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RESEARCH ARTICLE

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Key Points:

- Numerical simulations show that subglacial groundwater flow under a major paleo-ice stream is entirely reorganized compared with nonglacial times
- A mosaic of subglacial groundwater recharge and discharge areas moves time-transgressively with the advancing ice sheet with highest fluxes around the ice margin
- Drumlin formation is likely facilitated by upwelling groundwater weakening subglacial deposits and enhancing erosion

Correspondence to:

P. Hermanowski,
piotr.hermanowski@amu.edu.pl

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Groundwater Flow Under a Paleo-Ice Stream of the Scandinavian Ice Sheet and Its Implications for the Formation of Stargard Drumlin Field, NW Poland

Piotr Hermanowski¹ and Jan A. Piotrowski^{2,3}

¹Institute of Geology, Adam Mickiewicz University, Poznań, Poland, ²Department of Geoscience, Aarhus University, Aarhus C, Denmark, ³Faculty of Earth Sciences, Nicolaus Copernicus University, Toruń, Poland

Abstract Subglacial groundwater flow was an integral part of glaciological systems of past ice sheets, but its impact on the origin of active-ice landforms remains unexplored. Using numerical experiments, we attempt to constrain groundwater flow dynamics under a major paleo-ice stream of the southern Scandinavian Ice Sheet and its impact on the formation of the Stargard drumlin field. Flow models show a total reorganization of groundwater dynamics under the advancing ice stream to a depth of up to ~200 m. A mosaic of intervening groundwater recharge and discharge areas originates, whereby the same areas may experience multiple shifts in flow directions. A prominent time- and space-transgressive pressure pump recharges groundwater in a subglacial zone up to about 20 km within the ice margin and discharges it in front of the ice sheet. The simulated groundwater flow pattern suggests that drumlins occupying a portion of the ice stream area occur preferentially where groundwater upwells and discharges at the ice/bed interface. Consistent with the geological composition, a possible drumlin-forming mechanism involves an excess of pressurized water under the ice reducing the strength of the subglacial deposits and facilitating glacial erosion, erosion by turbulent meltwater flows or both, streamlining antecedent deposits into drumlin shapes. This study emphasizes the importance of causal feedback between subglacial groundwater flow and the ice sheet dynamics and suggests an impact of the former on the formation of streamlined subglacial landforms in general.

1. Introduction

Subglacial processes (Clarke, 2005) control the dynamic behavior of ice streams, arteries of fast-flowing ice which are important components of both contemporary and past glacial systems discharging disproportionately high volumes of ice and contributing to substantial sediment redistribution (Bennett, 2003; Stokes & Clark, 2001). Location and evolution of ice streams have been linked to subglacial parameters such as topography (Fransner et al., 2017; Hermanowski, 2015; Kessler et al., 2008; Margold et al., 2015; Rignot et al., 2011; Ross et al., 2009; Rychel & Morawski, 2017), geology (Anandakrishnan et al., 1998; Clark & Stokes, 2001; Peters et al., 2006; Rattas & Piotrowski, 2003; Studinger et al., 2001), hydrology (Bell et al., 2007; Christoffersen et al., 2014; Greenwood et al., 2016; Lelandais et al., 2018; Vaughan et al., 2008; Winsborrow et al., 2010), and thermal regimes (Bougamont et al., 2003; Christoffersen & Tulaczyk, 2003a, 2003b). The impact of these parameters on ice flow is, through a series of feedbacks, coupled with glacial meltwater (Bougamont et al., 2011; Christoffersen et al., 2018; Clason et al., 2014; Gudlaugsson et al., 2017; Winberry et al., 2009), which feeds the subglacial groundwater system.

Numerous studies have suggested that subglacial groundwater is an important part of the glacier system and should be considered in hydrological studies of both modern (e.g., Christoffersen et al., 2014; Gooch et al., 2016; Mikucki et al., 2015; Siegert et al., 2017) and past (e.g., Boulton et al., 1995; Lemieux et al., 2008a; McIntosh et al., 2012; Person et al., 2012; Piotrowski, 2006) ice sheets. During glaciations, high groundwater heads imposed by the overlying ice coupled with high volumes of meltwater generated at the ice/bed interface or reaching the bed from the ice surface (Adhikari & Tsai, 2015; Christoffersen et al., 2018; Zwally et al., 2002) caused reorganization of subglacial hydrogeology in relation to nonglacial times: groundwater flow is deeper, flow velocities and fluxes higher, and flow direction is generally oriented from the ice sheet interior toward the ice margin. Due to the typically insufficient capacity of the substratum to evacuate meltwater

from the ice sole as groundwater flow (Piotrowski, 2006), basal water pressures under contemporary ice streams are close to the ice floatation level (Engelhardt & Kamb, 1997; Smith et al., 2007) and similar conditions have been suggested also for the southern fringe of the Scandinavian Ice Sheet (Piotrowski et al., 2009; Piotrowski & Kraus, 1997; Piotrowski & Tulaczyk, 1999). This highly pressurized surplus meltwater affects the strength of coupling between the ice and the bed and consequently the ice sheet behavior (Bell, 2008; Breemer et al., 2002; Cutler et al., 2000; Kyrke-Smith et al., 2014; Piotrowski, 1997a) with a possible destabilizing effect.

