al., 1997; Neuzil, 2012). Indeed, this topography inherited from ice sheet growth phases impacts large-scale groundwater flow patterns because hydraulic potentials will be lowered below the ice sheet and increased in its forebulge (Ates et al., 1997; Lemieux et al., 2008a). In subglacial environments, groundwater is pressurized by the weight of the overlying ice and sediments, thus influencing hydraulic head values and characteristics of the potentiometric surfaces (cf. Section 12.4.3.1.).

12.4.2.2 Thermal structures of glaciers

The thermal structures of glaciers, whether temperate, cold, or polythermal is an indicator of meltwater production rates and therefore influence groundwater recharge, discharge, and flow characteristics. Glacier thermal structures are mainly controlled by climatic conditions (temperatures, snowfall) and the geothermal heat flux. Beneath polythermal glaciers (composed of both cold and warm basal ice), groundwater recharge is mainly restricted to the temperate basal zones of glaciers (Haldorsen and Heim, 1999). Indeed, the presence of warm-based ice with subsequent basal melting creates a potential for groundwater recharge rates much larger than under cold-based conditions or present, ice-free conditions (Provost et al., 1998). Under cold-based ice, no or very few meltwater is produced at the glacier bed, thus reducing considerably infiltration and decreasing or inhibiting groundwater recharge and discharge.

12.4.2.3 Ice sheet growth/decay

Numerical models show that phases of ice sheet growth or decay influence the groundwater flow characteristics because of fluctuating rates of meltwater infiltration (McIntosh and Walter, 2006; Vidstrand et al., 2010). Indeed, groundwater flow simulations beneath the LIS in Canada show that subglacial meltwater infiltration into the subsurface dominates during periods of ice sheet growth because meltwater is forced downward into the subsurface due to high pressures recorded at the ice–bed interface. During ice sheet retreat, groundwater predominantly exfiltrates because the weight of the ice overlying a given surface location decreases with time and subsurface pressure becomes higher than basal meltwater pressure. This pressure differential causes meltwater stored in the subsurface to discharge at the base of the ice despite the fact that basal melting continues (Lemieux et al., 2008a). Variations in the position of the ice front through episodes of ice sheet growth and decay influence the distribution of stresses exerted by the ice on the bed and therefore influence the sense and direction of groundwater migration. In addition, the migration of the ice margin through time influences the subglacial large-scale topography by redefining the position of the subglacial isostatic depression relative to the marginal forebulge, thus influencing groundwater flow movements (Ates et al., 1997).

12.4.2.4 Permafrost development

Permafrost can be compared to an aquiclude that develops at the land surface, thus affecting groundwater dynamics, groundwater recharge/discharge, and porewater pressure (Burt and Williams, 1976; Kleinberg and Griffin, 2005; Bense and Person, 2008; Vidstrand et al., 2010). The permafrost influence on groundwater hydrology has been evidenced by numerical models and also by geochemical analyses of palaeowaters. Indeed, a hiatus in the geochemical and chronological record of palaeowaters from several places in Europe and North America demonstrates the drastic reduction of the hydraulic conductivity during permafrost development (McEwen and de Marsily, 1991; Edmunds et al., 2001; Lemieux et al., 2008a; McIntosh et al., 2012). The presence of subglacial to marginal permafrost affects the potentiometric surface and therefore influences groundwater flow patterns, meltwater infiltration depths and length of meltwater flow paths (King-Clayton et al., 1997) (Fig. 12.5). Continuous

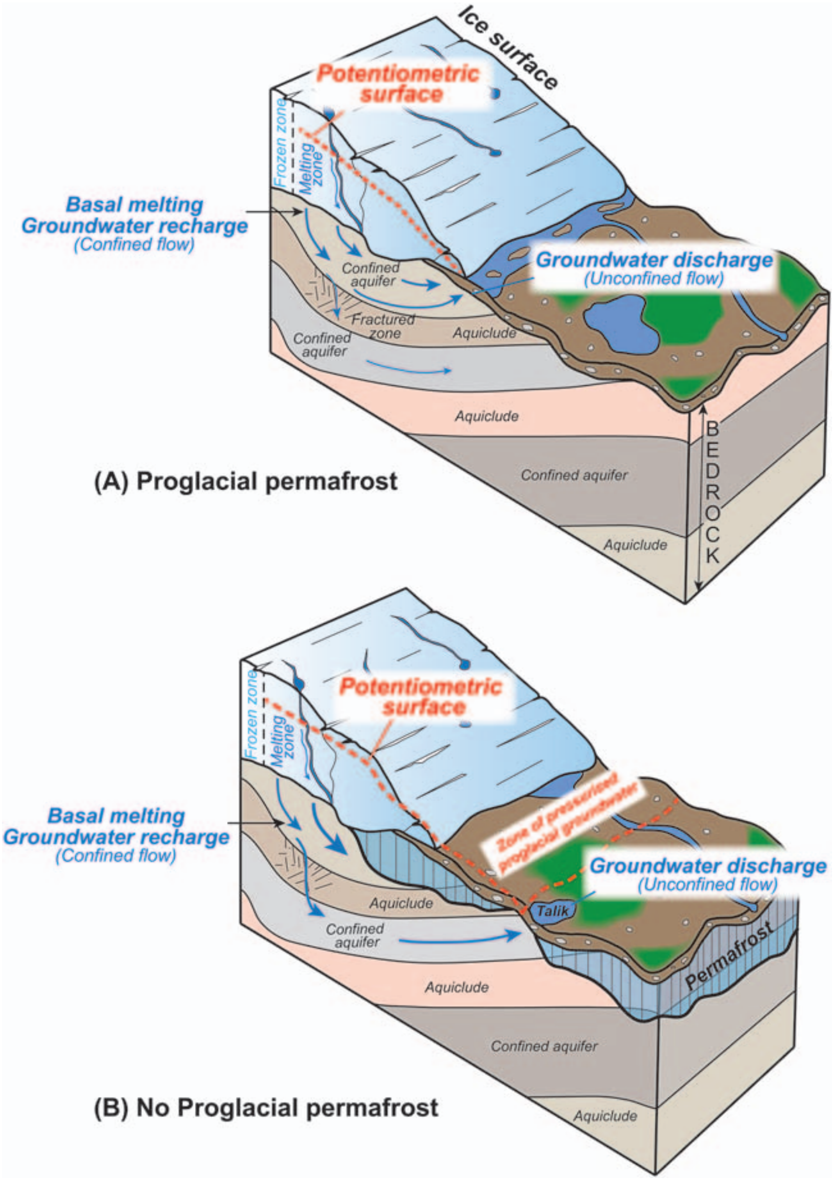


FIGURE 12.5

The occurrence of submarginal to marginal permafrost modifies significantly the potentiometric surface which defines the hydraulic gradient. Groundwater discharge occurs predominantly close to the margin of the ice when no permafrost cover occurs (A). Groundwater discharge occurs away from the margin through taliks if there is a periglacial permafrost (B). Blue arrows represent groundwater flow paths.

permafrost tends to block recharge and discharge of groundwater at large scales, thus influencing regional groundwater flow systems. Discontinuous permafrost has a more complex impact on groundwater flow as it tends to subdivide the regional flow system in local flow systems controlled by water recharge and discharge through taliks ('holes' in permafrost) (Fig. 12.5).

12.4.2.5 Substratum types and stratigraphy

Meltwater at the base of ice sheets may be discharged at the ice–bed interface or by groundwater flow; the balance between the two being controlled by the hydraulic conductivity of subglacial beds (Sections 12.2.2 and 12.2.3). Groundwater flow velocities are mainly controlled by the substratum hydraulic conductivity while groundwater flow paths in the subsurface are controlled by the distribution, the stratigraphic relationships, the lateral extent, and the geometry of aquifers and aquitards beneath an ice sheet (Flowers and Clarke, 2002).

