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Review

Models of ice-sheet hydrogeologic interactions: a review

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ABSTRACT

This study reviews the state-of-the-art and promising pathways to advance hydrologic models of groundwater flow systems and related transport processes in response to transient glacial loading. We also discuss the utility of hydrologic and geochemical data sets as a means of providing ground truth for these models. The paleohydrologic models presented herein should be used as analogues to assess high-level nuclear waste repository stability in response to future episodes of glaciations in countries such as Canada, Sweden, and Switzerland. The next generation of fully coupled ice-sheet-aquifer models may also be of use in assessing rates of ice sheet denudation on Greenland and Antarctica in response to global warming. However, significant uncertainty exists in paleoclimatic forcing, paleohydrologic boundary conditions, and effective basin-scale petrophysical parameters. Thus, model results must be viewed with some caution. Model results from studies reviewed herein suggest that during the last glacial maximum, recharge rates across glaciated basin margins increased by as much as 2–6 times modern levels. Paleohydrologic models predict that as ice sheets overran sedimentary basin margins, glacial melt water penetrated to depths of up to hundreds of meters. Recent ice-sheet models that incorporated the effects of groundwater flow suggest that the presence of a 1–10 mm film of water at the glacial bed can increase basal ice sliding rates by up to 4 orders of magnitude. No firm theoretical basis exists for coupling ice sheet and subsurface hydrogeologic models nor the effects of permafrost on hydraulic conductivity. These issues could be resolved, to some degree, by additional careful experimental studies. Analysis of fluid pressures and flow rates beneath modern ice sheets using geochemical tracers would help to reduce the uncertainty regarding suitable hydrogeologic boundary conditions, parameterization of poromechanical coupling, and transport processes. Glacial geologists should work closely with modelers to provide better constraints on model boundary conditions.

Key words: hydrogeololgy, ice-sheet, numerical models, review paper

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INTRODUCTION

During the last two decades, a series of hydrogeologic models have been developed to reconstruct the Pleistocene hydrology of North America and Europe with different goals, levels of complexity, scales, and resolution (Fig. 1; Table 1). The first attempts to quantify the hydrologic effects of glaciation were from a series of studies focused on reconstructing Late Pleistocene groundwater systems beneath the Fennoscadian ice sheet by Boulton *et al.* (1995), Boulton & Caban (1995), Frosberg (1996), Piotrowski (1997a,b), and Lerche *et al.* (1997). They were also motivated, in part, by issues pertaining to nuclear waste isolation. These papers assessed the impact of elevated fluid pressures beneath and beyond the ice sheet on sediment deformation and petroleum migration. More recently, a series of subsurface hydrologic models were constructed by Breemer *et al.* (2002), Person *et al.* (2003), Bekele *et al.* (2003), Hoaglund *et al.* (2004), Person *et al.* (2007), Lemieux *et al.* (2008a,b,c), Bense & Person (2008), and Cohen *et al.* (2009), assessing the role of the Laurentide Ice Sheet on

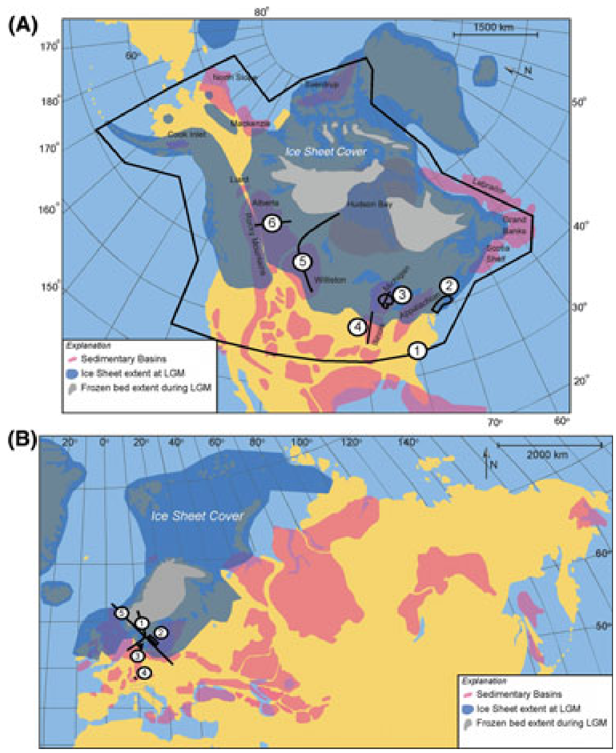

Fig. 1. Distribution of ice cover (dark gray), sedimentary basins (red), permafrost extent (black line), and frozen ice sheet bed (light gray) during at the last glacial maximum across Europe and Northern America (data from Denton and Hughes, 1981; Pewe, 1983; Kleman and Haettestrand, 1999). The locations of mathematical modeling studies for North America (A) and Europe (B) are shown by the black bullets and open circles. References for Figure 1A are as follows: Location-1, Lemieux *et al*. (2008a,b,c), Location-2, Cohen *et al*. (2009), Location-3, Hoaguland *et al*. (2004), Location-4, Breemer *et al*. (2002), Location-5, Carlson *et al*. (2007), Location-6 Bekele *et al*. (2003). References for Figure 1B are as follows: Location-1, Provost *et al*. (1998), Location-2, Boulton *et al*. (1995), Location-3, Piotrowski (1997a,b), Location-4, Vinard *et al*. (2001).

groundwater systems across North America. These studies were largely motivated by geochemical and environmental isotopic data sets suggesting the emplacement of fresh, isotopically depleted glacial meltwater at the margins of sedimentary basins overrun by ice sheets (Hathaway *et al.* 1979; Siegel & Mandle 1984; Siegel 1989, 1990, 1991; Grasby *et al.* 2000; McIntosh *et al.* 2002, 2004; Grasby & Chen 2005; McIntosh & Walter 2005). Nuclear waste repository sitting requirements has also resulted in the development of a number of groundwater models focused on evaluating the impact of ice sheets on subsurface groundwater systems in Switzerland (Vinard *et al.* 1993, 2001), Canada (Jaquet & Siegel 2003; Vidstrand *et al.* 2008a,b; Mayer & MacQuarrie 2007), Finland (Hutri & Antikainen 2002), and Sweden (Provost *et al.* 1998). Recently, glaciologists have begun to link groundwater flow models to ice-sheet codes to explore the coupling between subglacial water content/fluid pressures and basal sliding rates (Flowers *et al.* 2005; Le Brocq *et al.* 2009).

In this study, we summarize the types of models that have been developed and what has been learned from mathematical modeling studies of ice-sediment hydrologic interactions so far. We also discuss the uncertainty inherent in these model predictions owing to spatially and temporally varying material properties, as well as uncertain boundary and initial conditions. We discuss how hydrologic and geochemical data sets can be used to constrain model results and how paleohydrologic model results can be used to assess the suitability of proposed nuclear waste repositories in countries such as Canada, Sweden, and Switzerland. We make some suggestions regarding promising future research directions at the end of this study.

GROUNDWATER SYSTEM EVOLUTION AND GLACIAL PERTURBATIONS

The assumptions and transport processes represented in ice-sheet hydrogeologic modeling studies are summarized in Table 1. None of these hydrogeologic models fully couples all of the relevant processes discussed by Neuzil (2012). For example, we are unaware of any published study that fully couples ice sheet dynamics, groundwater flow, effective stress, and related porosity–permeability changes. Limitations in the complexity of ice-sediment hydrologic models may be due, in part, to the complexity of ice sheet hydrology (Jansson *et al.* 2007), uncertainty regarding suitable boundary conditions, uncertainty in parameterization of permafrost, and geomechanical effects. Model testing/validation has largely been limited to comparisons against geochemical and isotopic tracers rather than on direct pore-pressure measurements beneath modern continental glaciers. Some model intercomparison studies have been undertaken by the Swedish, Canadian, and Finnish nuclear waste management agencies considering the effects of glaciation (DECOVALEX III, 2004; Chan *et al.* 2005; Vidstrand *et al.* 2008a,b).

Coupling

Representation of glacial processes

With the possible exceptions of Flowers *et al.* (2005) and Le Brocq *et al.* (2009), all models of ice-sediment hydrologic interactions developed to date specify the growth and decay of ice sheets using either a polynomial function (e.g., Hoaglund *et al.* 2004; Bense & Person 2008) or output from ice sheet models (e.g., Boulton *et al.* 1995; Lemieux *et al.* (2008a,b,c)). Yet glacial, thermal, hydromechanical, and hydrogeologic processes are closely coupled. Basal ice-sheet velocity and ice sheet thickness are impacted by subsurface fluid pressures (Flowers *et al.* 2005). In turn, the temperature and fluid pressure at the base of the ice sheet strongly affect subsurface hydrology. Feedbacks between ice sheet dynamics and hydrogeologic processes can be observed in geomorphologic data sets. For example, Grasby & Chen

Table 1 Overview of modeling studies.

Authors	Model description	Study area	Hydraulic BC	Solute transport	Isotope Transport	Permafrost	Lateral grid (km)	Vertical grid (m)	Coupled flow and HM
North America									
Hoaglund *et al.* (2004)	MODFLOW	Saginaw Bay, Michigan	Head, 90% IST,	No	No	No	No details	No details	No
Breemer *et al.* (2002)	MODFLOW	Illinois Basin	Flux, 6 mm year^{-1}	No	No	Low K Layer, 50 km	No details	No details	No
Person *et al.* (2003)	Rift2d	Atlantic Continental Shelf	Head, 90% IST, no Flexure	Yes	No	No	3	75	Yes
Marksamer *et al.* (2007)	Femocp, parallel	Atlantic Continental Shelf	Head, 90% IST, Flexure	Yes	No	No	0.3–22	1–300	Yes
Person *et al.* (2007)	Flexpde	Generic Sag Basin	Head, 90% IST, Flexure	No	No	HTM	5–17	50–600	Yes
Bense & Person (2008)	Flexpde	Generic Sag Basin	Head, 70% IST, Flexure	Yes	Yes	HTM	5–17	50–600	Yes
Cohen *et al.* (2009)	PGEOFE, parallel	Atlantic Continental Shelf	Head, 90% IST, Flexure	Yes	No	Yes, Temp based	2–4	10–100	Yes
Sykes *et al.* (2011)	HYDRO-GEO-SPHERE	Eastern Flank of Michigan Basin, Canada	0%, 30%, 80%, and 100% of ice sheet elevation	Yes	No	Yes	No details	No details	Yes
Lemieux *et al.* (2008a,b,c)	HYDRO-GEO-SPHERE	Canadian Craton	Mixed head/flux; 1–7 mm year^{-1}, Flexure	Yes	No	Yes, Ice Sheet Model Output	25	50–1000	Yes
Europe									
Piotrowski (1997a,b)	MODFLOW	Northwest Germany	Head, 72% IST, no Flexure	No	No	No	1	approximately 50	No
Van Weert *et al.* (1997)	MODFLOW	Northern Europe	Flux boundary condition, constrained by calculated melt rate	No	No	No	No details	No details	No
Frosberg (1996)	Finite Element Code	Barentz Sea	Flux boundary condition, constrained by calculated melt rate	No	No	No	No details	No details	No
Boulton *et al.* (1995)	MODFLOW	Northern Europe	Flux boundary condition, constrained by calculated melt rate	No	No	Low K Layer	No details	No details	No
Iceland									
Flowers *et al.* (2005)	Coupled ice sheet dynamics with groundwater flow	Vatnajokull Ice Sheet	Flux boundary condition	No	No	No Details Provided	1.6 km × 1.8 km	Trans. Based Model	Yes
Antarctica									
Le Brocq *et al.* (2009)	Modified PSU Ice Sheet Model	West Antarctic Ice Sheet	Flux boundary condition	No	No	No Details Provided	20, 5 km	Trans. Based Model	No

Trans. – Transmissivity-based groundwater model (2D flow, vertical averaging).