Interactions between ice, water, and subglacial sediment generate an imprint in the geological and geomorphological record. High water pressure combined with fast ice flow promote subglacial erosion, sediment advection, and deformation (Tulaczyk et al., 2000; Damsgaard et al., 2015, 2016; Evans, 2018), which is reflected in such landforms and their deposits as drumlins (Menzies et al., 2018), mega-scale glacial lineations (Clark, 1993), tunnel valleys (Kehew et al., 2012), and glacial overdeepenings (Cook & Swift, 2012). Consequently, deciphering the geological and geomorphological signatures should enable reconstruction of the interactions between past glaciers and their beds. Supplemented by numerical modeling, such studies have a potential of quantitatively constraining the behavior of ice sheets.

This study focuses on subglacial hydrogeology in the terminal part of a major ice stream (Odra paleo-ice stream; OPIS) at the southern fringe of the Scandinavian Ice Sheet during the last (Weichselian) glaciation where one of the biggest drumlin fields in the central European Lowland (the Stargard drumlin field; SDF) is found. Using steady state and transient groundwater flow simulations, we attempt to inform about the dynamics of meltwater drainage through the glacier bed and evaluate its potential impact on the ice stream behavior. In particular, our aims are to (1) estimate flow patterns and fluxes of the subglacial groundwater and (2) discuss the potential influence of subglacial groundwater dynamics on the formation of SDF. The results may be relevant to deciphering the ice sheet behavior and the origin of drumlins in other areas of ice streams underlain by thick, permeable, and deformable deposits.

2. Study Area and Research Background

The study area is located in the Central European Lowland just south of the Baltic Sea in NW Poland and NE Germany (Figure 1). It comprises a geomorphologically well defined ice lobe originally referred to as “Odra lobe” by Keilhack (1898), Woldstedt (1931), and Galon and Roszkówna (1961) that marks the outermost part of one of the largest land-terminating ice streams of Scandinavian Ice Sheet (ice stream B² of Punkari, 1997, and Stokes & Clark, 2001) operating during the Pomeranian phase of the last (Weichselian) glaciation about 17 ka ago (Hardt & Böse, 2018; Hughes et al., 2016; Rinterknecht et al., 2005, 2012; Stroeven et al., 2016). The ice lobe margin can be traced as a distinct arcuate belt of ice-contact deposits and landforms projecting several tens of kilometres south of the main Pomeranian ice sheet margin starting at Feldberg in Germany and then continuing to the south through Joachimsthal and Hohenwutzen where it crosses the state border to Poland and runs through Moryn to Barlinek where it turns to the north to Insko and finally to the east where the ice lobe merges with the nonlobate main part of the ice sheet (Figure 1).

The Odra lobe area is a classic example of a major glacial lobe landsystem comprising a set of various glacial features with end moraines, outwash plains, kames, tunnel valleys, eskers, drumlins, and till plateaus (Karczewski, 1968; Starkel, 1997). The most common surficial sediment inside the ice lobe is thick glacial till deposited during the maximum Pomeranian advance and several ice-marginal oscillations during the ice retreat. The Odra lobe forefield hosts extensive outwash deposits (e.g., Pisarska-Jamroży, 2008) often arranged in thick, partly overlapping fans typically found at the mouths of tunnel valleys spreading radially from the ice lobe center. These thick outwash deposits together with the tunnel valley systems emphasize the role of subglacial meltwater as land-shaping and sediment-redistributing agent. Widespread glaciofluvial deposits also occur in the central part of the ice lobe area around the Szczecin Lagoon where they have been laid down during the ice retreat toward the Baltic Sea basin.

The eastern part of the ice lobe area hosts the most extensive drumlin field in the Central European Lowland—the Stargard drumlin field (Figure 1; Keilhack, 1898; Karczewski, 1987, 1995; Rachlewicz, 2001). In this area, Hermanowski et al. (2019) documented over 1,300 drumlins and related streamlined subglacial bedforms such as megascale glacial lineations and glacial flutings. These bedforms are arranged in an arcuate pattern broadly indicating ice flow to southwest, south, and southeast toward the ice margin in accord

Figure 1. Study area in northwestern Poland and northeastern Germany. Odra ice lobe marks the terminal part of Odra palaeo-ice stream (OPIS), a major ice stream draining the southern fringe of the Scandinavian Ice Sheet during the last (Weichselian) glaciation. The extent of the transient-state model of the subglacial groundwater flow is the entire area of this figure south of the Baltic Sea shore. Also marked is the extent of steady state model comprising the Stargard drumlin field.

with the general shape of the ice lobe. The streamlined bedforms are typically 600–800 m long, 200–250 m wide, 3–6 m high, and have elongation ratios mostly around 2 but in some cases as much as 15. Based on detailed geomorphological and sedimentological analysis, in particular, the lack of apparent correspondence between the drumlin deposits and drumlin shapes, Hermanowski et al. (2019) suggested that the drumlin field originated as a result of some combination of direct glacial erosion and subglacial meltwater erosion by removing older material from the interdrumlin areas and streamlining the resultant bumps.