12.4.2.6 Meltwater channels

The occurrence and distribution of channels at the ice–bed interface where meltwater circulates influence groundwater flow. 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 and Leung (1987) and Boulton and Hindmarsh (1987), and recently measured in piezometers set up around a channel under Breiðamerkurjökull in Iceland (Boulton et al., 2001, 2007a, 2009). Evidence from borehole heads and dye tracing in Iceland suggests that Darcian groundwater flow is the predominant mean by which tunnels scavenge meltwater from the glacier sole (Boulton et al., 2007a).

12.4.3 IMPACTS OF GLACIATION ON GROUNDWATER FLOW CHARACTERISTICS

The main impact of ice sheet development on groundwater flow is to reorganize groundwater flow systems from catchment-scale topographically driven recharge to continental-scale glacially driven recharge (Boulton et al., 1995) (Fig. 12.6). Groundwater flow during glacial periods has a strong impact on hydraulic heads, groundwater flow patterns and velocities, recharging rates, penetration depths of meltwater, and porewater pressures.

12.4.3.1 Hydraulic heads and potentiometric surfaces

During glacial periods, groundwater flow is driven by hydraulic heads diminishing toward the ice margin. Hydraulic head values are significantly higher beneath wet-based ice sheets compared to nonglacial conditions. This increase is attributed to the infiltration of subglacial meltwater into the rocks and to the load exerted by the ice sheet on the bed (Piotrowski, 2006). For an average ice sheet thickness of 1000 m, observed water levels in boreholes from a West Antarctic Ice Stream are typically 100 m below the ice surface (Engelhardt and Kamb, 1997). It can be hypothesized that changes in hydraulic head values and potentiometric surfaces completely overwhelmed existing land topography and could have contributed to flow reversal during glaciation (cf. Section 12.4.3.2) (Boulton et al., 1993; Person et al., 2003, 2007). Numerical modelling conducted on Pleistocene ice sheets in Canada displayed an increase in the hydraulic head values below the ice sheet by as much as 3000 m, down to a depth of 1.5 km into the subsurface (Lemieux et al., 2008b). Simulations conducted under the southern margin of the Scandinavian ice sheet in northwest Germany demonstrate

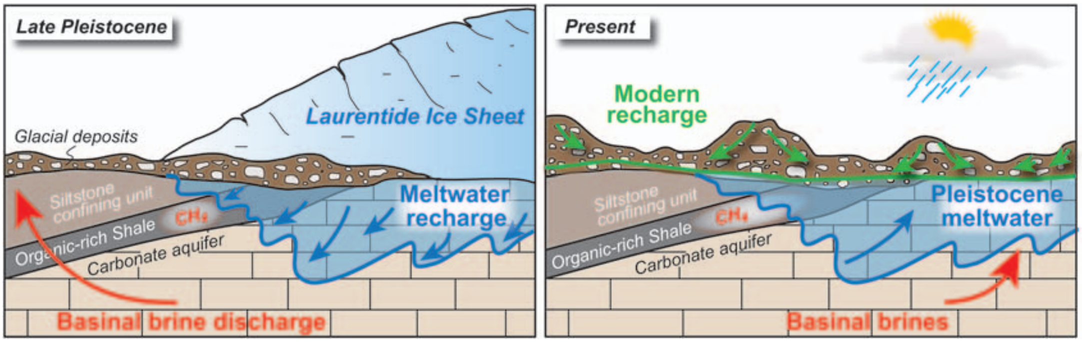


FIGURE 12.6

Conceptual model of groundwater hydrology in the Michigan basin comparing the glacial Late Pleistocene episode and the modern conditions. Meltwater recharge is topographically driven in modern conditions while influenced by ice load during the glacial episodes. Note the reverse direction of groundwater flow between modern and glacial times. Glaciation enhances fresh water recharge, thus diluting and flushing basinal brines and stimulating the generation of natural gas (CH_4).

Modified after McIntosh, J.C., Garven, G., Hanor, J.S., 2011. Impacts of Pleistocene glaciation on large-scale groundwater flow and salinity in the Michigan Basin. Geofluids 11 (1), 18–33.

the impact of glaciations on hydraulic heads since values were between 85 and 185 m under glacier coverage (for a calculated ice profile reaching up 250 m) during the Pleistocene glaciation, whereas at present, they are estimated to be between 2 and 14 m (Marczinek and Piotrowski, 2006).

Permafrost development also has a strong impact on hydraulic head values and distribution. Indeed, permafrost has a significant impact on the rate of dissipation of high hydraulic heads that build at depth as a consequence of subglacial infiltration and ice loading (Bense and Person, 2008; Lemieux et al., 2008b) (Fig. 12.5).

12.4.3.2 Groundwater flow pattern

Groundwater flow paths and directions are highly modified during glaciations. Groundwater flow paths are predominantly controlled by the lateral distribution of ice overburden pressure values (Hoaglund et al., 2004). Indeed, the subglacial flow pattern is characterized by a net downward flow vector in response to vertical ice loading and a strong upward flow beyond the ice sheet due to decreasing overburden pressure (Boulton et al., 1993). During glacial retreat, the weight of the overlying ice decreases, causing a significant pressure release in the subglacial bed. Results from a regional-scale groundwater flow model of the Fennoscandian shield show that this pressure release triggers a flow direction inversion illustrated by upward flow vectors of groundwater, despite the continued potential for recharge of basal meltwater (Provost et al., 1998).

According to several groundwater flow models developed for formerly glaciated areas, pressure exerted by an overlying ice sheet can be sufficient to reverse hydraulic gradients and regional flow within subglacial aquifers (Boulton et al., 1993; Glynn and Voss, 1996; Piotrowski, 1997; van Weert et al., 1997; McIntosh et al., 2011). Examples taken from simulations underneath Pleistocene ice sheets in Europe and North America demonstrate this flow reversal (Grasby et al., 2000; Breemer et al., 2002; Bekele et al., 2003; Carlson et al., 2007). The development of permafrost beneath or at the periphery of ice sheets is also responsible for considerable modifications of the groundwater flow paths as it modifies the potentiometric surface (King-Clayton et al., 1997; Provost et al., 1998) (Fig. 12.5).

12.4.3.3 Groundwater flow velocities

Hydrologic models of subglacial Pleistocene groundwater flow beneath the LIS indicate that flow velocities within aquifers differed substantially compared to modern interglacial conditions (Breemer et al., 2002). Based on simulations beneath the Weichselian ice sheet in Germany, subglacial groundwater flow has been estimated to be faster by a factor of 30 than under nonglacial conditions, and the discharge along the down-ice boundary is c. 4 $m^3\ s^{-1}$ (Piotrowski, 1997; van Weert et al., 1997). Provost et al. (1998) argued that groundwater flow velocities increase with basal melting and that development of permafrost prevent recharge and discharge of groundwater, with flow velocities slowing slightly relative to their present-day magnitudes.

12.4.3.4 Penetration depth of meltwater

The increase of hydraulic heads and groundwater flow velocities tends to increase the penetration depth of meltwater during glacial periods. Using geochemical analyses and radiocarbon ages, the penetration depth of Pleistocene meltwater can be investigated (Section 12.8). Many studies show a meltwater penetration depth increase during glacial periods, principally in response to ice-induced hydraulic loading (Provost et al., 1998; McIntosh and Walter, 2005; McIntosh et al., 2012). Indeed, beneath ice sheets high groundwater gradients in overpressured beds result in an increase in groundwater penetration energy (Wallin, 1995; Piotrowski, 1997; Smellie and Frape, 1997; Tullborg, 1997). Results from palaeorecharge studies during the Pleistocene in Europe and North America suggest that meltwater penetration depth may have reached 1000 m (Glynn et al., 1997) (Fig. 12.7). 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 by meltwater under the Saalian ice sheet (Boulton et al., 1993). Provost et al. (1998) even suggested that recharging basal meltwater could have reached depths of a few kilometres in a few thousand years in deep Cambrian aquifers in Estonia. Clark et al. (2000) demonstrated that pressurized glacial meltwater penetrated the bedrock to depths of over 1.6 km in the North West Territories (Canada). The formation of permafrost in glaciated terrains will tend to drive groundwater flow deeper than it would otherwise penetrate (King-Clayton et al., 1997; Person et al., 2007). Indeed, beyond the continental ice sheet margins, permafrost development may have inhibited meteoric recharge and may have promoted deeper, longer-distance circulation of glacial meltwater.