(2005) note that regional esker systems terminate where relatively permeable carbonate aquifers crop out, suggesting that the basal fluid pressure regime changed abruptly. These authors hypothesize that as glacial meltwater recharged this permeable carbonate aquifer, effective stress would increase (associated with a decrease in pore pressure) and likely reduced ice-sheet basal sliding rates. Grasby & Chen (2005) report that in some instances, end moraines occur at the boundary between the low-permeability Canadian shield and the carbonate rocks in Manitoba and NE Alberta, supporting the notion that a reduction in basal fluid pressure prevented the ice lobe from advancing.

The general form of the groundwater flow equation solved in many recent studies of ice-sheet sediment hydrogeologic interactions is given by

$$\nabla_x \cdot [K \nabla_x h] = S_s \left[\frac{\partial h}{\partial t} - \zeta \frac{\rho_{\text{ice}}}{\rho_f} \frac{\partial \eta}{\partial t} \right] \qquad (1)$$

where t is time, ∇_x is the gradient operator, $\mathbf{K}$ is the hydraulic conductivity tensor, h is hydraulic head, ρ_{ice} is ice density, ρ_f is the fluid density, η is the elevation of the top of the ice sheet, ζ is the loading efficiency (a poromechanical parameter for consolidated, water-saturated media that indicates how much of the ice sheet load is transferred to the fluid phase), and S_s is the specific storage. Many early models of groundwater flow beneath ice sheets (e.g., Breemer *et al.* 2002; Person *et al.* 2003; Hoaglund *et al.* 2004) neglected the effects of hydromechanical ice sheet loading (second term on the right-hand-side of Eq. 1) on groundwater flow. Other studies assumed a loading efficiency of one (e.g., Bense & Person 2008; Cohen *et al.* 2009) for well-consolidated sedimentary rocks implying that all of the weight of the ice sheet is transferred to the fluid. Provost *et al.*'s (1998) study was one of the few studies that varied loading efficiency as part of a model sensitivity study. These authors found that assigning a high loading efficiency reduced sub-ice-sheet recharge and discharge during glacial advance and retreat, respectively. If the ice sheet is frozen at its base, then no liquid water is available for recharge and the ice sheet loading term will act as a mechanical load capable of generating excess pressures within compressible, relatively low-permeability confining units during ice sheet advance. Conversely, rapid glacial unloading can lead to underpressure generation in confining units during and after periods of deglaciation (Vinard *et al.* 1993, 2001; Bekele *et al.* 2003; Bense & Person 2008). The studies of Lerche *et al.* (1997) and Marksamer *et al.* (2007) also included the effects of rapid sediment loading associated with deglaciation on overpressure formation on New England's continental shelf. Few studies have considered loading or unloading effects in multiple spatial dimensions (e.g., Vinard *et al.* 2001). Surface exposure dating studies have been of great benefit in determining the number and extent of glaciations. For example, tills and paleosols from Kansas, Nebraska, Missouri, and Iowa suggest that there were at least seven pre-Wisconsin advances of the Laurentide Ice Sheet that reached a position further south in the mid-continent region, USA, than it did during the Wisconsin Glacial Stage (Roy *et al.* 2004). The Laurentide Ice Sheet reached a similar southern extent (around 39°N) in Missouri during the Early Pleistocene (around 2.4 Ma) (Balco & Rovey 2010). However, reliable information regarding the rates of ice sheet advance and retreat, as well as ice sheet thickness is generally limited to the Wisconsin stage of the Pleistocene (Peltier 2011).

Permafrost and thermal processes

While ice sheet models have represented the coupling between permafrost formation and ice sheet dynamics for some time (e.g., Moores, 1990, Cutler *et al.* 2000), rigorous coupling of thermal and hydrogeologic processes is rare (Lemieux *et al.* (2008a,b,c); Bense & Person 2008). Coupling between permafrost formation and hydrogeologic processes is important because of the effects of freezing on subsurface permeability (Woo 1986; Kleinberg & Griffin 2005; McKenzie *et al.* 2007; Woo *et al.* 2008) and hence recharge. This is substantiated by numerous watershed studies documenting fundamental changes in stream discharge relationships during the past few decades in arctic regions in response to global warming trends (Yang *et al.* 2002; Ye *et al.* 2003; Walvoord & Striegl 2007; McKenzie *et al.* 2007; Bense & Person 2008). Some areas covered by permafrost experienced little or no recharge during the last glacial maximum (Edmunds 2001). Permafrost thickness can vary between 10 and 600 m in high-latitude regions. Permafrost thickness can be spatially variable depending on mean annual surface temperature conditions, land surface orientation, geology, and basal heat flow conditions. As thick ice sheets develop at lower latitudes, permafrost will decay owing to the thermal blanketing effect of ice. Only a few quantitative studies listed in Table 1 have represented heat transport accounting for freezing–thawing. These have done so to represent permeability reduction associated with permafrost formation or because the basins are so deep that fluid properties change in response to elevated temperatures. Beyond the continental ice sheet margins, permafrost development likely promoted deeper, long-distance circulation of glacial melt water beyond the ice sheet toe (e.g., Edmunds 2001). Except for the studies by Lemieux *et al.* (2008a,b,c) and Bense & Person (2008), permafrost has been prescribed as a boundary condition rather than solving for subsurface heat transfer and phase changes. This is likely because the numerical overhead (in terms of spatial and temporal discretization) is large in representing phase changes. Cohen *et al.* (2009) and DeFoor *et al.* (2011) adopted a simplified approach where permafrost was prescribed along the top node if the prescribed land surface temperature dropped below −2°C. Under these conditions, permeability was reduced by three orders of magnitude. This approach did not permit permafrost to propagate down below the top layer of elements. Thus, the thickness of the upper element layer is equal to the potential thickness of the permafrost zone. Such empirical approaches represent a compromise from needing to develop models that can run efficiently over large temporal and spatial scales on the one hand and needing to represent coupling between different subsurface transport processes on the other hand. The validity of this empirical approach of including permafrost has not been assessed to date. We are aware of only one laboratory study that has quantified the relationship between permeability and ice content (Kleinberg & Griffin 2005). That study focused on unconsolidated clastic materials. To the best of our knowledge, laboratory experiments measuring the permeability of fracture bedrock and consolidated sedimentary rocks with different ice contents have not been conducted.

Geomechanical and hydrogeologic coupling

There is close coupling between pore pressures, porosity, and permeability in groundwater systems (Neuzil 2003). To date, however, few quantitative studies have attempted to consider the coupling between excess pore-pressure generation and associated permeability changes during glaciations (Chan *et al.* 2005; Vidstrand *et al.* 2008a,b). There is compelling field evidence such as sand dykes, fractured sedimentary units, and thrust and diapir features, suggesting that pore pressures have modified sediment permeability (Mollard & James 1984; Boulton *et al.* 1995). Within the Illinois Basin, potential evidence for permeability changes can be seen in spatial patterns of salinity data within the New Albany Shale (McIntosh & Walter 2005). Areas where salinity are relatively low in the shale corresponds to the lateral extent of the Laurentide Ice Sheet. High excess pore pressures induced by glacial loading may have caused permeability changes during the last glacial maximum by hydrofracturing shale beds, allowing for the emplacement of relatively fresh groundwater in these confining units (Schegel *et al.* 2011).

One possible approach that could be used to address permeability changes with pore pressure would be to relate the effective stress to porosity and permeability changes as is done by petroleum geoscientists (e.g., Bethke & Corbet 1988):

$$\sigma_e = \sigma_V - P \tag{2}$$

$$\phi = \phi_o \exp[-\beta\sigma_e] \tag{3}$$

$$\log(k) = -a_1 + a_2\phi \tag{4}$$

where σ_e is effective stress, σ_V is downward vertical load of the sedimentary column, P is pore pressure, ϕ is porosity, ϕ_o is porosity at the land surface, β is sediment compressibility, k is permeability, and a_1 and a_2 are lithology-dependent fit coefficients. High pore pressures lead to low effective stress (Eq. 2). Conditions of low effective stress can create conditions of compaction disequilibrium (Eq. 3) where porosity can increase. An increase in porosity can in turn lead to a nonlinear increase in permeability (Eq. 4). Hysteretic behavior can occur during glacial loading/unloading complicating this picture. The values of sediment compressibility are reported to decrease by a factor of five between loading and unloading (Corbet & Bethke 1992). The aforementioned approach also does not address failure mechanisms such as hydrofracturing. Representing rock failure would require representation of additional failure parameters as well as including an effective stress fracture permeability relation (e.g., Walsh 1981).