Groundwater flow under the Odra ice lobe was subject to a preliminary steady state simulation (Hermanowski & Piotrowski, 2009), and the eastern section of the lobe located in Poland was part of low-resolution, large-scale numerical simulations ($\sim 66,500 \text{ km}^2$) of both steady state and transient groundwater flow using finite difference codes (Piotrowski et al., 2009). These simulations focused on subglacial groundwater drainage patterns and water budgets with possible feedback on ice sheet stability. The results suggested inefficiency of meltwater evacuation from the ice/bed interface as groundwater flow due to thick low-permeability deposits in the substratum (mainly tills of Saalian and Elsterian glaciations). This geological setting, in combination with high meltwater production rates, was believed to have caused high pressure of basal meltwater and decoupling of the glacier from its bed. The modeling by Hermanowski and Piotrowski (2009) indicated groundwater discharge in front of the glacier but also suggested upwelling of groundwater toward the ice sole in the SDF area, which was not elaborated any further.

Drumlins and megascale glacial lineations are considered proxies of fast ice-flow in ice sheets (Benediktsson et al., 2016; Colgan & Mickelson, 1997; Eyles & Putkinen, 2014; Smith et al., 2007; Stokes & Clark, 2002), while their internal composition may (Iverson et al., 2017; Menzies et al., 2016) or may not (Eyles et al.,

2016; Möller & Dowling, 2016) reflect processes operating at the ice/bed interface. Although the latter scenario considers that the deposits in drumlins predate the drumlinizing process, it is still consistent with fast flow of ice. Due to its location within a well-constrained terminal part of OPIS, abundant geological and hydrogeological data, and new sedimentological information documenting the internal composition of the drumlins (Hermanowski et al., 2019), SDF as an example of an area of fast ice flow is well suited for numerical simulation of subglacial groundwater dynamics and its impact on landforming processes.

3. Hydrogeology of the Study Area

The hydrogeological system in the Odra ice lobe area comprises extensive Cretaceous, Paleogene-Neogene, and Quaternary deposits. Deep Cretaceous and Paleogene-Neogene aquifers consist of marls, limestones, and sands. These aquifers discharge to the Odra River through a complex hydraulic system or directly to the Baltic Sea in the north. Along the Odra River valley, an artesian groundwater table of the Paleogene-Neogene aquifer was reported (Paczyński et al., 1972). Aquifers within the Quaternary succession are mainly located in Pleistocene glaciofluvial sands. At the regional scale, four Quaternary aquifers are distinguished: one unconfined aquifer at the top and three confined aquifers separated by tills (Wiśniowski, 2007). The lowermost Quaternary aquifer occupies buried valleys eroded during the Elsterian glaciation. The overlying confined aquifers consist of outwash sand of Saalian and Weichselian glaciations. They reach a total thickness of over 100 m in places and host significant groundwater resources. Recharge of these aquifers is along the Pomeranian phase ice marginal zone directly from precipitation or by infiltration from the upper, unconfined Quaternary aquifer through numerous hydrogeological windows, whereas the discharge is to the Odra River and its tributaries (Wiśniowski, 2007).

In the conceptual model adopted in this study, deep groundwater flows in the Cretaceous and Paleogene-Neogene aquifers are considered minimal and the simulations embrace only the Quaternary deposits, which typically rest on thick low-permeability Neogene silts. This assumption is consistent with the limited depth of any significant groundwater circulation estimated as up to ~200 m (Piotrowski et al., 2009). Thus, special interest is on the shallow aquifers expected to be most affected by the overriding ice sheet (e.g., Bense & Person, 2008; Boulton, 2010; Boulton et al., 2007; Carlson et al., 2007; Grasby et al., 2000; McIntosh & Walter, 2005; Moeller et al., 2007; Person et al., 2007; Piotrowski, 1997a, 1997b, 2006; Piotrowski et al., 2009). Moreover, hydrogeological conditions in shallow aquifers are of primary importance for the processes at the ice/bed interface.