12.4.3.5 Recharging rates

Beneath warm-based ice sheets, recharging rates significantly increase. Infiltration rates in glaciated terrains can increase between 2 and 10 times the modern conditions (Person et al., 2012). Simulations in intracratonic sedimentary basins at the southern limit of the LIS show that recharging rates are strongly elevated during glaciation compared to nonglacial periods (Person et al, 2007). Permafrost tends to slow or stop groundwater recharge (McEwen and de Marsily 1991; Boulton and de Marsily, 1997; Lemieux et al., 2008a) and its distribution beneath and at the margin

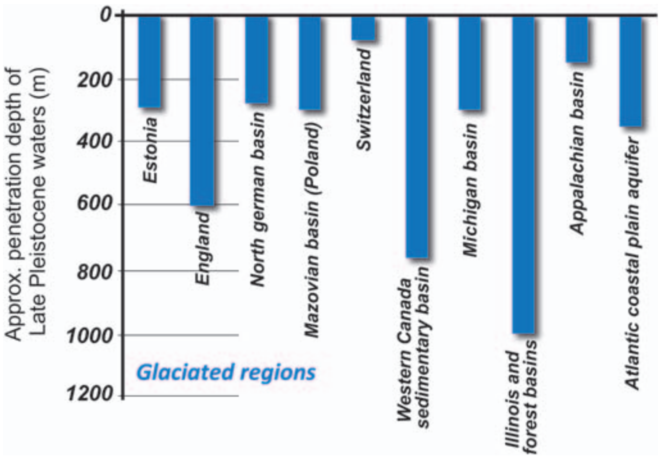


FIGURE 12.7

Reported approximate maximum penetration depths of Late Pleistocene meteoric waters in sedimentary basins across North America and Eurasia.

From McIntosh, J.C., Schlegel, M.E., Person, M., 2012. Glacial impacts on hydrologic processes in sedimentary basins: evidence from natural tracer studies. Geofluids 12 (1), 7–21.

of the ice sheet will affect the zones of recharge/discharge and the recharge rates (King-Clayton et al., 1997).

12.4.3.6 Porewater pressure

Groundwater flow within subglacial substratum is sometimes responsible for the development of overpressure and underpressure zones. Within confined aquifers, the pressure exerted by overlying ice associated with the infiltration of meltwater favours the development of high porewater pressure in the substratum, often referred to as fluid overpressure. Overpressurized groundwater is common beneath and at the ice sheet margin (Boulton and Caban, 1995). Overpressurized zones formed during the Pleistocene glaciations have been determined using zones of anomalously high-amplitude seismic reflections in the continental shelf of Massachusetts (Siegel et al., 2014). In this case, overpressure subsists in modern times as the result of ice loading from the late Pleistocene glaciation and rapid sedimentation. The development of marginal permafrost also enhances overpressure development as permafrost prevents groundwater from welling up directly beyond the ice sheet margin (Piotrowski, 1994, 1997). This effect generates high heads and water overpressures in the proglacial zone in which pressures would be atmospheric in the absence of permafrost.

Based on drill stem test data from wells located in the Cretaceous aquifers of the Alberta Basin, zones of anomalously low fluid pressures have also been measured (Bachu and Underschultz, 1995). The numerical modelling of these underpressure zones shows that basin hydrodynamics cannot produce these zones without mechanical unloading processes (Bekele et al., 2003). Although a significant component of the low-pressure zones is related to unloading by high erosion rates (from Early Eocene), an additional component such as glacial unloading is needed to fit the low-pressure values recorded in the Cretaceous aquifer of the Alberta Basin. Since ice sheets retreat much faster than they advance, the resulting high unloading rates could have led to the formation of underpressure zones within confined aquifers (Vinard et al., 2001; Bekele et al., 2003) (Box 12.1).

BOX 12.1 TWO MODELS BASED ON GROUNDWATER FLOW SIMULATIONS IN NORTH AMERICA DURING THE LAST GLACIATION ILLUSTRATING CHANGES OF HYDRAULIC HEADS VALUES OVER A WHOLE GLACIATION/DEGLACIATION CYCLE (A), AND CHANGES IN GROUNDWATER FLOW PATTERNS, DIRECTIONS, AND VELOCITIES BETWEEN GLACIAL AND INTERGLACIAL PERIODS (B) (FIG. 12.8A AND B)

(A) *Changes in hydraulic heads in intracratonic sedimentary basins* (based on the study of Bense, V.F., Person, M.A., 2008. Transient hydrodynamics within intercratonic sedimentary basins during glacial cycles. J. Geophys. Res.-Earth Surf., 113 (F4)).

The simulations apply to intracratonic sag basin basins with maximum depths of 5000 m and a width (length) of 700 km at the southern edge of the LIS during the last glacial maximum (20 ka BP), in the Williston, Michigan, and Illinois basins. The sedimentary basin consists of four aquifers separated by three semiconfining units (shale beds). The results reveal characteristic spatial patterns of underpressure and overpressure that occur in aquitards and aquifers as a result of recent glaciation and deglaciation. Four main stages can be defined:

- *Stage I: Ice Sheet Advance.* During the advance of the ice sheet toward the south, the flexural response to ice sheet loading decreases the effective ice surface elevation, thus causing a reduction of hydraulic head at the surface. This process occurs in combination with the generation of elevated fluid pressures in the low-permeability confining units due to mechanical loading. This loading triggers hydraulic gradients that force fluid flow toward the surface.
- *Stage II: Last Glacial Maximum (LGM).* During the LGM, hydraulic head disturbances induced by mechanical loading propagate rapidly through the aquifers. High fluid pressures in the aquifers will diffuse into the confining units in between the aquifers. When compared to the initial steady-state conditions, horizontal fluid fluxes in the deeper aquifers steadily rise by up to two orders of magnitude by the end of the LGM.
- *Stage III: Ice Sheet Retreat.* Once the ice sheet starts to retreat, hydraulic heads start to decline. This is a result of the unloading of the basin by the rapid reduction in ice sheet thickness. Below, where the ice sheet was covering the basin, hydraulic heads in the confining units become subhydrostatic because of mechanical unloading.
- *Stage IV: Present-Day Conditions.* In the present-day, low and high hydraulic heads are found in the basin. Near the centre of the basin, overpressures occur at all depths. The location of these anomalous heads is near the position of maximum extent of the former ice sheet. The presence of these anomalously high hydraulic heads can be understood by realizing that lateral migration of high hydraulic heads occurred beyond the toe of the ice sheet during the LGM.

(B) *Variations in groundwater flow patterns, directions, and velocities between glacial and nonglacial times* (based on the study of Breemer, C.W., Clark, P.U., Haggerty, R., 2002. Modeling the subglacial hydrology of the late Pleistocene Lake Michigan Lobe, Laurentide Ice Sheet. Geol. Soc. Am. Bull. 114 (6), 665–674).

Groundwater flow simulations beneath the Lake Michigan Ice Lobe are based on a two-dimensional flow line parallel to ice flow. The model includes a hydrostratigraphic description of six regional aquifer units and four regional confining units that were present below the Lake Michigan Ice Lobe. Under nonglacial conditions (i.e., modern conditions), groundwater is typically recharged in topographically high areas and discharged from topographic lows. However, simulations under glacial conditions indicate that the Lake Michigan Ice Lobe altered or reversed topographically driven pressure gradients, resulting in groundwater flow patterns and velocities that differed substantially from those of current modern conditions. Under glacial conditions, simulated groundwater flow velocities are high. Beneath the ice, groundwater has a relatively strong downward component in the aquifer although an upward component is observed

(*Continued*)

BOX 12.1 (CONTINUED)

close to the margin. The upward component is more expressed in the case of a 7-mm meltwater film at the ice–bed interface (water film thickness is based on results from Ice Stream B in Alley, 1989). Under glacial conditions, when a subglacial water film is absent, groundwater velocities are higher, and flow is consistently directed toward the south.