Brine transport processes

Many sedimentary basins contain saline brines at depths greater than 200–1000 m because of the presence of soluble evaporite formations and evaporated seawater which was emplaced during deposition (e.g., Hanor 1994; McIntosh & Walter 2006). Densities for Na-Ca-Cl brines in sedimentary basins can be on the order of 1200 kg m^{-3}. At depth, this can lead to a stably stratified system (Park *et al.* 2009). The presence of dense brines could modify flow patterns that result from ice sheet loading. Several recent models (e.g., Person *et al.* 2003; Lemieux *et al.* 2008a,b,c; Bense & Person 2008; Vidstrand *et al.* 2008a,b; Marksamer *et al.* 2007; Provost *et al.* 1998; Cohen *et al.* 2009) represented the effects of variable-density groundwater flow and solute transport. One challenge for these studies arises from the need to assign an initial salinity distribution within sedimentary units that have seen many glacial cycles and additionally have undergone complex fluid migration events through geologic time in response to tectonism (Bethke & Marshak 1990; Garven *et al.* 1993). Given the long response time associated with diffusive solute transport, estimating an initial salinity condition will have significant uncertainty. It is likely that current salinity conditions within glaciated basins represent the integrated effects of hydrologic and glacial loading imposed over millions of years. Typically, an initial vertical salinity gradient is assigned as an initial condition using a polynomial function (e.g., McIntosh *et al.* 2010). The study of Vidstrand *et al.* (2008a,b) was unique in attempting to address the issue of salt rejection owing to permafrost formation. However, this was done by imposing a high concentration at the upper boundary rather than coupling permafrost formation and salt exclusion. Few quantitative studies have tried to address brine formation owing to freezing (Panday & Corapcioglu 1991).

Coupling of groundwater flow and ice sheet dynamics

Flowers *et al.* (2005) was the first to couple a two-layer hydrogeologic model to the ice sheet dynamics code of

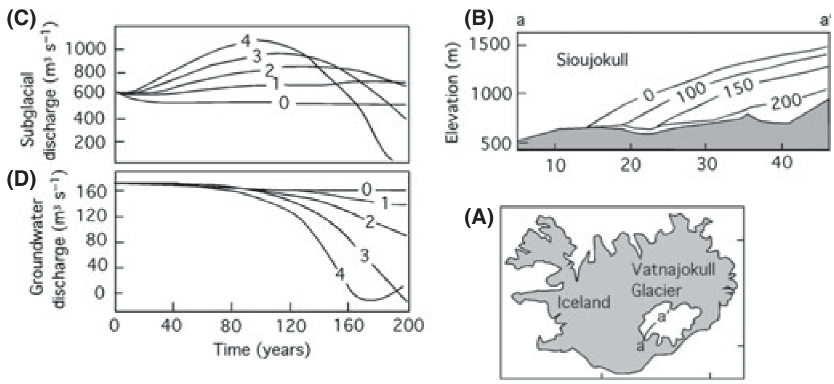

Fig. 2. (A) Base map of Iceland showing location of Vatnajokull Ice Sheet, (B) simulated changes in Vatnajokull Ice Sheet thickness through time along the Sioujokull lobe of the Vatnajokull Ice Sheet, assuming a warming rate of 2°C per century. Numbers in B denote forecasted ice sheet elevations since 2005. The location of a-a′ is shown in A. Changes in net subglacial (C) and groundwater (D) discharge rates through time, assuming different proposed warming rates (0–4°C per century). The numbers on the lines represent the warming rate per century (after Flowers *et al.* 2005).

Marshall *et al.* (2000). These authors used the code to assess the effects of proposed future climate change on ice sheet volume, basal/subsurface discharge, and ice sheet dynamics for the Vatnajokull Glacier, Iceland (Fig. 2A). These authors proposed that basal ice sheet sliding was related to subsurface fluid pressure using the following parameterization scheme:

$$U_{\mathrm{b}} = -B\frac{P_{\mathrm{w}}}{P_{\mathrm{ice}}}\rho_{\mathrm{ice}}\, gH\nabla\eta \tag{5}$$

where U_b is the basal ice sheet sliding rate, B is a basal slide parameter, P_w is basal water pressure, P_{ice} is ice overburden pressure, H is ice sheet thickness, η is height of ice sheet, ρ_{ice} is ice density, and g is gravity. Fluid pressures were calculated using a two-layer hydrogeologic model for the glacial bed and underlying till layer using two linked transmissivity-based flow equations. The two layers were coupled using vertical leakage terms. These authors considered the effects of warming trends between 0 and 4°C per century on ice sheet thickness (Fig. 2B) and subsurface groundwater transport processes (Fig. 2C,D). These authors did not find a strong coupling between basal sliding rates and sub-ice sheet fluid pressures for the Vatnajokull Glacier under conditions of climate change. However, the authors concluded that this is to be expected because this ice sheet is already wet-based and there is no shortage of water. These authors found that during melting, subsurface discharge (Fig. 2D) was about 10 times less than subglacial drainage rates (Fig. 2C). Groundwater discharge rates are predicted to decrease through time in Iceland in response to warming trends. This is attributable to reductions in ice-sheet thickness during ice-sheet retreat (Fig. 2B).

Le Brocq *et al.* (2009) developed a somewhat different parameterization scheme for basal ice-sheet sliding conditions based on water film thickness for the West Antarctic Ice Sheet. These authors hypothesized that the presence of a thin water film between the ice and basal till layer could accelerate basal sliding. These authors allowed the basal traction law in their model to be a function of water film thickness:

$$U_{\mathrm{b}} = -B(d)\rho_{\mathrm{ice}}\, gH\nabla_x\eta \tag{6}$$

where $B(d)$ is the basal traction parameter, d is water film thickness, H is ice sheet thickness, η is height of ice sheet, ρ_{ice} is ice density, and g is gravity. The relationship between basal sliding and water film depth is not known a priori. Le Brocq *et al.* (2009) used InSar measurements of surface ice velocities to estimate U_b. They estimated water film thickness in their model using a steady-state, plane-view transmissivity-based flow equation:

$$\nabla_x \cdot \left[\frac{d^3}{12\mu_{\mathrm{w}}}\nabla\Phi\right] = -M \tag{7}$$

where Φ is the groundwater potential ($\Phi = P_w + \rho_w gz$), P_w is basal water pressure, ρ_w is water density, g is gravity constant, z is bedrock elevation, M is the ice melting rate, and μ_w is water viscosity. This model uses a simple parallel-plate formulation for permeability ($k = d^2/12$) to represent the conductance of the water film. Once these authors obtained a solution for d, they could determine a functional relationship for basal sliding (Fig. 3). They found that the basal sliding parameter needs to be increased by four orders of magnitude to account for ice stream velocities derived from InSar data. Both of these schemes are somewhat ad hoc and have not been tested against experimental data.

Boundary conditions

An important difference between the various published ice-sediment hydrologic models presented in Table 1 is how hydrologic and thermal upper boundary conditions are

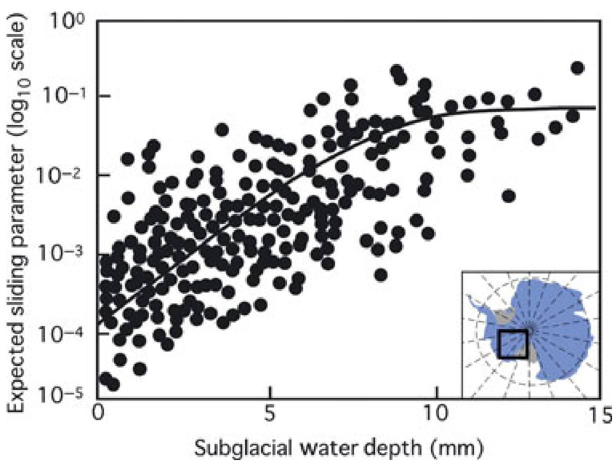

Fig. 3. Relationship between basal sliding parameter (B) and simulated water film thickness (d) for West Antarctic Ice Sheet (From Le Brocq *et al.* 2009).

prescribed. In this section, we compare and contrast different hydrologic, thermal, and geochemical tracer boundary conditions considered in prior studies.

Hydrological boundary conditions

Some quantitative studies have imposed a specified head boundary condition beneath the ice sheet, assuming that the water at the base of the ice sheet is partially supporting the weight of the overlying ice (specified head equals 7/10th the ice sheet height; Piotrowski 1997a,b) or the entire load (specified head equals 9/10th the ice sheet height; Person *et al.* 2003; Hoaglund *et al.* 2004; Person *et al.* 2007; Bense & Person 2008). The presence of esker networks in which liquid water is present in an ice tunnel network supports, in part, this boundary condition. Other groups have specified a flux boundary condition beneath the ice sheet based on estimated rates of basal ice sheet melting (Breemer *et al.* 2002; Carlson *et al.* 2007). Others used a mixed boundary condition (Provost *et al.* 1998; Lemieux *et al.* 2008a,b,c) specifying a flux rate based on a basal melting rate provided that the resulting heads did not exceed 9/10th the ice sheet height. This could occur in areas where a wet-based ice sheet overruns a low-permeability unit. In this case, a head-based condition is imposed equal to 9/10th the ice sheet height. There are strengths and weaknesses to all of the aforementioned approaches. Clearly, specifying a head-based boundary condition can overestimate the amount of available water for recharge, and it is clear from field studies that while subglacial pressures may approach 9/10th the ice sheet height (Engelhardt and Kamb, 1997), heads can fluctuate considerably (Anderson *et al.* 2004) between summer and winter months. However, while there is some uncertainty with prescribing a head condition beneath an ice sheet, there are more uncertainties with prescribing a flux condition (Iverson and Person 2012). The use of a flux condition requires information about the melting rate of a glacier. While basal melting rate estimates vary less than an order of magnitude (1–6 mm per year), recent investigations suggest that surface meltwater can drain to the base of the ice sheet in continental glaciers over 1 km thick (Zwally *et al.* 2002; Joughin *et al.* 2008). The amount of surface melting can vary widely both in space and in time, making the use of a flux boundary condition more uncertain than a head-based boundary condition. Piotrowski (1997a,b) used a prescribed subsurface potentiometric head in his groundwater model at locations where the ice sheet covers the bedrock. Rather than assigning a floating head condition as described (90% of ice sheet thickness), he inferred sub-ice-sheet heads from paleo-porewater pressures estimated from the stress characteristics of the fine-grained sediments overridden by the ice sheet (Piotrowski & Kraus 1997). According to these proxy estimates, the potentiometric surface was on average equal to 72% of the ice thickness.

Another important consideration related to implementing a specified hydrologic head boundary condition is that of lithosphere flexure. The weight of a 3-km ice sheet is great enough to result in lithosphere deflection up to about 0.84 km, assuming full isostatic compensation (approximately 28%; Marshall & Clark 2002). Thus, mechanical loading would lower the magnitude of the imposed head boundary condition. The formation of a far-field flexural bulge (about 10% of maximum downward deflection) would generate uplift beyond the ice sheet margin, potentially creating a small counter flow (Barnhardt *et al.* 1995). While some models have taken this into account (e.g., Lemieux *et al.* (2008a,b,c)), most studies either neglect this effect or have assumed local isostatic compensation of the lithosphere (e.g., Bense & Person 2008).