4. Methods

4.1. Structural and Hydrogeological Parameters

The geological complexity of the study area necessitated a detailed reexamination of the available data and subsequently a simplification and generalization of the geometrical and hydrogeological parameters to deliver a manageable data set for the numerical model. The area has been studied using cross sections from 30 German (Lithofazieskarten Quartär) and Polish (Szczegółowa Mapa Geologiczna Polski) geological map sheets at a scale 1:50,000 supplemented by a scrutiny of 2,037 borehole logs on the Polish side of the Odra lobe. This set of borehole logs was selected from an extensive database (BankHydro, provided by the Polish Geological Survey) comprising documentation of in total 5,876 boreholes. The selection was based on the location, depth, and description reliability of the boreholes.

The available data have been generalized yielding four hydrogeological layers with two aquifers (model layers 3 and 5) intervening with two aquitards (model layers 2 and 4; Figure 2). The point data were geostatistically extrapolated onto the entire area using the kriging method preceded by variogram analyses. Kriging is especially useful for irregularly spaced data points (Davis, 1986) and is commonly applied in groundwater studies (de Marsily, 1986). The regionalized depths and thicknesses of the layers together with the relevant hydrogeological parameters (see below) were then taken as input to the numerical models.

Hydraulic conductivity values (K) of aquifers were derived from 14 hydrogeological map sheets at a scale of 1:50,000 (Hydrogeologische Kartierung, HK50) on the German side of the lobe, and from pumping tests on the Polish side. Overall 439 hydraulic conductivity data points were interpolated (165 for layer 3 and 274 for

Figure 2. Setup of the numerical model (transient and steady state) of subglacial groundwater flow with hydrogeological parameters assigned to model layers.

layer 5) and generalized (Table 1). Despite the uneven spatial distribution, the number of hydrogeological data points for aquifers was considered sufficient to estimate hydraulic conductivities in the entire area.

Determining the K values of aquitards was more uncertain. By far most of the aquitards consist of glacial tills known to have highly variable hydraulic conductivities across 10 orders of magnitude between $\sim 1 \times 10^{-12}$ and 1×10^{-2} m/s (Piotrowski, 2006) largely controlled by the content of clay. This range of values, typically derived from field experiments, pumping tests, and laboratory tests (e.g., Allred, 2000; Hubbard & Maltman, 2000; Iken et al., 1996; Jones, 1993; Meiri et al., 1990), reflects the potentially large error embedded in the estimation of the hydraulic conductivity of tills, which is additionally magnified by the structural complexity of tills usually not taken into account. If not directly recorded in the available sources, we estimated the conductivities of tills in the study area from their grain-size descriptions supplemented by laboratory measurements that we conducted on 12 samples from the Weichselian till (yielding K values between 2×10^{-6} and 3×10^{-8} m/s). For all model layers, K_z values (vertical direction) are taken as 1 order of magnitude lower than K_x and K_y values (horizontal directions), which reflects the till anisotropy (e.g., Boldt-Leppin & Hendry, 2003; Hart et al., 2006).

The remaining aquifer parameters required as model input (effective porosity— n_e , and specific storage— S_s) were taken from hydrogeological studies on similar deposits elsewhere (Baker & Pavlik, 1990; Domenico, 1972; Gonen & Gvirtzman, 1997; Johnson, 1967; McWorter & Sunada, 1977; Morris & Johnson, 1967; Urish, 1981) considered representative for the deposits at hand.

Table 1
Statistical Parameters of All Collected Hydraulic Conductivity (K) Values for Aquifers (Layers 3 and 5)

Statistical parameter	Value	
	Model layer 3	Model layer 5
No. of values	165	274
Minimum (m/s)	5×10^{-5}	5×10^{-5}
Maximum (m/s)	1×10^{-3}	1×10^{-3}
Average (m/s)	5×10^{-4}	4.5×10^{-4}
Median (m/s)	5×10^{-4}	5×10^{-4}
Variance	856.6	723.8
Standard deviation	29.3	26.9

Note. Data were subsequently generalized and arranged in six spatial zones.

The lowermost model layer (layer 5; Figure 2) overlying the Neogene silt considered as model base is an aquifer made up of sandy deposits from the Holstenian interglacial and gravelly outwash deposits from the early Saalian glaciation. This layer has a mean thickness of 27.6 m, whereby the greatest thicknesses (up to ~140 m) are found in buried valleys (Kopczyńska-Żandarska, 1970); its top surface is relatively even. Horizontal hydraulic conductivities are for gravel 1×10^{-3} m/s, coarse sand 8×10^{-4} m/s, medium sand 6×10^{-4} m/s, fine sand 4×10^{-4} m/s, and very fine sand 2×10^{-4} m/s.

Model layer 4 (Figure 2) is an aquitard mainly consisting of Saalian till. Its geometry mimics to some extent the present-day land surface, whereby it gradually rises toward the lobe margin. It has a mean thickness of 25.9 m except in northwest where it locally exceeds 70 m and rarely falls below 5 m. This aquitard has been assigned a uniform hydraulic conductivity of 2.5×10^{-7} m/s (horizontally), which is similar to K values of tills used in some other paleo-groundwater simulations (cf. Carlson et al., 2007; Piotrowski et al., 2009). The effective porosity of this till is taken as 0.31, which is similar to the typical range of till porosities between 0.25 and 0.3 (Fountain & Walder, 1998). We note that under specific subglacial conditions related to till deformation and dilation its porosity can be as high as 0.4 (Engelhardt et al., 1990), which significantly increases groundwater flow velocities and fluxes.