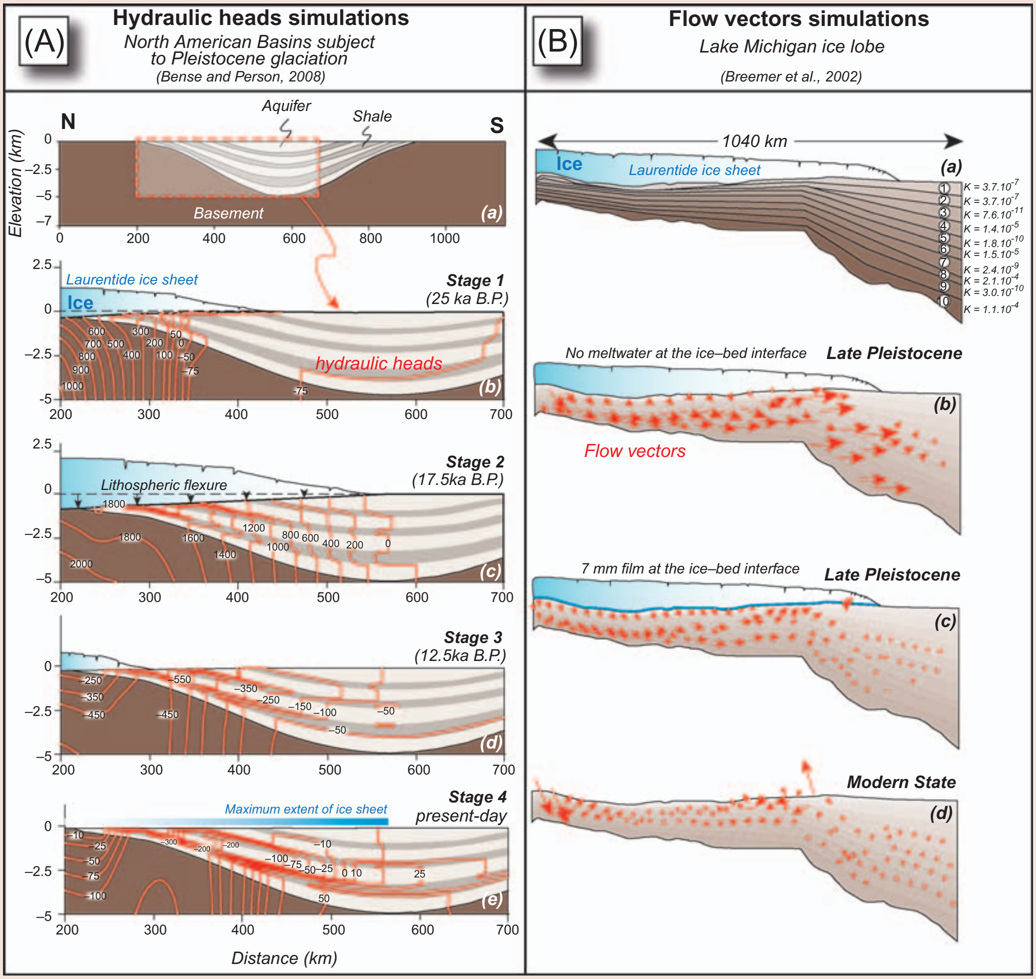


FIGURE 12.8

(A) Simulations of hydraulic patterns within glaciated intercratonic basins. (a) Basin stratigraphy used in the model composed of aquifers and aquitards (shales). (b) Stage 1: hydraulic head values pattern during ice sheet advance. (c) Stage 2: hydraulic head values during the LGM. (d) Stage 3: hydraulic head values during ice sheet retreat. (e) Stage 4: hydraulic head values in modern conditions. (B) Simulations of groundwater flow directions and velocities beneath the Late Pleistocene Lake Michigan Lobe. (a) Model geometry and hydrostratigraphic units used in simulations. (b) Groundwater vectors under glacial conditions with no drainage at the ice–bed interface. (c) Groundwater vectors with a 7-mm-thick water film at the ice–bed interface. (d) Groundwater flow vectors in present-day conditions.

12.4.4 POSTGLACIAL GROUNDWATER FLOW EVOLUTION

During interglacial periods, the zone of active groundwater circulation in formerly glaciated sedimentary basins is mostly limited to uppermost glacial sediments and shallow fractured rocks (McIntosh and Walter, 2005) (Fig. 12.6). However, present-day subsurface hydrogeological conditions across formerly glaciated areas are likely still out of equilibrium and reflect the impact of the last glaciation. This suggests that the current groundwater flow system cannot be interpreted solely on the basis of present-day boundary conditions and it is likely that several thousands of years of additional equilibration time will be necessary for the system to reach a new quasi-steady-state. Zones of overpressure and underpressure that still exist in many formerly glaciated sedimentary basins demonstrate this non-equilibrium configuration of modern groundwater flows. These particular zones are related to processes of glacial loading and unloading experienced by confined aquifers during glaciation and deglaciation (Bense and Person, 2008). Long timeframes are required to dissipate these pressure anomalies at large depths (Vinard et al., 2001; Bekele et al., 2003). After complete retreat of ice sheets, zones of overpressurization or displaced water of high salinity may provide the driving force for modern movement of groundwater (Ferguson et al., 2007) (Fig. 12.6). Evidence for this modern groundwater movement is provided by the presence of saline springs along basin margins in the Michigan and Willinston Basin (Grasby et al., 2000; Grasby and Betcher, 2002; McIntosh and Walter, 2005). Cold water brines formed when subglacial meltwater intruded deep enough into the sedimentary basin during the Pleistocene glaciations to dissolve Palaeozoic evaporites. Nowadays, the regional-scale flow system is reversed as a result of glacial unloading, thus leading to extrusion of brines and subsequent saline springs developments of saline springs along sedimentary basin margins.

12.5 EFFECT OF GROUNDWATER DRAINAGE ON GLACIER DYNAMICS

Groundwater flow in a pressurized subglacial system influences ice sheet stability and ice flow mechanisms. The occurrence and importance of basal sliding and/or subglacial sediment deformation to ice sheet motion are linked to the subglacial hydrology of the ice sheet. Recently, glaciologists have begun to link groundwater flow models to an ice sheet dynamics model to explore the coupling between subglacial water content, fluid pressures, and basal sliding rates (Le Brocq et al., 2009). As an example, small groundwater discharge capacity coupled with high meltwater production rates result in high porewater pressures in the substratum, which can in turn lead to rapid ice flow, a surge, and even initiate ice sheet collapse (Piotrowski, 1997). Groundwater flow models developed for the Lake Michigan Ice Lobe during the Late Wisconsinan glaciation demonstrate the inefficiency of subglacial aquifers to transmit the estimated meltwater flux, implying that pressurized meltwater must have been present within the bed or at the ice–bed interface (Breemer, 2006). This pressurized meltwater can lubricate the base of the ice and contribute to fast and unstable ice flow. Conversely, extremely permeable sediments would have the capacity to transmit all of the basal meltwater through the bed and beyond ice sheet margins. This would reduce subglacial pore pressure and/or drain any film at the ice–bed interface, significantly increasing the resistance to ice flow at the bed and lowering flow velocities. Such a decrease of ice flow velocities in response to efficient bed drainage has been described for Ice Stream C in West Antarctica (Anandakrishnan and Alley, 1997; Breemer et al., 2002).