Thermal boundary conditions

The ice-sediment thermal models that have represented heat transport have generally assumed a specified temperature at the sediment–ice interface. The temperature distribution within an ice sheet is controlled by both conductive and convective heat transfer mechanisms (Hooke 2005), latitude-dependent temperature variations along the ice-sheet surface, and melting owing to frictional heating (Van der Veen 1996). If the temperature at the base of the ice sheet equals/exceeds the pressure melting temperature, melting will occur and a liquid water phase will be present. Person *et al.* (2007) and Bense & Person (2008) assigned a temperature at base of the ice sheet of 0.5°C. DeFoor *et al.* (2011) used one-dimensional, steady-state analytical heat transfer solutions to estimate basal ice temperature. However, these assumptions have not been tested rigorously.

Typically, a basal heat flux is assigned to the bottom of the solution domain based on local continental heat flow.

These typically range between 50 and 70 mW m^{-2} for continental crust (e.g., Lachenbruch & Sass, 1977). The sides of the solution domain are generally assumed to be insulated boundaries ($dT/dx = 0$). Initial conditions for heat transfer are less problematic than for solute transport because of the conductive nature of subsurface heat transport. Generally, an initial vertical geothermal gradient of 30°C per km is assumed. Another important assumption involved in these paleohydrologic models is assigning land surface temperatures through time at different latitudes. Typically, the modern lapse rate (ΔT) is used to vary temperature at a given latitude. Temperature changes through time are determined from paleoclimatic records. Fortunately, ice core and marine isotopic records provide excellent constraints on changes in temperature through time (e.g., Imbrie *et al.* 1984). However, these do not represent local temperature conditions, and thus, some uncertainty exists.

Geochemical tracer boundary conditions

Boundary conditions for solute transport are relatively straight forward. Areas above sea level are assigned a salinity of 0 mg l^{-1}. Areas below sea level are assigned sea water salinity. No-flux boundary conditions are assigned to the sides and base of the domain. Assigning sub-ice-sheet isotopic boundary conditions can be more problematic. A number of studies report $\delta\ ^{18}O$ estimates of basal ice and meltwater to be isotopically heavier than excepted (Knight 1987; Siegel 1991; Ma *et al.* 2004). Laurentide Ice Sheet glacial meltwater in the mid-continent of the United States has $\delta\ ^{18}O$ up to −9‰. This is presumably because of isotopically heavy Pleistocene precipitation (Klump *et al.* 2008; Siegel 1991). Isotopic exchange with rock flour incorporated into the basal ice may also act to increase the $\delta\ ^{18}O$ of melting ice. DeFoor *et al.* (2011) found that to match the observed $\delta\ ^{18}O$ composition in offshore wells on Greenland's southeastern continental shelf, basal meltwater had have an isotopic composition of −3‰. Greenland's average ice sheet oxygen isotopic composition is about −40‰.

Spatial dimensions and discretization of model studies

The level of discretization and spatial dimensions represented in numerical models are linked because of the overhead involved in three-dimensional versus cross-sectional models. The complex geometry of sedimentary basins and ice sheets as well as the continental scale of glaciations as illustrated in Fig. 1 argues for the need to represent ice-sediment hydrologic interactions in all three spatial dimensions, as was done by Hoaglund *et al.* (2004), Lemieux *et al.* (2008a,b,c), and Cohen *et al.* (2009). As noted in Fig. 4, the geometry of sedimentary units (especially the presence of structural arches) can have a significant impact on the emplacement of ice sheet meltwater. Hoaglund *et al.* (2004) constructed a detailed three-dimensional hydrologic model of the Saginaw Bay and Saginaw lowlands, Michigan, USA, using MODFLOW. Limiting their study area to the margin of the Laurentide Ice Sheet and only solving for groundwater flow allowed these authors to use a relatively refined grid and still represent ice-sediment hydrologic interactions in three dimensions. That said, they relied on salinity data to test their model and had to use simple mixing rules that calculated model concentrations. On the other end of the spectrum are the studies of the Waterloo group. These authors were the first to develop a continental-scale ice-sheet groundwater flow model. The computational overhead for this task required Lemieux *et al.* (2008a,b,c) to discretize the Canadian shield and sedimentary basins to a depth of about 10 km using only 10 vertical nodes in a column. As a result, sedimentary

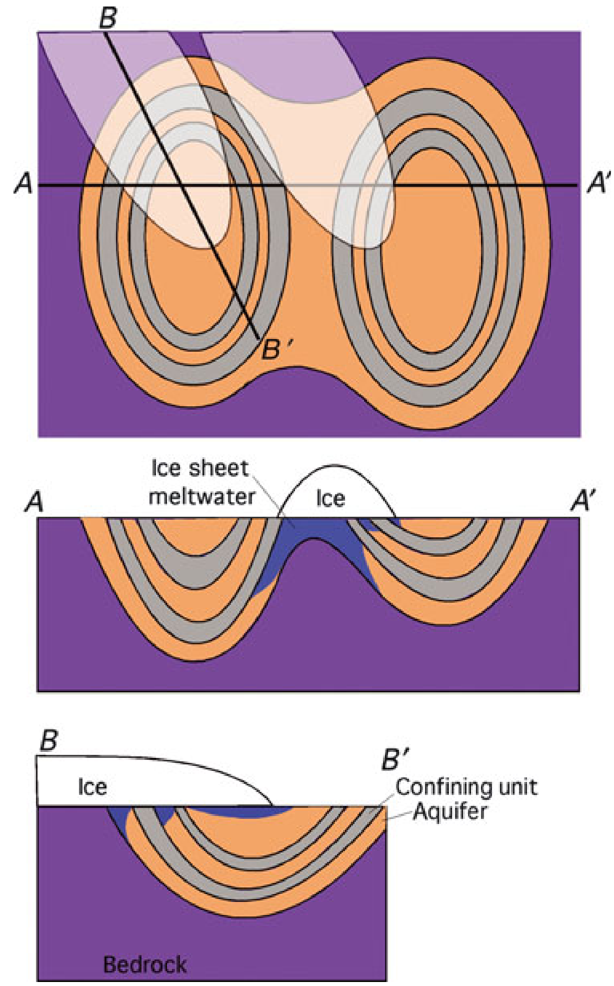

Fig. 4. Schematic diagram illustrating how sedimentary basin and ice sheets geometry impact the emplacement of freshwater. In some situations, sediment-ice hydrologic interactions are highly three-dimensional. The presence of structural arches where aquifers crop out at the surface over long lateral distances can be critical to the emplacement of large volumes of recharge (modified from Mayer & MacQuarrie 2007).

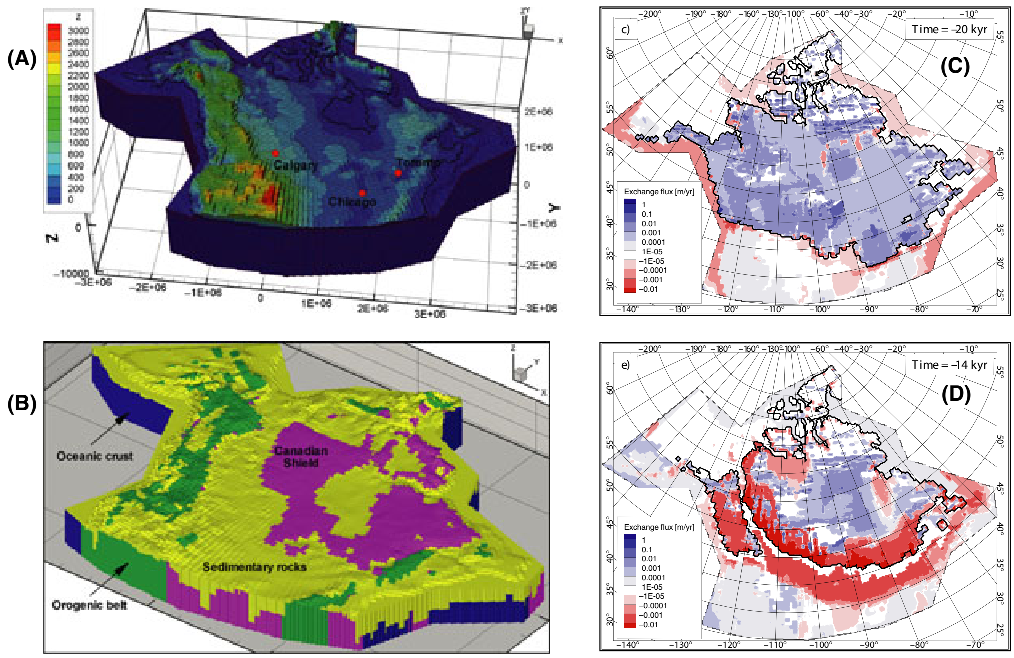

Fig. 5. (A) Topography and (B) stratigraphic grid used in three-dimensional model of Lemieux *et al.* (2008a,b,c). Computed spatial patterns of groundwater fluxes at 20 (C) and 14 (D) KaBP across Canada. Positive numbers (blue pattern) denote groundwater recharge, while negative numbers (exfiltration) denote discharge (red pattern). (from the study of Lemieux *et al.* 2008b).

basins were discretized in the vertical dimension using only a few layers. While these studies were ground breaking in representing the ice–sediment interactions of the entire Laurentide Ice Sheet, this level of discretization using HydroGeoSphere (a serial computer model) precluded representing the hydrologic effects of glaciation on individual confining units and aquifers within sedimentary basins (Fig. 5B). As most of the field data needed to test models are derived from individual aquifers or confining units, there is a need to represent individual hydrostratigraphic units to undertake model validation (Bense & Person 2008). In addition, using coarse discretization invariably results in inaccuracies in solving the solute transport equations. Cohen *et al.* (2009) addressed these problems by turning to high-performance computing and developing a three-dimensional ice-sediment hydrologic model of the Atlantic continental shelf using a parallel code running on 256 processors on a grid of about 1 million nodes (Fig. 6B). Individual confining units and aquifers were discretized to a level of about 10–100 m in the vertical dimension. This allowed them to calibrate their model to well salinity data (Fig. 6C).

MODEL RESULTS

Here, we briefly discuss the results reported by the above-mentioned quantitative models of ice-aquifer hydrologic interactions (Table 2). We first discuss a series of heuristic models developed by Person *et al.* (2007) and Bense & Person (2008). Then, we discuss model studies from North America and Europe and how the results have provided important insights into the occurrence of anomalous fluid pressures, glaciotectonic deformation features, and the occurrence of isotopically depleted groundwaters at the margins of sedimentary basins.