Above the Saalian till are highly permeable fluvial and lacustrine sediments of Eemian and Early Weichselian age that were merged into one hydrogeological unit constituting the uppermost aquifer (model layer 3; Figure 2). The bottom surface of this layer is between ~36 m below sea level and ~72 m above sea level, and its maximum thickness is 41 m. In the central, and partly southeastern part of the Odra lobe the thickness of this aquifer falls below 2 m. In about 70% of the study area this aquifer is less than 5 m thick, and the hydraulic conductivities are arranged in six zones.

Model layer 2 (Figure 2) is an aquitard mainly consisting of Weichselian till. It has different geometries in the transient and steady state models. In the transient model a uniform thickness of 2 m of basal till deposited everywhere under the advancing glacier was applied. In the steady state model the ice/bed interface is analogous with the present land surface, and consequently, the mean thickness of this layer is 26.5 m. As in layer 4, we applied here a uniform hydraulic conductivity. Its horizontal value of 6×10^{-7} m/s constrained by laboratory analyses is somewhat higher than in layer 4 reflecting its coarser-grained texture. We assume that the present-day K values represent what K was under the ice sheet because subglacial tills in the southern Peribaltic area typically are not overconsolidated due to high pore-water pressure under the weight of the ice sheet (e.g., Møller Just, 2017; Piotrowski & Kraus, 1997).

4.2. Setup of the Numerical Models

Water flow through the glacial substratum in the study area can be considered as a special case of Darcian flow through anisotropic, heterogeneous porous medium in which the groundwater flux is expressed as

$$q = -\underline{K}\nabla h$$

where $\underline{K}$ is a second order symmetric tensor of hydraulic conductivity, which has dimensions of L/T, and h is the hydraulic head, which is a function of elevation head and pressure head. Groundwater flow is governed by the mass conservation law in a fully saturated porous medium. Mathematically, for anisotropic and heterogeneous porous material under transient flow conditions the mass conservation law states that

$$\frac{\partial}{\partial x} \left(K_x \frac{\partial h}{\partial x} \right) + \frac{\partial}{\partial y} \left(K_y \frac{\partial h}{\partial y} \right) + \frac{\partial}{\partial z} \left(K_z \frac{\partial h}{\partial z} \right) = S_s \frac{\partial h}{\partial t}$$

where S_s is the specific storage, which reflects the water storage release in elastic media and has dimension of L^{-1} , and t is time (e.g., Domenico & Schwartz, 1998; Flowers, 2015; Freeze & Cherry, 1979; Haque, 2015; Hiscock & Bense, 2014). For steady state conditions the right-hand side of this equation is zero.

The response of the subglacial groundwater system to the overriding by the Odra ice lobe was simulated by two numerical models, one for transient conditions during the ice advance covering the entire study area (~18,500 km²) and another one for steady state conditions covering the area of Stargard drumlin field (~6,000 km²; Figure 1). The former was designed to provide generalized time-dependent overview of groundwater response to ice overriding, while the latter focused on the geometry of groundwater flow. Both

Figure 3. Schematic visualization of the numerical model of subglacial groundwater flow showing the boundary conditions and conceptualized groundwater flow pattern.

simulations use the hydrogeological architecture consisting of four layers described above with the base of layer 5 considered impermeable (Figures 2 and 3). No-flow boundary is also applied along the western and eastern edges of the models. This simplification is justified by the fact that these boundaries run approximately parallel to the ice movement direction, that is, along the maximum ice slope gradient that under ideal conditions would also correspond to the hydraulic gradient across which no water flow would occur. Along the northern edge of the area a prescribed-head boundary is applied to simulate groundwater inflow from farther up-ice, whereas the southern boundary is open to enable free exchange of groundwater between the subglacial and proglacial areas.

The uppermost model layer (layer 1; Figures 2 and 3) conceptually simulates the ice sheet and is used to implement a prescribed-head boundary, which is set at 90% of the ice thickness. This position corresponds to the ice floatation pressure, a situation that facilitates fast flow of ice consistent with the location of the model within an ice stream area. The ice thickness (h) was calculated using the empirical formula $h = A L^{0.5}$ where L is a distance from the ice margin and A is a coefficient dependent on the thermal and rheological properties of ice and the bed (Orowan, 1949; Paterson, 1994; Piotrowski & Tulaczyk, 1999). In the axial part of the ice lobe where flow velocities were highest and ice slope lowest, we apply the A coefficient of 1.0, whereas for the marginal parts of the lobe, we take $A = 1.3$. These values are similar to low-profile ice lobe reconstructions elsewhere (e.g., Clark, 1992). A equal to 1.0 corresponds to basal shear stress of ~ 5 kPa, which is somewhat lower than shear stresses under some present (Joughin et al., 2006) and past (Boulton, 2010) ice streams.