12.6 LANDFORMS RESULTING FROM INEFFICIENT GROUNDWATER DRAINAGE

When ice overlies sediments with low hydraulic conductivity, drainage pathways at the ice–bed interface such as channels, tunnels, or canals have to form to counterbalance the inefficient groundwater drainage (Shoemaker, 1986; Boulton and Caban, 1995; Ng, 2000). The meltwater conduits preferentially form at locations near the ice sheet margin where the subglacial heads increase and balance the ice weight (Shoemaker and Leung, 1987; Boulton et al., 2007b). It has been suggested that basal meltwater recharge rates and groundwater flow characteristics exert a major control on the frequency and location of eskers and tunnel valleys (Boulton and Caban, 1995; Boulton et al., 2007a). Subglacial groundwater flow model characteristics allow the spacing of meltwater channels and eskers to be determined (Boulton et al., 2009). Simulations beneath the James Bay Lobe of the LIS show that the subglacial drainage system at the ice–till interface composed of conduits with widths of 0.5–1.0 m and heights of 0.1 m–0.3 m, would most likely have been spaced on the order of tens–hundreds of metres apart to have drained the excess meltwater (Carlson et al., 2007). Where bed transmissivity is low, tunnel valleys occur and eskers are rare. Indeed, tunnel valleys form preferentially where the hydraulic transmissivity is low, causing reverse hydraulic gradients from the bed toward the ice–bed interface (Piotrowski et al., 2009).

Tunnel valleys form to secure the stability of the ice sheet by reducing the water pressure at the ice–bed interface, thereby preventing catastrophic surges and ice sheet collapse (Marczinek and Piotrowski, 2006; Piotrowski et al., 2009). Two models of tunnel valley formation are directly linked to hydraulic properties of the substratum and groundwater flow characteristics. The first model suggests that development of ice marginal permafrost considerably reduces the bed hydraulic conductivity and blocks the subglacial drainage up-ice from the extent of the permafrost wedge (Johnson, 1990; Piotrowski, 1994; Cutler et al., 2000; Clayton et al., 2001; Hooke and Jennings, 2006). In this model, meltwater is unable to drain and therefore accumulates in subglacial lakes up-ice, resulting in a drastic increase in basal meltwater pressure. When the meltwater pressure exceeds the strength of the permafrost seal, rupture and catastrophic drainage of the impounded meltwater occur and trigger the formation of tunnel valleys (Piotrowski, 1994; Cutler et al., 2002; Hooke and Jennings, 2006). The second model concerns glaciers that rest directly on aquifers or are separated from them only by a thin layer of low-permeability sediments. In this model, the bed inefficiency to drain all meltwater produced subglacially is responsible for the formation of channels at the ice–bed interface. Channel formation reduces the water pressure and generates a hydraulic gradient driving groundwater from the surrounding sediment into the channel, thus creating a catchment area along the channel (Boulton and Hindmarsh, 1987). This hydraulic gradient triggers groundwater flow toward the channel leading to porewater pressure increase, sediment liquefaction, and injection into the channel (Boulton and Hindmarsh, 1987). This second model of tunnel valley formation was suggested independently by Shoemaker and Leung (1987) and Boulton and Hindmarsh (1987), and measured in piezometers arranged around a channel under Breiðamerkurjökull in Iceland (Boulton et al., 2001).

12.7 IMPACTS OF SUBSURFACE INCREASING MELTWATER PRESSURE

12.7.1 OVERPRESSURIZED MELTWATER

Overpressure corresponds to a state where porewater pressure in the sediments is very high. In glacial environments, high porewater pressures may develop in the substratum because of ice and sediment loading, high meltwater recharge, and low drainage capacity of the bed (i.e., low hydraulic conductivity). Changes in porewater pressure have a considerable impact on the processes of sediment deformation. The increase in porewater pressure induces changes in the mechanisms controlling deformation, from brittle to ductile and from hydroplastic to liquefaction/fluidization (Lowe, 1975; Maltman, 1994). Over large areas, a subglacial drainage system is likely to produce high porewater pressures in the subglacial soft bed, reducing the effective stresses to a level low enough to induce deformation by liquefaction or fluidization.

12.7.2 STRUCTURES AND LANDFORMS LINKED TO OVERPRESSURIZED MELTWATER

A range of structures and landforms derives from fluid overpressure development in the bed, including hydrofractures, clastic injections, diapirism, extrusion moraines, or tunnel valleys, for example.

Hydrofracturing is triggered when the porewater pressure within sediments exceeds the tensile strength of the encasing unit, and injection is produced as a response of porewater excess release from an overpressurized parent bed. Recent works conducted in glacial environments illustrate that hydrofracturing and clastic injections predominantly occur within subglacial to marginal environments (Meer et al., 2009; Phillips et al., 2013; Ravier et al., 2015). Boulton and Caban (1995) suggested, using modelling, that hydrofracturing can occur to depths of 400 m and tens of kilometres ahead of the glacial margin.

Clastic injections form in response to the injection of remobilized liquefied or fluidized sediments within hydrofractures (Cosgrove, 2001; Jolly and Lonergan, 2002). Previous studies seem to indicate a preferential distribution of these structures according to three main subenvironments: (1) subglacial; (2) submarginal; and (3) proglacial marginal (Ravier et al., 2015). (1) In subglacial environments, the combination of high porewater pressures and basal shear stresses induced by flowing ice triggers the formation of *per descensum* clastic dykes dipping down-ice (Larsen and Mangerud, 1992; Dreimanis and Rappol, 1997; Meer et al., 1999) (Fig. 12.9A and C). (2) In submarginal environments, the decrease of ice overburden pressure (decreasing ice thickness) tends to promote the formation of sills because the propagation of hydrofractures is mainly controlled by bedding anisotropy (Kumpulainen, 1994; Phillips et al., 2013) (Fig. 12.9B). (3) In proglacial marginal settings, the absence of ice overburden pressure promotes the upwards release of overpressurized meltwater and thus the formation of *per ascensum* dykes (Fig. 12.9D). When porewater pressure is very high and ice decoupled from its bed, sediments are in hydrostatic stress, implying that no stress exerted by the ice is transmitted to the bed. In this stress configuration, hydrofractures