Heuristic models

A series of heuristic two-dimensional models were constructed to assess ice-sediment interactions within idealized sedimentary basins by Person *et al.* (2007) and Bense & Person (2008). The idealized sag basin represented in these studies is probably applicable to geologic settings such as the Michigan Basin, USA, or perhaps the Williston Basin of North Dakota, USA. These authors used the commercial modeling package Flexpde© and a modified version of Rift2d to construct their ice-sediment hydrologic models (see Table 1). The primary goal of these models was to explore the coupling between permafrost, groundwater flow, and solute transport patterns. A secondary goal was to determine whether or not anomalous pressure patterns were preserved in aquifers and the extent to which isotopic and geochemical tracers associated with glacial meltwater could be preserved at the basin margins. These authors showed that recharge rates are elevated during glaciation and that remnant overpressures and underpressures are still preserved 21 000 years after glaciation (Fig. 7D). They also showed that preserved ‘fossil’ head anomalies are up

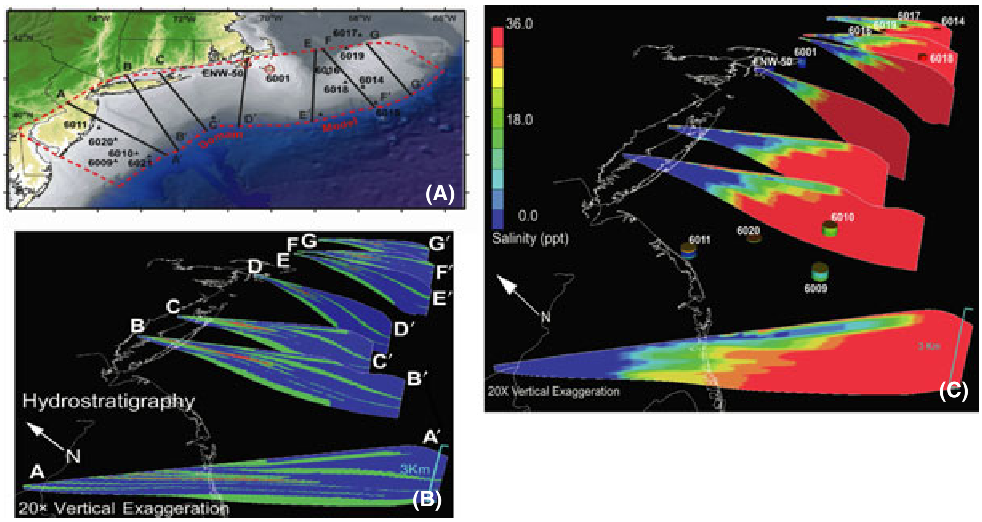

Fig. 6. (A) Aerial footprint of three-dimensional paleohydrologic model (dashed purple line) along Atlantic continental shelf in New England and location of cross sections and wells. (B) Cross-sectional transects depicting hydrostratigraphic units represented in three-dimensional model: silt-clay (blue), fine sand (red), and medium to coarse sand (green) units. (C) Simulated concentrations (in parts per thousand) along cross sections extracted from the three-dimensional finite element model of the Atlantic Continental Shelf. Cylinders depict concentration of offshore AMCOR, COST, and ODP wells (well radii not to scale). (From Cohen *et al.* 2009).

Table 2 Overview of modeling study results.

	Recharge			
Authors	Ice sheet	No ice sheet	Head anomaly	Freshwater penetration
North America				
Hoaglund *et al.* (2004)	No details	No details	No details	No details
Breemer *et al.* (2002)	6 mm year^{-1}	38 cm year^{-1}	No details	No details
Person *et al.* (2003)	0.33 m year^{-1}	0.15 m year^{-1} meteoric recharge	No details	approximately 100 km lateral distance
Marksamer *et al.* (2007)	Ice-sheet recharge	Meteoric recharge	Excess (+8 m, overpressure) and negative heads (underpressure)	approximately 100.0 km lateral distance
Person *et al.* (2007)	No details	No details	Excess and negative heads	No details
Bense & Person (2008)	No details	No details	Excess (+120 m) and negative heads (−50 m)	<1.0 km depth,
Cohen *et al.* (2009)	2.0 m year^{-1}	m year^{-1}	Excess head (+700 m)	approximately 100.0 km lateral distance
Lemieux *et al.* (2008a,b,c)	1–7 mm year^{-1}	No details	Predicted remnant excess head	No details
Europe				
Piotrowski (1997a,b)	Flux, 23–30 m^2 day^{-1}	Flux, 10–50 m^2 day^{-1}	Likely to produce hydraulic lifting of the ice sheet	Penetrated Quaternary sediments of thickness (approximately 200 m)
Van Weert *et al.* (1997)	Ice sheet recharge	No details	Formation of high pressures	No details
Frosberg (1996)	5–20 mm year^{-1}	No details	High sub-glacial water pressure in the aquifers	No details
Boulton *et al.* (1995)	25 mm year^{-1}	No details	Presence of very high and very low effective pressure	No details

to hundreds of meters (100 m of pressure head equals 1 MPa). This study also showed that in select locations, lower confining units were underpressured and overlying confining units were overpressured. This study, like many others, noted reversals in groundwater flow patterns and found that permafrost had little effect on vertical infiltration. This is because permafrost disappears underneath wet-based ice sheets on time scales of less than 10 000 years. On the other hand, permafrost was important in focusing shallow groundwater flow out beyond the ice sheet margin. Salinity (Fig. 8A) and oxygen-18 isotopic (Fig. 8B) patterns computed in this study showed that the

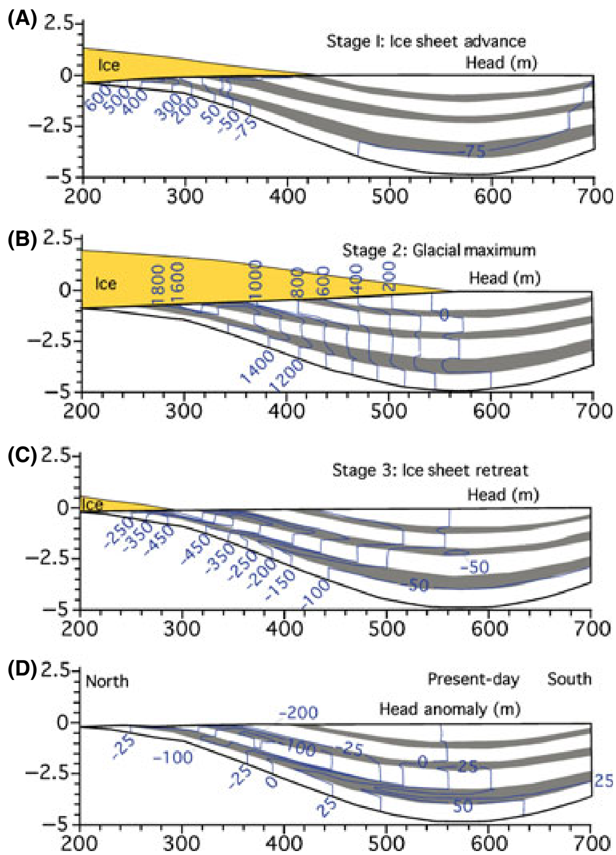

Fig. 7. Computed transient heads during one glacial cycle (A-C) as well as anomalous heads after one glacial cycle (31 k years) for an idealized sedimentary basin. Stages I-III represent ice sheet advance (I), glacial maximum (II) and ice sheet retreat (III). The anomalous heads (D) were computed by subtracting steady-state-computed heads from transient head patterns which include glacial loading. Note that there are regions of both overpressuring and underpressuring at different depths. The light gray pattern denotes aquitards.

vertical depth of fresh water penetration in the basin was up to 500 m per glacial cycle. The influx of fresh, isotopically depleted glacial melt water was greatest in the most permeable (shallowest) aquifer rather than the deepest (least permeable) aquifer whose recharge area was ice covered the longest.

North America models

One of the few quantitative studies to consider the effects of glaciations on anomalous pressures in North America was presented by Bekele *et al.* (2003) within the Alberta Basin. These authors used the hydrogeologic model Rift2d (Person and Garven, 1992) to try to reconstruct 35 subhydrostatic pressure measurements collected by Parks & Toth (1995) from 24 boreholes near Buck Lake, Alberta. Maximum underpressures measured were as high as 6.5 MPa at

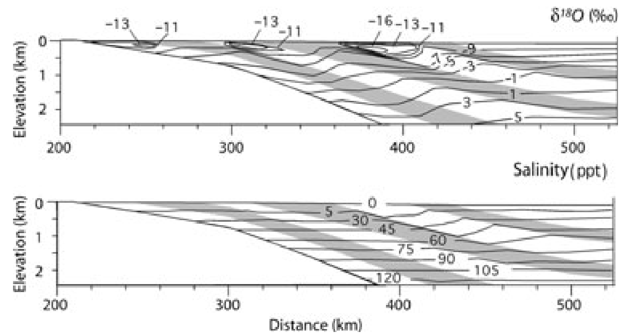

Fig. 8. Computed salinity (parts per thousand; ppt) and oxygen isotopic composition (per mille) after one cycle of glacial loading. The lines denote the boundaries between sedimentary rock aquifers (Aq) and confining units. Beneath the layers is bedrock (from Bense & Person 2008).

a depth of about 1000 m. These authors tested both erosional unloading and glacial unloading hypotheses. They noted that typical sediment erosion rates were relatively low (approximately 0.034 mm year^{-1}; Steiner *et al.* 1972), whereas the ice sheet ablation/melting rate was orders of magnitude higher at about 100 mm year^{-1}. However, using representative permeability and compressibility data, these authors concluded that even during rapid ice sheet retreat, these unloading rates where not sufficient to explain the observed 6.5 MPa underpressure measurements.

More recently, Lemieux *et al.* (2008a,b,c) consider the hydrologic consequences of ice-sediment hydrologic interactions at the continental scale in three spatial dimensions (Fig. 5C,D). These authors used HydroGeoSphere to assess the effects of the Laurentide Ice Sheet on groundwater flow dynamics, anomalous head conditions, and recharge during the late Pleistocene. These authors calibrated their model comparing simulated and observed salinity data from the Canadian Shield and the Michigan Basin. One important finding from Lemieux *et al.* (2008b) is that during periods of ice sheet advance, there is a net influx of meltwater into the continental crust (Fig. 3C). Groundwater discharge (exfiltration) was more important during ice sheet retreat (Fig. 3D). Lemieux *et al.* (2008b) as well as Bense & Person (2008) also predicted the presence of remnant excess heads (up to 1500 m) within the Canadian Shield at high latitudes. Unfortunately, there is a lack of pore-pressure data at depth across the Canadian Shield to substantiate these results. The most important finding of this study was that a significant fraction of basal meltwater (15–70%) from the Laurentide Ice Sheet could have infiltrated the Canadian Shield and sedimentary basins of North America. On average, about 40% of meltwater infiltrated into sedimentary and crystalline rocks.