In the transient flow model simulating the ice advance the pressure exerted on the groundwater by the overriding glacier is time-dependent. The modeling was performed with finite difference code VisualMODFLOW 2011.1 (McDonald & Harbough, 1988), whereby the area was discretized into a grid of rectangular cells 500×500 m in x and y direction giving a total of 136,675 cells. The ice advance was divided into 35 time steps where each time step lasts 25 years and represents a different position of the ice margin advancing from the Baltic Sea area to the maximum extent of the Pomeranian phase. This gives a total duration of the simulated ice advance of 875 years and a mean ice flow velocity of 137 m/year, which is broadly consistent with earlier estimates (e.g., Stankowski, 1983). The transient simulation allows us to examine the immediate impact of the pressure wave moving time-transgressively with the ice lobe on groundwater flow patterns and fluxes close to the ice sole and in the vicinity of the ice margin.

The steady state model comprises only the eastern part of the Odra ice lobe (Figure 1) hosting the Stargard drumlin field, and it was intended to inform about the glacial meltwater drainage through the bed because the subglacial hydrogeological conditions may influence the drumlin-forming processes (Iverson et al., 2017; Rattas & Piotrowski, 2003). Groundwater flow under the glacier has been simulated using finite element code FEFLOW 6.1 (Diersch, 2014), which facilitates a better resolved representation of the spatial and hydrogeological characteristics by taking advantage of mesh flexibility (e.g., Fetter, 2001; Haque, 2015). The model has an irregular outline grid and a mesh of irregular triangles covering the area of simulation. The

simulations depict a series of 16 steady state situations during the Pomeranian ice advance through the drumlin field area up to the line of maximum ice sheet extent. Boundary conditions as well as the initial conditions for each model layer were imported from the output of the finite difference model.

Our models simulate past conditions that cannot be directly validated using established modeling protocols (e.g., Anderson & Woessner, 1992). Therefore, it is not possible to reduce the uncertainty arising from the generalization and simplification of structural, glaciological, and hydrogeological data used as model input, and to better scrutinize the specified head boundary conditions. However, the robustness of the model was tested through a sensitivity analysis investigating the influence of hydraulic conductivity on the model results (cf. van Griensven et al., 2006; Spear & Hornberger, 1980). This was done by multiple runs with different (yet realistic) hydraulic conductivity values and scrutinizing its impact on the groundwater discharge to the ice forefield.

Due to a limited control, the likely occurrence of permafrost in front of and beneath the ice sheet margin is not taken into account in the numerical model (see Discussion). Permafrost wedge would have migrated with the advancing ice and, due to its extremely low hydraulic conductivity (Williams & Smith, 1991) reduced the groundwater fluxes. Finally, the modeling does not consider the isostatic depression of the terrain (which is justified by the location of the study area at the ice margin) and the subglacial channels known to occur in the field that acted as local discharge conduits for the groundwater.

5. Results

5.1. Subglacial Groundwater Flow Pattern

Both models indicate a strong impact of ice advance on the groundwater system in the Odra lobe area (Figures 4–6). This impact is evident in the entire area, both within each model layer and at the interfaces between them. Caused by the pressure gradient imposed by the potentiometric surface sloping parallel with the ice surface, the general groundwater flow is in the down-ice direction from the ice sheet interior toward the ice margin (Figure 4). In the vicinity of the ice margin the flow lines reflect the shape of the lobe with flow directions spreading radially away from the lobe center. In the cross-sectional view (Figure 5), groundwater flow vectors are also generally oriented toward the ice margin. In front of the glacier, groundwater typically discharges at the ground surface. Below the ice margin and several kilometers up-ice, there is a distinct zone of groundwater recharge into the bed. Farther up-ice, the dominating flow direction is also down into the substratum, but there are local areas of flow oriented upward. The aquitards take up the role of mainly vertical water transfer between the aquifers, whereas the latter mostly transmit the water laterally. The steady state model (Figure 6) reveals both, deep groundwater circulation and shallow flow paths. Water particles started in model layer 3 cross all layers in the down-ice direction and some leave the model in the vicinity of the ice margin. In the shallow flow system the particles pass much shorter distances, typically within one or two model layers and finally flow upward to the ice sole.

Groundwater velocities differ significantly throughout the model domain (Figure 4). Model layers 3 and 5 characterized by relatively high hydraulic conductivities constitute preferential flow paths and are crucial for water transfer to the ice forefield. The highest flow velocities occur in close proximity to the ice margin (high hydraulic gradient) and where these aquifers are thin (to maintain flux continuity). Average linear velocities are up to $\sim 5 \times 10^{-5}$ m/s, that is, significantly higher than the flow velocities in these aquifers under present and likely past interglacial conditions. Values lower by several orders of magnitude are observed in the low-permeability layers 2 and 4.