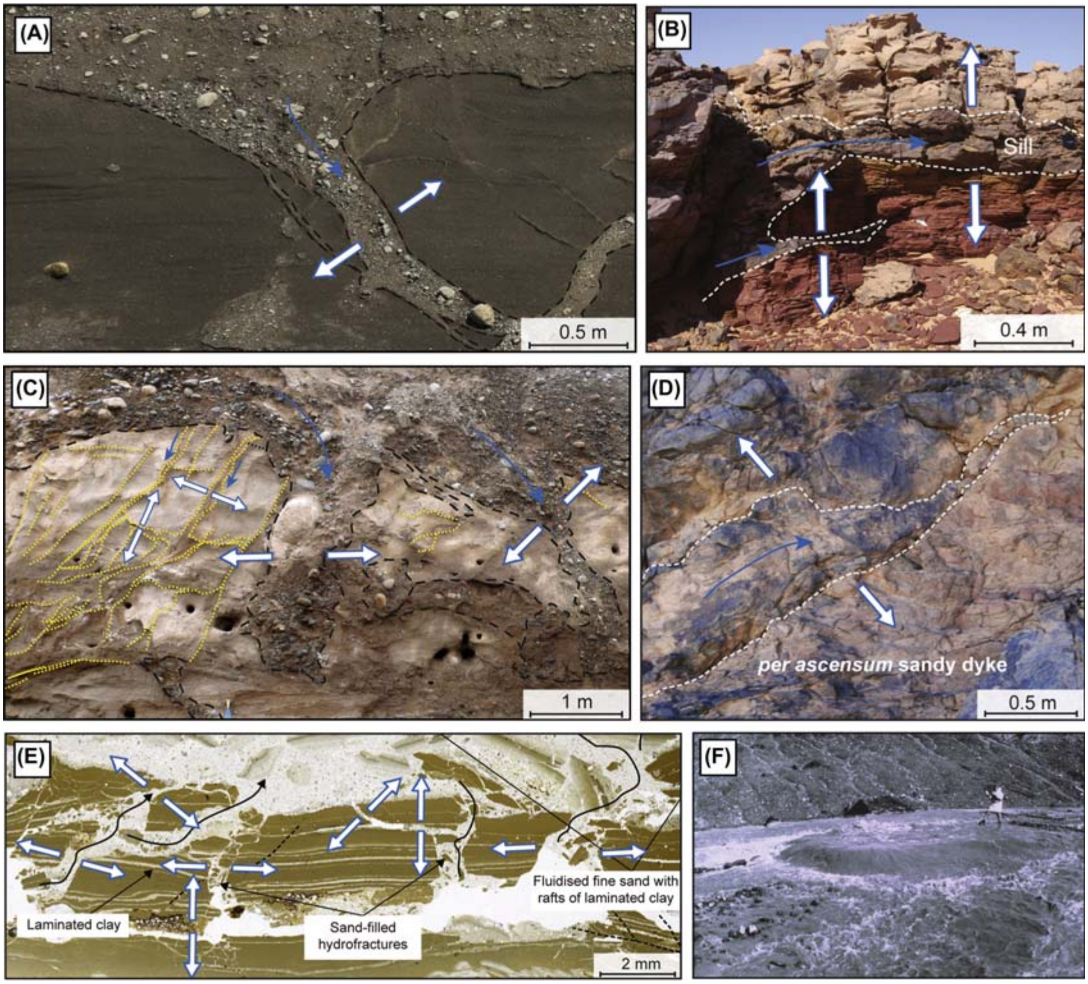


FIGURE 12.9

(A) Dyke propagating downward and dipping down ice (Iceland, Holocene). (B) Sandstone sill propagating parallel to bedding in shales (Niger, Ordovician). (C) Multiphased hydrofracturing and injection of sediments (Ireland, Pleistocene) (Clerc et al., 2012). (D) Dyke propagating upward (Morocco, Ordovician) (Ravier et al., 2015). (E) Microscale hydraulic brecciation formed as a response of dense hydrofracturing (Scotland, Pleistocene) (Phillips et al., 2007). (F) Strong upwelling of meltwater beyond the Breiðamerkurjökull margin (Iceland) (Boulton et al., 2007a).

and injections of sediments develop in every direction, potentially leading to in situ hydraulic brecciation (Fig. 12.9E).

At the glacier margin, strong flow of groundwater toward low-pressure areas is a source of upwelling that can produce sediment liquefaction and instability, and may be responsible for the formation of diapiric structures or blowout structures (Christiansen et al., 1982; Kjær et al., 2006) (Fig. 12.9F). The circumstances which produce diapirism have generally been assumed to be changes in ice or sediment overloading (Szuman et al., 2013). For example, overpressure development in saturated sand leads to a decrease in its bulk density. When overlain by clays, this decrease in sand bulk density is responsible for a density imbalance between these two sediments causing sand to rise in clay (i.e., diapirism). The extrusion of sediments in marginal environments induced by diapirism leads to the formation of small landforms at the glacier front, sometimes referred to as extrusion moraine. It has also been suggested that groundwater overpressures associated with narrow proglacial permafrost plates promote the formation of large push moraines (Waller et al., 2012).

The development of fluid overpressure in subglacial soft bed is also responsible for the formation of certain tunnel valleys. Unlike previous tunnel valley formation models invoking erosion of the substratum by meltwater flow (Brennand and Shaw, 1994; Hooke and Jennings, 2006; Boulton et al., 2009), this new model would suggest that single-tunnel valleys are formed subglacially under the influence of localized development of overpressure in the unlithified substratum (Box 12.2). Increasing porewater pressure triggers hydrofracturing and removal of fractured bed materials through subglacial channels. This model is based on field evidence of hydraulic fracturing in Morocco (Ordovician) but also through the study of subsurface data in the North Sea (Pleistocene) (Janszen et al., 2012; Ravier et al., 2014). This model is based on a strong lithological control where the hydrostratigraphy and hydraulic conductivities of the different layers define multiple confined aquifers promoting the development of high porewater pressure within the subglacial bed. As suggested by Sandersen and Jørgensen (2012), a preglacial substratum with abundant low-permeable

BOX 12.2 PORE-WATER PRESSURE CONTROL ON SUBGLACIAL SOFT SEDIMENT REMOBILIZATION AND TUNNEL VALLEY FORMATION: A CASE STUDY FROM AN ORDOVICIAN TUNNEL VALLEY IN MOROCCO

In the eastern part of the Moroccan Anti-Atlas Mountains, the Alnif area exposes a buried Ordovician glacial tunnel valley (5 km wide, 180 m deep) cut into preglacial marine sediments. The preglacial sedimentary sequence, deposited in a marine environment, is characterized by an alternation of sand and clay beds, thus forming confined aquifers. At the base of the tunnel valley, a discontinuous and fan-shaped glacial conglomeratic unit 10−15 m thick occurs, deposited over preglacial marine sediments (Fig. 12.10A). The conglomeratic unit is composed of angular preglacial intraclasts embedded within a sandy matrix. Kinematics and relative chronology of deformation structures allow the role of fluid overpressure in the process of tunnel valley genesis on soft beds to be understood. The occurrence of deformation structures related to increasing porewater pressure, within preglacial sediments, represents major evidence for a porewater pressure-driven model. Ball structures, clastic dykes, and hydraulic breccia, locally observed below glacial erosional contact surfaces, illustrate the increase in porewater pressure within preglacial strata due to the combination of meltwater input and the stress exerted by the ice on preglacial confined aquifers (Fig. 12.10A and B). The tunnel valley formed through multiphased episodes of increasing porewater pressure that led to hydrofracturing, brecciation and remobilization of the subglacial soft sediments (Fig. 12.10D). Transport of the resulting conglomerate composed of preglacial intraclasts and fluidized sand occurred through subglacial pipes (Fig. 12.10D). Conglomeratic fans observed at the base of the tunnel valley are related to the different episodes of hydraulic brecciation and piping of the preglacial sediment layers. Principal parameters controlling the evolution of the porewater pressure in the preglacial sediments are: (1) preglacial lithology and stratigraphy, (2) ice thermal regime and dynamics, and (3) the meltwater production and infiltration within the bed.

(*Continued*)

BOX 12.2 (CONTINUED)