Breemer *et al.* (2002) developed a series of cross-sectional hydrologic models along the NE–SW traverse of the Lake Michigan Lobe across the Illinois Basin, using the model MODFLOW. The model reconstructions focused on Illinoian glaciation because of its large spatial extent. The goal of this study was to assess the development of basal pore-pressure conditions beneath the ice sheet and to better understand sliding and basal sediment deformation mechanisms beneath this ice lobe. Their model was calibrated to present-day head observations across the Illinois Basin. No apparent correction was made for density effects associated with basinal brines. Solute transport was not considered in this study. These authors noted that Illinoian glaciation resulted in reversals of groundwater flow directions when compared to Holocene subsurface flow patterns. These authors also found that using a flux-based boundary condition resulted in computed heads far higher than could be supported by the overlying ice sheet. To compensate for this, they added a permeable layer at the ice-sediment interface to represent a thin film of water which acted as a thin esker system. Two important points are worth noting. The hydrostratigraphic framework model these authors employed did not allow the Paleozoic carbonate and sandstone aquifers to outcrop near the margins of these basins, thus limiting the amount of subsurface inflow and promoting high heads in shallower layers. Second, the constant-density simulations presented in this study did not consider the dense brines in the Illinois basin. Thus, groundwater flow dynamics may have been misrepresented.

Carlson *et al.* (2007) studied basal pressure conditions beneath the James Bay Lobe as it crossed the Williston Basin. This study attempted to refine the work of the Breemer *et al.* (2002) by developing a more physically based conceptual model of the basal drainage system at the ice-sediment interface. These authors developed cross-sectional representations of groundwater flow using MODFLOW and applied a 6 mm year^{-1} recharge rate to represent sediment-ice hydrologic interactions. To bleed off high heads resulting from the specified flux boundary condition, the authors developed an 'esker' layer varying in thickness from 0.1 to 1 m at the sediment-ice interface. This model was calibrated to modern head measurements prior to reconstructing last glacial maximum conditions. These authors also noted important changes in flow directions resulted from ice-sheet loading. Their study did not attempt to match salinity or isotopic data sets reported by Grasby *et al.* (2000). The point made regarding the Breemer *et al.*'s (2002) study applies here as well: the Williston Basin contains brines, and a constant-density-flow model will not properly represent groundwater flow dynamics in deep, saline formations.

Hoaglund *et al.* (2004) developed three-dimensional models of present-day and last glacial maximum hydrogeologic conditions within the Michigan Basin in Wisconsin. Their study covered both the Saginaw Bay area of Lake Huron and the Saginaw Lowlands. They used the groundwater modeling package MODFLOW. The primary objective of the models was to explain the presence of isotopically light, saline groundwater observed within the Saginaw Lowland aquifers of Michigan to a depth of 300 m (Kolak *et al.* 1999; Klump *et al.* 2008). They did not consider solute transport directly. They used simple mixing model calculations to compare observed geochemical anomalies to their model results. Hydrogeologic parameters used in this model were calibrated using state-wide heads and stream fluxes. These authors noted a head discrepancy in their model calibration between −7 and 18.9 m (computed minus observed heads). It does not appear that they took into account the effects of fossil heads (Bahr *et al.* 1994) in their model calibration, but these discrepancies are small in comparison with the imposed ice sheet load. Probably, the most relevant hydrologic finding from their study was the documentation of flow direction reversals associated with glaciations. They also concluded that a mixing event involving subglacial recharge could have produced the groundwater chemistry currently observed in the Saginaw Lowlands area. McIntosh *et al.* (2012) developed a cross-sectional, variable-density flow, heat and solute mass transport model (CPFLOW) along a north–south transect across the Michigan Basin. The model was run for 1 million years with no glacial loading in an attempt to approach steady-state conditions using modern topography. Then, one glacial advance and retreat was imposed over a period of 17 000 years. Glacial loading was imposed using a polynomial function to represent ice sheet geometry along the upper boundary. Recharge rates during glaciations were up to 20 times higher than modern conditions. Glacial meltwater penetrated laterally by about 60 km into a series of Paleozoic confined aquifers. Model results were compared to observed anomalous pore pressures (Bahr *et al.* 1994), salinity patterns (McIntosh & Walter 2005), noble gas (Klump *et al.* 2008; Ma *et al.*, 2004), and environmental isotopic composition of groundwater samples (McIntosh & Walter 2005). Penetration of freshwater along the northern margin of the Michigan Basin extended to a depth of about 1000 m within the Cambrian Mount Simon Formation (a sandstone aquifer). Sykes *et al.* (2011) developed detailed one-dimensional two-phase (natural gas), basin-scale cross-sectional and site-scale (about 200 km by 140 km) three-dimensional paleohydrologic models of the Michigan Basin to assess the viability of a proposed medium-level nuclear waste site on the eastern flank of the Michigan Basin in the Municipality of Kincardine, Ontario, Canada (Fig. 9A,B). The proposed repository will be sited at a depth of about 680 m below 200 m of ultralow-permeability Ordovician shale's (Fig. 9D). Detailed hydraulic measurements collected along a series of test holes revealed a complex pattern of underpressures in the mid-Ordovician shales and modest overpressures within basal Cambrian sandstone (Fig. 9C).

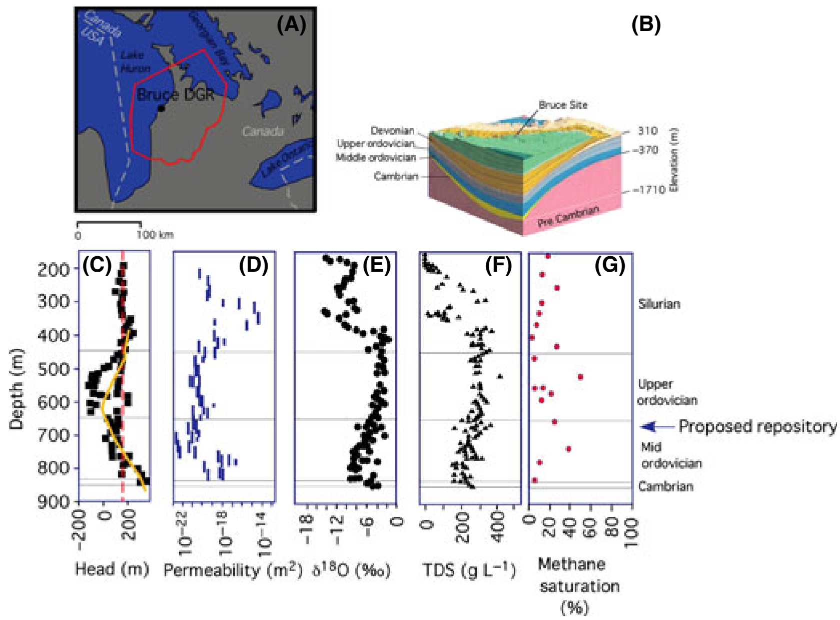

Fig. 9. (A) Location of Bruce site on eastern edge of Michigan Basin in Ontario, Canada. (B) Idealized three-dimensional block diagram of stratigraphy at proposed waste site. (C) Observed (squares) and computed heads. Computed heads are from a one-dimensional TOUGH2 two-phase flow model simulation that considered a methane gas phase. Anomalous low heads are thought to be related to capillary effects associated with the presence of a methane phase. (D) Measurements of permeability (m^2). Observed $\delta^{18}O$ (‰) (E), salinity (F), and methane saturation (G). (after Sykes *et al.* 2011).

These authors concluded that the underpressures were not attributable to glacial loading but rather were related to capillary effects associated with the presence of a methane gas phase (Fig. 9G). The effects of glaciations were quantified using both cross-sectional and three-dimensional models (FRAC3DVS-OPG). The effects of two-phase flow and pore pressures were assessed using a one-dimensional multiphase model (TOUGH2). Simulated effects of glaciation suggest that emplacement of isotopically depleted freshwater did not extend to beneath the mid-Ordovician shales (400 m depth; Fig. 9E,F). They concluded that the greatest effects of any future glaciations would be to increase pore pressures within the repository zone.

A large quantity of freshwater (<5 parts per thousand total dissolved solids, or ppt TDS) is sequestered within the permeable sands and carbonate rocks along the Atlantic continental shelf of the USA (wells in Fig. 6A,C) from New Jersey to Massachusetts (Hathaway *et al.* 1979). This freshwater was likely emplaced during Pleistocene sea-level low stands when the continental shelf was exposed to meteoric recharge and overrun by the Laurentide Ice Sheet at high latitudes from Maine to Long Island, NY, during the last glacial maximum (Uchupi *et al.* 2001). Person *et al.* (2003) and Marksamer *et al.* (2007) evaluated the relative importance of sub-ice sheet recharge versus meteoric recharge during the Late Pleistocene using the cross-sectional basin simulator Rift2d. These authors considered both variable-density flow and solute transport. Permafrost was not considered in these studies. These authors tested their hypotheses using salinity and anomalous fluid pressure data from the offshore wells. They found that sea-level changes and subsequent meteoric recharge alone could not account for the observed distribution of freshwater. By imposing ice sheet loading and recharge beneath pro-glacial lakes, they were able to drive freshwater far offshore (Person *et al.* 2011). More recently, Cohen *et al.* (2009) developed a three-dimensional high-resolution paleohydrologic model of groundwater flow and solute transport for the continental shelf from New Jersey to Maine over the last 2 million years using a parallel hydrologic model (*GW_ICE*) and represented both glaciated and unglaciated areas (Fig. 6). During the last glacial maximum, predicted recharge/discharge rates near the toe of the ice sheet reached about 2 m $year^{-1}$, which is about 4 times higher than modern estimates for Nantucket Island (Knott & Olimpio 1986). Models for which the effect of the ice sheet was removed predicted higher salinity concentrations in regions proximal to the ice-sheet toe. On the other hand, during ice-sheet loading within the shallow Miocene sand unit off New Jersey, fresh to brackish water is transported out to 100 km offshore. This is because these shallow sands created submarine springs, which were active during sea-level low stands. These authors also calculated the volume of sequestered freshwater in New England to be 1300 km^3. To put this freshwater estimate into perspective, during the 20th century, the cumulative volume of freshwater withdrawn from the High Plains Aquifer, USA, is estimated at about 270 km^3 (Konikow 2002).