Controlled by the evolving thickness of the ice sheet and the corresponding distribution of pressure at the ice/bed interface, the groundwater dynamics vary strongly during the ice advance. This is visualized in Figure 7 showing time-dependent changes in discharge/recharge patterns in the layer 3 aquifer. Most parts of this layer have water budgets oscillating around $0 \pm 1 \times 10^{-4}$ m/day but there are spots, typically from several square kilometers to more than 100 km² in size where the fluxes are more heterogeneous. The area of most diversified flows occurs in the vicinity of the advancing ice margin. Strong groundwater discharge (negative fluxes) occurs in front of the ice sheet, while recharge (positive fluxes) dominates on the up-ice side of the margin. These two well-defined zones are usually several kilometers wide and may reach up to ~ 15 km. This pattern emphasizes that the hydrogeologically most dynamic part of the glacial system is

Figure 4. Plan view of the average linear velocity vectors of subglacial groundwater in model layer 3 (upper aquifer) during the ice maximum extent in the transient flow model. Note the overall pattern of down-ice groundwater drainage direction and velocity increase toward the ice margin. Location of cross-section A-B from Figure 5 is shown.

around the ice margin and moves time-transgressively through the area with the advancing ice sheet. It may be envisaged as a pressure pump inserting meltwater to the bed just up-ice from the margin and discharging it in front of the glacier, a process capable of refushing significant volumes of groundwater in the subglacial aquifers in relatively short periods of time due to the high hydraulic gradient. One consequence of the recharge and discharge areas shifting in time and space during the ice advance is that most of the subglacial terrain experiences multiple changes in groundwater flow patterns often involving opposite flow directions, as well as significant changes in flow velocities. The magnitude of these changes induced by the overriding ice is much greater than that of any other natural causes during interglacial times.

Figure 5. Cross section A-B through the groundwater flow field (transient model, maximum ice extent) represented by velocity vectors. Note the strong groundwater recharge short distance up-ice from the glacier margin. Location of cross section is given in Figure 4.

Figure 6. Three-dimensional view of groundwater flowlines (red lines) derived from the steady state model. All flowlines start in model layer 3 (upper aquifer). Note the deep groundwater circulation reaching the model base as well as shallow groundwater flow discharged at the ice/bed interface. The preferential flow of groundwater from layer 3 is toward the glacier base.

In the steady state simulation, cumulative groundwater fluxes in all layers are mainly in the range of 0.025–0.1 m/day (Figure 8). The highest fluxes reaching up to ~ 0.325 m/day (0.175 m/day on average) are observed in a ~ 10 - to 20-km-wide subglacial zone parallel to the ice margin. This is where the lower aquifer (layer 5) is thickest, as also seen in the cross-sectional view of Figure 5.

Special attention is given to the transient evolution of the groundwater flow pattern in the upper aquifer (layer 3) in the area of SDF, depicted in 16 stages of ice advance from the northern border of the area to the maximum ice extent in the south (Figure 9). During the early stages of ice advance, in the northern part of the drumlin field two large areas of strong groundwater recharge (positive fluxes) develop (p_1 and p_2 in Figure 9). These fluxes gradually decline in the subsequent stages but remain positive in the entire simulation time and do not shift laterally a lot. South of these areas, a zone of negative fluxes (n_1 in Figure 9) develops in the early stages and remains as such until the end of simulation. Closer to the ice maximum two distinct zones of positive fluxes originate (p_3 and p_4 in Figure 9), whose areal extent increases as the ice continues to advance. Areas in the drumlin field outside the zones mentioned above experience in general an upward movement of groundwater from layer 3 toward the ice sole (negative fluxes) with flux rates of $\sim 5 \times 10^{-3}$ to 1×10^{-2} m/day. Cumulatively for the whole area of Figure 9 during the maximum ice extent the negative fluxes cover $\sim 1,556$ km² and the positive fluxes $\sim 2,602$ km² (Figure 9d). Approximately 200 km² experiences positive as well as negative fluxes during the simulated 16 stages of the ice advance. Strictly within the drumlin field ($\sim 2,600$ km²), about 41% of the area experiences distinctly positive and about 38% distinctly negative fluxes averaged over time. Most of the positive fluxes occur in marginal zones of the drumlin field, while negative fluxes concentrate in the area of ~ 100 km² in the central part of the field.