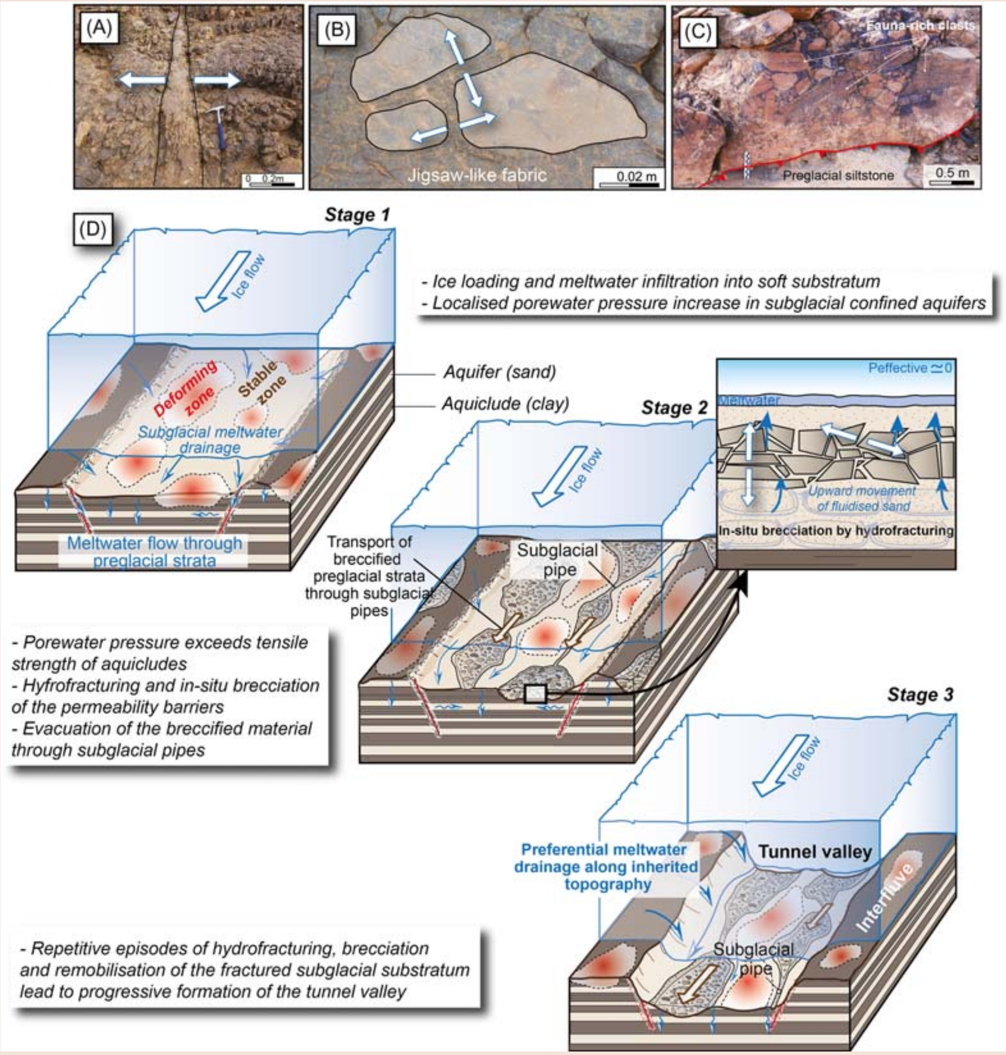


FIGURE 12.10

(A) Clastic dyke in preglacial sediments. (B) In situ breccia formed by hydrofracturing. (C) Breccia composed of angular preglacial clasts floating in a fluidized sandy matrix. (D) Model of tunnel valley formation by hydrofracturing and removal of preglacial strata during episodic porewater pressure increase in unconsolidated sediments.

Modified after Ravier, E., Buoncristiani, J.-F., Guiraud, M., Menzies, J., Clerc S., Portier, E., 2014. Porewater pressure control on subglacial soft sediment remobilization and tunnel valley formation: a case study from the Alnif tunnel valley (Morocco). Sediment. Geol. 304, 71–95.

layers will preferentially contribute to the development of tunnel valleys induced by fluid overpressure. The formation of tunnel valleys by fluid overpressure development in the subsurface is also controlled by the rates of meltwater production and the meltwater recharge areas.

12.8 HYDROGEOCHEMISTRY

Hydrogeochemical studies are common in groundwater sciences, notably because they allow the palaeohydrogeology to be investigated. Hydrogeochemistry is therefore of great use in reconstructing past groundwater circulation under ice sheets because meltwater recharging the subsurface has a geochemical signature that can be discriminated from other sources of water generated under warmer climates (Smellie and Frape, 1997; Glynn et al., 1997). Several hydrogeological factors may affect the geochemical composition of groundwater: the climate conditions, the lithology of the substrate, and its hydraulic conductivity, the weathering, the length of the flow path, the penetration depth, the groundwater flow velocity, the residence time of groundwater, and the rock:water ratio. Groundwater recharge during the Pleistocene has an impact on: the stable isotope composition, the amount of dissolved solid, noble gas temperatures, regional water salinity, and microbial gas production (Siegel, 1990; McIntosh and Walter, 2005).

12.8.1 GEOCHEMICAL SIGNATURE OF GLACIAL MELTWATER

Glacial meltwater has distinct chemical and isotopic characteristics. Meltwater is generally isotopically depleted. Past glacial meltwater recharge is evidenced by low $\delta^{18}O$, δD, or δ^2H signatures in groundwater compared to precipitation formed under modern warmer climate conditions. Low isotopic ratios of glacial palaeowaters are principally inferred from colder glacial climatic conditions and especially snow formation conditions (temperatures of condensation of the precipitations and mechanism of snowfall). Isotopically light pore fluids identified as meltwater have been found in aquifers located within sedimentary basins at mid to high latitudes across North America and Europe (Siegel, 1990, 1991; Remenda et al., 1994; Grasby et al., 2000; Edmunds et al., 2001; Person et al., 2003, 2007; Grasby and Chen, 2005; McIntosh and Walter, 2005; McIntosh et al., 2012; Ferguson and Jasechko, 2015). Groundwater at depths of up to 300 m in Silurian-Devonian aquifers and shales along the northern margin of the Michigan Basin have $\delta^{18}O$ of -15 to -10‰ and δD values of -100 to -75‰, indicating Pleistocene recharge (McIntosh and Walter, 2005, 2006). Glacial groundwater may have $\delta^{18}O$ values as low as c. -30‰. Most old deep glacial groundwater, however, shows slightly higher signatures, probably due to mixing with heavier preglacial water. Noble gas recharge temperatures were approximately 6°C cooler than at present across Europe in permafrost-covered regions. The presence of isotopically depleted groundwater combined with noble gas temperatures close to the freezing point (1–3°C) and an age close to the last glacial maximum (21,000 ka) is a good indicator for water recharged by melting ice sheet (Person et al., 2007).

Ice composition data from the base of Greenland Ice Sheet and glacial meltwater data from Iceland and from the European Alps suggest that glacial meltwaters may be highly enriched in dissolved oxygen, with concentrations at least three to five times greater than would be at atmospheric equilibrium (Glynn et al., 1997). Glynn and Voss (1996) have suggested that highly oxygenized water may cause a major decrease in dissolved solids and changes in the geochemical composition of meltwater.

12.8.2 IMPACTS ON SALINITY AND BRINE MIGRATION

During the Pleistocene, modification in the penetration depth of meltwater and meltwater flow patterns can significantly depress the freshwater−saline water interface (McIntosh and Walter, 2006; Vidstrand et al., 2010). Profound changes to the salinity structure and fluid composition basins by relatively recent freshwater invasion have major implications for the stability of basinal brines, mass transport of solutes in the subsurface, and chemical evolution of basinal fluids (Fig. 12.6). In North America, glacial meltwaters dissolved large quantities of halite along the margins of several sedimentary basins, including the Michigan, Appalachian, and Western Canada basins, generating 'Pleistocene age' brines. The dissolution of salt and generation of dense brines may have enhanced or impeded the penetration of glacial meltwaters and altered basinal-scale fluid and solute transport (McIntosh et al., 2012). Glaciation was able to reorganize salinity gradients and drive basin-scale fluid migration within only several thousands of years for the Michigan Basin (McIntosh et al., 2011). Under glacial conditions characterized by ice-sheet growth and permafrost development, the position of the brine interface may be expected to change. The interface may most likely be pushed down as glacial waters are introduced into the groundwater system, but the possibility of it rising is also considered (King-Clayton et al., 1997; McIntosh and Walter, 2005). Studies led on the Canadian Shield show that circulation of glacial meltwaters at the end of the Pleistocene flushed brines to depths as great as 1600 m (Douglas et al., 2000).

12.8.3 METHANOGENESIS

Recent studies carried out on the Michigan and Illinois Basins in North America demonstrate that groundwater flow during glacial periods may have been involved in the generation of microbial gas (McIntosh et al., 2002; McIntosh and Walter, 2005). Indeed, glacial meltwater has migrated into Silurian-Devonian carbonate aquifers and enhanced generation of microbial methane as meltwater circulated through overlying fractured organic-rich Upper Devonian shales (Martini et al., 1996, 1998; McIntosh et al., 2002). Glaciation likely promoted microbial methanogenesis by significantly diluting the salinity of ambient formation waters and dilating natural fractures. Microbial gas is generated in situ at relatively shallow depths, and CO_2 and CH_4 are adsorbed onto the organic matrix.