European models

A series of quantitative reconstructions of ice-sediment hydrologic interactions and glaciations were carried out in Europe in the mid-1990s (Boulton & Caban 1995; Boulton *et al.* 1995; Van Weert *et al.* 1997). These studies were motivated, in part, by the search for suitable sites to host medium-level to low-level radioactive waste repositories. These 'first-generation' studies focused on Scandinavia and the southern parts of the North Sea Basin. Boulton *et al.* (1993) discussed the possibility of enhanced ice-sheet-derived recharge in permeable sedimentary basins and argued that areas of relatively low permeability limited the amount of groundwater movement. This, in turn, played a crucial role in the development and morphology of subglacial channels. Boulton *et al.* (1993) strengthened and illustrated these arguments by reconstructing numerical models of a groundwater flow system using MODFLOW in the northern part of the Netherlands during the previous glacial period (approximately 130–190 ka BP). They concluded that during glaciations, groundwater flow directions in sedimentary aquifers underlying ice sheets can be strongly altered compared to present-day conditions and that fluid fluxes can be more than two orders of magnitude larger. This is probably the largest estimate of changes to groundwater recharge owing to glaciations reported in the literature.

A regional study in the North Sea Basin and Scandinavia based on the same principles as laid out in Boulton *et al.* (1993) was reported in a second set of papers by Boulton *et al.* (1995) and Boulton & Caban (1995). A continental-scale model based on MODFLOW was presented in these studies extending from the low-permeability basement rocks of Sweden, across the North Sea Basin and then toward the permeable sedimentary aquifers of the Netherlands and Germany. Boulton *et al.* (1995) again concluded that the regional groundwater flow system is completely reorganized during glaciations and that fluid fluxes increased by several orders of magnitude. Although no heat transport is considered in these models, the role of permafrost is briefly discussed in sediment mobilization where the effective stress becomes negative. Boulton & Caban (1995) discussed in some detail the implications of their findings for the development of glaciotectonic structures and moraine formation. In the above set of studies, the basal melt rate of the glacier is calculated from a simple ice-sheet model, which is then used to constrain the hydraulic head underneath the ice load. In this approach, the authors evaluate whether the underlying sediments would have been permeable enough to discharge all the available sub-glacial meltwater. If the hydraulic head becomes so large that the effective pressure is reduced to zero, it is assumed that the excess water will be discharged through developing subglacial tunnel systems.

Van Weert *et al.* (1997) developed a three-dimensional groundwater flow model similar to that of Boulton *et al.* (1993). In this study, the entirety of the North Sea Basin has been considered including flanking areas in Scandinavia, England, the Netherlands, Denmark, and Germany. The six simulations presented by Van Weert *et al.* (1997) are steady-state hydraulic head calculations of representative stages during the last glacial period and mostly confirm the previous smaller-scale studies by Boulton *et al.* (1993). Frosberg (1996) presents a suite of steady-state cross-sectional hydrogeological models for interglacial and glacial situations across a sedimentary basin SE of Spitsbergen. A section about 200 km in length and 4 km in depth was modeled using a fluid flux boundary condition where the ice sheet is present. The results and conclusions drawn from these models are very similar to those presented in Boulton *et al.* (1995).

Piotrowski (1997a,b) presented a MODFLOW-based numerical model of groundwater dynamics associated with the late Pleistocene advance of the Scandinavian ice sheet in northwestern Germany during the last glacial maximum. Piotrowski (1997a,b) applied a fixed hydraulic boundary condition (about 70% of reconstructed ice-sheet thickness) in his model simulations based on field observations by Piotrowski & Kraus (1997). The surface hydraulic conditions for this model were further constrained with calculations of basal meltwater production rates and compared with field evidence of meltwater erosional features such as tunnel valley development across the area. Although Boulton *et al.* (1995) argued that the North Sea basin aquifers where sufficiently thick and permeable to drain the available subglacial melt water, Piotrowski (1997a,b) came to a different conclusion that the transmissivity of sedimentary aquifers in Northern Europe was not everywhere sufficient to drain all the meltwater produced underneath ice sheets during the last glacial periods over Europe.

Heathcote *et al.* (1996) presented a suite of cross-sectional and three-dimensional models Sellafield, NW England, to interpret borehole data from 18 wells (hydrogeological and hydrochemical data to depth up to 2000 m). These models were used to address the present-day hydrogeological conditions at the site, and no attempt was made to use them to evaluate how conditions might have changed during glacial times. However, one main conclusion drawn by Heathcote & Michie (2004) was that at the location of the Sellafield site, subglacial recharge must have been relatively limited and no emplaced meltwater was found. This could suggest an important role of permafrost in limiting infiltration.

Provost *et al.* (1998) developed a late Pleistocene hydrologic model that included the last cycle of glaciation, variable-density groundwater flow, and solute transport along a 1500-km NW–SE transect across Norway and Sweden using the US Geological Survey model SUTRA and was among the first to represent glacial-loading effects. A salinity source term was used to represent geochemical processes associated with deep brine formation. Flexural adjustments to the lithosphere were not considered. A sensitivity study was carried out in which loading efficiency, permeability anisotropy, and porosity were varied. One of the most important findings of this study was that varying the loading efficiency between 0 and 1 changed the position of the freshwater saltwater transition zone by hundreds of meters.

Vidstrand *et al.* (2006) considered the effects of cryogenic brine formation on flow and solute transport. A generic solution domain was used in this study although conditions closely resembled the Swedish Aspo hard-rock laboratory site. Permafrost formation was not represented formally. Rather, brine exclusion was represented indirectly by specifying a high concentration along the upper boundary. Fracture bedrock permeability was assigned in the model using a Poisson point process. Results indicate that brine fingers formed in near-surface permeable zones and that downward solute transport occurred faster than the rate of permafrost formation. What is unclear is whether brine exclusion effects have a large impact of flow dynamics during permafrost formation or not.

Vinard *et al.* (1993, 2001) studied the mechanisms responsible for the existence of large (4.8–7.1 MPa; diamonds and circles in Fig. 10C) underpressures within a tight, 2-km-thick marl unit (red unit in Fig. 10B) in central Switzerland near Wellenberg (Fig. 10A). Detailed hydraulic testing reported by Vinard *et al.* (1993, 2001) indicated that the permeability of these confining units was between 10^{-18} and 10^{-20} m^2. The underpressures are thought to be caused by both mechanical erosion and glacial unloading. These authors developed one- and two-dimensional poromechanical models using ABAQUS to try to reconstruct the pore-pressure distribution for two test wells (SB1 and SB3) where detailed pore pressures were collected. They varied hydrologic and elastic properties of the Marl as well as erosion and deglaciation rates, and boundary conditions (drainage at the base of the marl) as part of a numerical sensitivity study. Vinard *et al.* (2001) concluded that both erosion and glacial unloading were required to account for the observed underpressures. While the models were able to match the magnitudes of underpressures, they were not able to match the spatial patterns closely (lines in Fig. 10C).

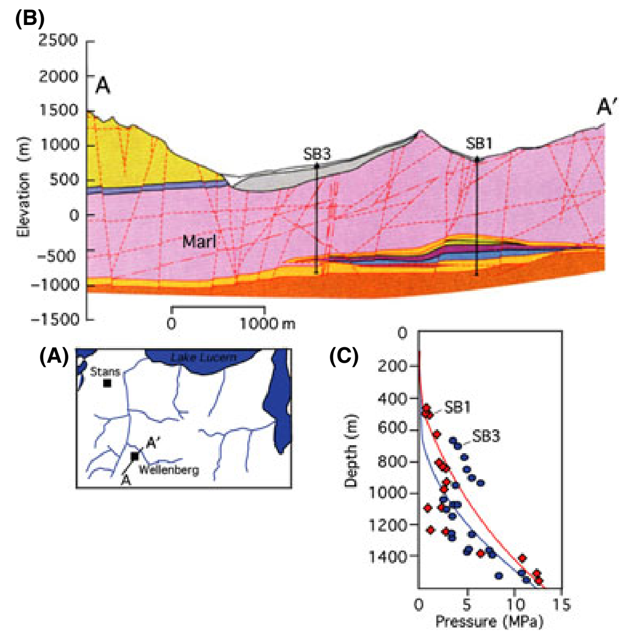

Fig. 10. (A) Basemap showing location of Swiss Wellenberg site for nuclear waste isolation studies. (B) Geologic cross section depicting thick unit of low-permeability Marls. The location of two test wells (SB1, SB3) where detailed pore-pressure measurements were taken. (C) Comparison between observed pore-pressure data from SB1 (red diamonds) and SB3 (blue circles) and poromechanical model ABUQUS (red and blue lines, respectively), assuming a combination of erosional and glacial unloading with a basal drain boundary condition (from Vinard *et al.* 2001).

DISCUSSION AND CONCLUSIONS

Summary of model findings

The models discussed herein provide important insights into past hydrogeologic conditions during periods of glaciation. For nuclear waste repository performance assessments, these models are required in coordination with field investigations to better instill an understanding of the systems response to glacial loading. What is consistently suggested by these models is that infiltration rates in ice-covered areas can potentially increase between two and six times the modern conditions. The study of Lemieux *et al.* (2008a) suggested that about 40% of basal meltwater generated beneath the Laurentide Ice Sheet recharged sedimentary and crystalline rocks across North America during the last glacial maximum. Some of these models were able to account, in part, for observations of extreme underpressures in tight, thick confining units (e.g., Vinard *et al.* 2001; Bekele *et al.* 2003). Models that included solute transport processes suggest that glacial meltwaters could penetrate to depths of hundreds of meter along the margins of sedimentary basins over a glacial cycle. While model results in many instances were found to be consistent with the observed data sets, uncertainties in paleohydrologic and geochemical boundary and initial conditions remain.