12.9 ISSUES AND APPLICATIONS

Although studies led on groundwater movement in glaciated domains remain scarce, today's issues concerning water recovery, hydrocarbon reservoirs, and nuclear waste treatment have revealed the importance of this discipline.

12.9.1 NUCLEAR WASTE TREATMENT

Several studies on subsurface flow during glacial periods emphasize the potential impact of groundwater flow modifications on high-latitude nuclear waste repositories and carbon dioxide repositories in Switzerland, Sweden, and Canada (Vinard et al., 2001; Sheppard et al., 1995; Talbot, 1999; Heathcote and Michie, 2004). High-level nuclear waste must be isolated from the biosphere for at least 10^6 years (Fyfe, 1999). However, climate models indicate that glacial episodes involving widespread ice cover in northern Europe and North America will occur several times during the next 10^5 years or more, thus potentially impacting the behaviour of deep geological repositories. The assessment of the long-term safety of radioactive waste disposal requires assimilation of evidence for the impact of climate change and especially glaciation on the groundwater flux and flow pattern, rock stresses, and groundwater chemistry (King-Clayton et al., 1997; Person et al., 2007; Iverson and Person, 2012). The minimum penetration depth of meltwater estimated being over 1000 m should be modelled before evaluating the site capacity to host nuclear waste. In addition, the high dissolved oxygen content of glacial groundwater may potentially increase the solubility and mobility of many radionuclides by several orders of magnitude, therefore increasing risk for contamination (Glynn et al., 1997; van Weert et al., 1997). Several potential nuclear waste repository sites subject to groundwater flow changes during glaciations have been investigated in Canada and Sweden. In Canada, the Palaeozoic sedimentary rocks of southern Ontario display thick Ordovician shales (>100 m) of low hydraulic conductivities that could be suitable for hosting a deep geologic repository (DGR) (Mazurek, 2004). The Canadian Shield has also been investigated to host a DGR as crystalline rocks potentially offer stable mechanical and biogeochemical conditions for thousands of years. However, the deep penetration depth of meltwater that is usually enriched in dissolved O_2 has to be seriously considered because of the redox stability of deep repositories (Spiessl et al., 2008). Two sites of southeastern Sweden (Forsmark and Laxemar-Simpervarp) have been studied for future DGR in order to quantify hydrogeological conditions associated with future glacial periods (Vidstrand et al., 2010). The impact of the glacially induced groundwater flow at the Forsmark site was estimated using numerical simulations of groundwater flow with and without the presence of a subglacial to marginal permafrost. Results show that the increase of hydraulic gradients and Darcy fluxes and the distortion of the fresh water−saline water interface during glaciations have to be taken into account to assess safety applications for future DGRs at the Forsmark site (Vidstrand et al., 2010).

12.9.2 POTABLE WATER RESOURCES

Meltwater movement in subglacial permeable rocks and sediments can influence renewal of groundwater resources. Isotopically light water of glacial origin is usually fresh to slightly brackish and is therefore a suitable domestic and commercial water supply (Ferguson et al., 2007). In addition, meltwater is generally of excellent quality and because of its age, typically 14−21 ka, it is not tainted by groundwater contamination associated with the post-Industrial Revolution (Edmunds, 2001). During glaciations, the large influx of fresh meltwater in sedimentary basins associated with changes of groundwater flow characteristics related to ice loading (penetration depth, flow reversal, and velocities) are responsible for the dilution and displacement of basinal brines. These brines impact the distribution of shallow drinking water resources by affecting the current position of the freshwater boundary. This reorganization needs to be well constrained for determining the sustainability and potable water resources and water quality and their distribution in sedimentary basins (McIntosh and Walter, 2006). The management of water resources in formerly glaciated areas requires to understand the palaeohydrogeology of regional aquifers using numerical modelling and hydrogeochemistry. The understanding of palaeohydrogeology of regional aquifers has major implications for the residence times and sustainability of drinking water resources. Pleistocene tunnel valleys' sedimentary infills are known to provide good water reservoirs consisting of several tens of metres of thick sand and gravels protected from subsurface pollution by some tens of metres of low-permeability strata (clay and/or tills) (Ehlers and Linke, 1989). Geophysical investigations and drillings conducted on Pleistocene tunnel valleys in Denmark, Germany, the Netherlands, and North America have located and characterized well-protected aquifers in buried tunnel valleys (Gabriel et al., 2003; Jørgensen et al., 2003; Kluiving et al., 2003; Cummings et al., 2012).

12.9.3 HYDROCARBON RESERVOIRS

Groundwater flowing in glaciated terrains has major impacts on generation of methane, tilting of oil/water contacts, hydrocarbon migration pathways, and development of hydrocarbon reservoirs. As described in Section 12.8.3, circulation of glacial meltwater can be responsible for the production of biogenic gas. It is likely that repeated cycles of glacial meltwater invasion across this region produced a unique class of unconventional shale-hosted gas deposits. Such resources are currently being exploited in the Michigan and Illinois Basins (Martini et al., 1996, 1998; Hill and Nelson, 2000; McIntosh et al., 2002, 2004). It is also known from observations in wells that oil has been flushed or has migrated because of increasing meltwater recharge during glacial periods in the Barents Sea (Forsberg, 1996). In addition, water/oil contacts may have been tilted and gas in reservoirs compressed. The models predict a tilting of the water/oil contact by 1/10 under simulated glacial conditions (Forsberg, 1996; Flowers and Clarke, 2002). Fluid pressures measured in petroleum wells are commonly used to generate models for basin-scale fluid flow and to predict hydrocarbon migration pathways. Terrains subject to glaciation show variations in basin-scale flow systems through time, which need to be taken into account for reconstructing hydrocarbon migration pathways. For example, the thickness of the continental ice sheet over Manitoba (Canada) would have provided sufficient potentiometric to modify the hydrocarbon migration pathway in the Williston basin (Grasby et al., 2000). Groundwater also has an influence on hydrocarbon reservoir development since tunnel valley genesis is directly or indirectly controlled by groundwater flow. Indeed, tunnel valleys represent elongated depressions that can accumulate thick successions of sediments constituting potentially good reservoirs (i.e., sand and gravels) for hydrocarbons (Ghienne and Deynoux, 1998; Hirst et al., 2002; Clerc et al., 2013). These valleys represent considerable repositories of glaciogenic deposits of high preservation potential (compared to intervalleys area) and therefore constitute potential pay zones. Reservoirs in glaciogenic sediments have been identified through subsurface imaging and boreholes data in Neoproterozoic, Late Ordovician, Permo-Carboniferous, and Late Pleistocene sedimentary record from all continents (Huuse et al., 2012, and references therein). Due to their significant production, Late Ordovician tunnel valleys from North Africa (Algeria and Libya) and the Middle East (Jordan, Syria, Iraq, and Saudi Arabia) are certainly petroleum and gas plays that have received the most attention in the past decades (Wender et al., 1998; Le Heron et al., 2004; Al-Ameri and Wicander, 2008; Hirst, 2012). Late Ordovician glaciogenic clastic sediments of North Africa are good plays because they are entrapped between two mature source rocks, respectively from Ordovician and Silurian, that favour migration of hydrocarbon into the glaciogenic reservoirs (Lüning et al., 2000). In most glaciogenic hydrocarbon systems, reservoirs are topped by condensed organic-rich sediment accumulation (future source rock) as a result of rapid transgression and high sea-level stands triggered by deglaciation (Craig et al., 2009). Tunnel valleys are potentially good water and hydrocarbon reservoirs, however much remains to be known for being able to predict their locations and therefore facilitate their economical exploitations.

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