Glacial loading and anomalous pressure generation

Many mechanisms have been proposed to account for anomalous pore pressures in confining units and reservoirs within sedimentary basins including erosional unloading (Neuzil & Pollock 1983), osmotic effects (Neuzil 2000), petroleum generation (Bredehoeft *et al.* 1994), and diagenesis (Vrolijk *et al.* 1991). Attempts to link observations of anomalous pressures in fine-grained confining to glacial loading have been equivocal at best. Studies in Switzerland (Vinard *et al.* 1993, 2001), and the Alberta Basin, Canada (Bekele *et al.* 2003) both concluded that glacial unloading was not the sole mechanism responsible for observations of extreme underpressures. Sykes *et al.*'s (2011) study of underpressures in tight confining units along the eastern flank of the Michigan Basin called into question the glacial unloading mechanism. These authors proposed instead that capillary effects associated with the presence of a gas phase as a more likely explanation for the observed underpressures. Other studies have identified overpressures within Michigan Basin aquifers (Bahr *et al.* 1994). That study called on geochemical pressure seal formation to explain the anomalous pressure data within the Michigan Basin. We conclude that it is unlikely that any single mechanism such as glacial loading is responsible for anomalous pressures observations within a sedimentary basins. Pore-pressure data should be used in combination with geochemical/isotopic/noble gas tracer data as a means of identifying glacial-loading effects. Analysis of pressures beneath modern ice sheets may help to provide additional ground truth of the effects of glacial loading. However, these data are very expensive to obtain.

Because there is considerable uncertainty in the permeability of clastic units (Gelhar 1993), sensitivity studies in which different realization of permeability of confining units and aquifers are varied systematically could help bracket the range of possible outcomes of pore-pressure distributions and fluid migration patterns. Not taking into account scale effects on core permeability measurements may also bias model results. Many early quantitative studies of the hydrologic impacts of glaciations did not include a glacial-loading source term in their governing groundwater flow equation (e.g., MODFLOW). As noted by Provost *et al.* (1998), this will tend to overestimate recharge and meltwater penetration into the subsurface because the weight of the ice sheet is not 'felt' at depth.

Geochemical and isotopic tracers

Comparison between computed and observed geochemical (salinity) and environmental isotopic (^{2}H, ^{18}O), and noble gas tracers as well as groundwater residence times (^{4}He, ^{14}C) represents probably one of the best means to provide ground truth for the paleohydrologic models discussed above. 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 years before present) is a good indicator of ice sheet recharge. We argue that greater certainty in the application of geothermal tracers will come from using multiple tracers (noble gas, environmental isotopes, and salinity). However, these tracers can sometimes provide ambiguous results. For example, analysis of noble gas–derived temperatures and ^{14}C ages of pore fluids presented by Ma *et al.* (2004) for the Michigan Basin suggest that the presence of Late Pleistocene glacial meltwater is more enriched (−8.6 to −10.3‰) than that of the typical ice sheet end member composition (−20‰; Remenda *et al.* 1994). This could be explained, in part, by mixing processes. Finding highly depleted oxygen-18 groundwaters in aquifers is rare (McIntosh *et al.*, 2012). Collection of isotopic data beneath modern ice sheets will help to constrain the variability that could be expected in assigning isotopic boundary conditions for hydrogeologic models.

Effects of basin geometry, boundary, and initial conditions on model results

The choice of the boundary conditions beneath ice sheets (i.e., flux verses head), representation of permafrost effects on subsurface permeability, the magnitude of aquifer/confining permeability, and sedimentary basin geometry (i.e., that whether or not sedimentary basin aquifers crop out in regions overrun by glaciers) are likely the most important factors controlling the calculated distribution and maximum depth of chemical and environmental isotopic tracers. An imposed flux boundary condition controlled by basal melting rates (about 5 mm year^{-1}) will reduce freshwater penetration as an ice sheet overruns a permeable aquifer when compared to a head-based upper boundary condition. The importance of basin geometry on simulated freshwater penetration can be seen by comparing model results from different paleohydrologic studies of the Michigan Basin. Comparison of simulated freshwater penetration depths along the eastern (Sykes *et al.* 2011, 400 m) and northern (McIntosh *et al.*, 2012, 1000 m) margins of the Michigan Basin illustrates how basin stratigraphic configuration influences model outcome. The basal Mount Simon sandstone cropped out along the northern margin of the Michigan Basin in the model presented by McIntosh *et al.* (2012). This allowed for the ice sheet to directly access this basal aquifer. In the three-dimensional model of the Bruce site on the eastern flank of the Michigan Basin, this basal Cambrian aquifer did not crop out and freshwater did not extend below the mid-Ordovician shales (Sykes *et al.* 2012). We have noted that permafrost is quickly removed in low-latitude regions during ice sheet advance once an area is covered by a thick layer (1000 m) of ice. However, at high latitudes, permafrost may never disappear owing to low surface temperature conditions. Thus, permafrost's effects on recharge (and hence tracer distributions) may be very site specific.

All paleohydrologic model reconstructions discussed herein suffer from uncertainty regarding choices of imposed boundary and initial conditions. Many of these models represent broad range of coupled processes and with many degrees of freedom that can lead to solution non-uniqueness. This has led some to suggest that such models are of little use (Morner 2001). There is considerable uncertainty regarding the spatial extent and duration of glaciations to the Wisconsin Glacial Stage. How many times has a given region been overrun by ice sheets during the Pleistocene? How many times did the ice sheet readvance within a glacial cycle? Insights on the waxing and waning of ice sheets are available from marine and ice sheet isotopic records (e.g., Imbrie *et al.* 1984), but these represent global averages and cannot reliably predict the behavior of individual continental ice sheets. Glacial rebound analysis considering mantle flow and lithosphere flexure can provide more local information on changes in ice sheet thickness during the Late Pleistocene (Peltier 2011). As noted in section Representation of glacial processes, surface exposure dating has documented at least seven glacial cycles across the mid-continent of North America.

The initial salinity conditions used in the paleohydrologic models are difficult to accurately constrain. It is likely that salinity conditions in sedimentary basins are never in equilibrium with present-day climatic forcing. McIntosh *et al.* (2012) used an initial empirical salinity–depth relationship to spin up their model for 1 million years without ice sheet forcing. This type of approach is necessary. Many quantitative studies have represented just one glacial cycle. It is unclear, however, how many glaciations have impacted porewater chemistry.

Hydrogeologic models and gravity-based estimates of changes to Greenland's ice sheet thickness

Ice-sheet hydrogeologic models have primarily been applied to hind-cast past conditions during the Late Pleistocene. They could also be of benefit to the Earth science community in assessing the potential impacts of future climate change on subsurface hydrologic conditions beneath the Greenland ice sheet. Today, large areas of Greenland and Antarctica are underlain by permafrost with frozen beds and likely have deep water tables owing to the absence of recharge. As these ice sheets melt under conditions of future climate change, permafrost beneath ice sheets could decline and sub-ice-sheet infiltration could cause large changes in water table elevation. Greenland's continental ice sheet is widely expected to disappear over the next 200 years (Parizek & Alley 2004). Quantitative estimates of changes in ice sheet thickness have benefited from satellite-based gravity measurements. Hydrogeologic models could be used to improve GRACE satellite estimates of ice-sheet melting and runoff on Greenland (e.g., Velicogna & Wahr 2005). Gravity-based estimates of ice-sheet thickness changes require corrections for changes in subsurface water storage (Swenson & Wahr 2003; Rodell *et al.* 2007). This is because GRACE gravity measurements integrate total vertical water content at any point. Changes in water table height need to be quantified in estimating changes in ice sheet thickness.

High-performance computing

Many of the models discussed above are cross-sectional. They are typically constructed parallel to the ice sheet flow lines. Reducing the spatial dimension allowed these authors to represent coupling between heat and flow that could probably not have been done easily in three spatial dimensions or without using high-performance computing. On the other hand, the continental-scale models from the Waterloo group (Lemieux *et al.* 2008a,b,c) provided the first estimate of continental water balances, which could not be done using cross-sectional models. Future heuristic studies of ice-sediment interactions that consider complex feedbacks (e.g., ice sheet dynamics, permafrost formation-brine exclusion, and effective stress–porosity–permeability coupling) should start by developing cross-sectional models. High-performance computing, while having a high overhead in terms of code development costs, represents a promising avenue for three-dimensional ice-sediment hydrologic model development. One potential solution would be to minimize development costs by using freely available open-source codes recently developed in glaciology such as PISM (Bueler and Brown, 2009; Winkelmann *et al.*, 2010) or Elmer (Gagliardini & Zwinger 2008) and coupling them to hydrogeologic models.

The way forward

Hydrologic models developed by glaciologists are groundbreaking in that they consider two-way coupling between ice-sheet dynamics and subsurface flow. However, their models only include a 1- to 2-m-thick till aquifer. What are the consequences of this assumption? Would simulated pore pressures be different if an underlying 1000-m-thick sedimentary or fractured rock layer was also included? Hydrogeologists, on the other hand, have represented ice sheets using idealized polynomial functions (one-way coupling) and neglected ice-sheet dynamics. It is unclear what are the consequences of this assumption. Pressure-dependent basal sliding rates could affect ice-sheet topography and hence subsurface heads. New laboratory experiments in a large ring-shear device where ice is sheared above a soft sediment bed are underway to study the coupling between pore pressure, till deformation, and basal sliding (Iverson & Petersen 2011). By mimicking subglacial conditions in a controlled environment, these experiments should provide a much-needed assessment of the factors that control feedbacks between basal sliding rates and till hydrology.

Virtually all models presented in this review have represented hydromechanical effects using vertical stress formulations (i.e., Terizhaghi's law). The validity of this assumption needs to be tested by comparing computed head patterns to multidimensional hydromechanical models (i.e., Biot theory). The effects of flexure rebound associated with glaciations on the buildup of crustal pore pressure, as proposed by Neuzil (this issue), requires further quantitative analysis. Careful model intercomparison studies between codes that use Biot theory verses vertical loading seem warranted.

Glaciologists have recently incorporated isotope transport processes into their ice sheet models (e.g., Clarke & Marshall, 2002). Linking this to a subsurface isotope transport code may shed light why it is not common to find highly depleted groundwater in sedimentary basin aquifers overrun by ice sheets (e.g., Ma *et al.* 2004; DeFoor *et al.* 2011; Schegel *et al.* 2011). While theory exists for quantifying solute rejection during permafrost formation (Panday & Corapcioglu 1991), this has not been formally considered in Pleistocene paleohydrologic modeling studies.

Studies by nuclear waste management agencies

Our review has mainly focused on the state of knowledge from peer-reviewed publications. There are a large number of excellent quantitative studies undertaken by nuclear waste management agencies in Canada (NWMO; e.g., Hobbs *et al.* 2011; INTERA, 2011; NWMO, 2011; Peltier 2011; Sykes *et al.* 2011; Finland (STUK; Tsang et al. 2005) and Sweden (SKB; e.g., Geier 1996; Jaquet & Siegel 2003) that were not discussed in great detail in this report.

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