5.2. Groundwater Budget in the Drumlin Field Area

Both models inform about water budget and groundwater transfer between adjacent hydrogeological layers. Because of the distinct zone of groundwater discharge occurring in front of the glacier, the advancing ice margin strongly impacts the water budget of the overridden areas. For the last 16 stages of the ice advance across the SDF depicted in Figure 9 we comment below on the water budget and focus on model

Figure 7. Groundwater fluxes for model layer 3 (upper aquifer) from the transient flow model representing stages of ice advance from the northern fringe of the study area (a) to the ice maximum position during the Pomeranian phase (h). Negative values indicate areas with groundwater upwelling toward the ice/bed interface and positive values show areas where groundwater recharge into the bed dominates. Note the zone of high groundwater dynamics moving time-transgressively around the ice margin (white line) with groundwater recharge up-ice and discharge down-ice from the glacier margin.

Figure 8. Cumulative groundwater fluxes for all layers obtained from the steady state model for the maximum ice extent during the Pomeranian phase. The highest fluxes occur in the area of thick lower subglacial aquifer (model layer 5).

layers 2, 3, and 4, which are most relevant for the subglacial hydrology and thus for the ice sheet dynamics.

Water budget of the upper aquifer (model layer 3) shows increasing groundwater fluxes (both recharge and discharge) during the first six stages of ice advance (Figure 10a). This is mainly due to the increasing recharge of groundwater from the overlying layer 2 and, to a lesser extent, discharge into the underlying layer 4. In stage 6, the groundwater fluxes through the upper aquifer are approximately twice as high as in the early stage of ice overriding. Noteworthy is that stage 6 is when most of the drumlin field area became overridden by the ice sheet. From stage 6 there is a continuous decrease in groundwater transfer through the upper aquifer. Groundwater inflow into the upper aquifer from the underlying aquitard (model layer 4) remains almost constant during the first six stages and afterward it steadily decreases. During the maximum ice extent, the upper aquifer is recharged mainly from the overlying aquitard and most of this water is transferred deeper into the bed as clearly seen toward the final stages of simulation. During stages 1-6 the water fluxes between the upper aquifer and the neighboring layers are approximately the same, whereas stages 7-16 witness a more positive balance, meaning that this aquifer acts preferentially as groundwater sink.

Of special importance is groundwater exchange between layers 2 and 3 (Figure 10b) as this bears directly on the hydrology of the contact zone between the ice and its substratum. A prevailing upward oriented flow

Figure 9. Evolution of groundwater fluxes in layer 3 (upper aquifer) during 16 steady state stages of ice sheet advance in the Stargard drumlin field area. (a) Stages 1–6. (b) Stages 6–11. (c) Stages 11–16. (d) Stages 1–16 cumulated. Areas of negative fluxes indicate groundwater flow toward the ice sole and areas of positive fluxes are areas of water evacuation from the ice base into the bed. p_1 , p_2 , p_3 , and p_4 are large areas with positive fluxes and n_1 is a large area with negative fluxes. Note in (d) that almost no drumlins occur in the large area P with cumulative positive fluxes close to the ice margin whereas most well-developed drumlins occur in the large area with cumulative negative fluxes N some distance up-ice from the former.

from layer 3 to layer 2 implies water discharge at the ice/bed interface, whereas the opposite flow direction indicates evacuation of water from the ice sole into the bed. During the early stages of ice advance (1–6; Figure 10b) the ice/bed interface experiences discharge of groundwater from the substratum, whereas in the later stages (7–16) water drains from the ice sole into the bed. The shift around stages 6 and 7 is primarily caused by the formation of the increasingly larger zones (p_3 and p_4 in Figure 9c) under the marginal part of the advancing ice sheet where layer 3 acts as a drain capturing groundwater from the overlying layer 2. We note that the till constituting layer 2, despite its relatively low hydraulic conductivity, likely played an important role in groundwater transfer similar to the modern system below some the Antarctic ice streams (Christoffersen et al., 2014).

Figure 10. Subglacial groundwater fluxes between model layers during the 16 stages of ice advance shown in Figure 9. Note the change in fluxes around model stage 6, which corresponds with the ice sheet margin reaching the southern fringe of Stargard drumlin field. (a) Fluxes between model layer 3 (upper aquifer) and the surrounding aquitards (model layer 2 above and model layer 4 below). (b) Fluxes between model layer 2 (upper aquitard) and model layer 3 (upper aquifer) with the upper part of diagram showing inflow from layer 3 to layer 2 and the lower part inflow from layer 2 to layer 3. In stages 1–6 groundwater flow is preferentially toward the ice/bed interface, whereas in stages 7–16 groundwater flow is mainly oriented down into the bed.

Zones p_3 and n_1 (Figure 9c) of respectively positive and negative fluxes in layer 3 are the two largest areas of consistent flow pattern. This is also reflected in the spatial distribution of cumulative groundwater fluxes generalized over all 16 simulation stages (Figure 9d). Interestingly, the generalized zone P in Figure 9d coincides with the part of the ice lobe area with nearly no drumlins, whereas zone N hosts abundant drumlins including some of the most prominent drumlins in the entire field. This situation may indicate a following causal relationship. Drainage of basal water through layer 2 and into layer 3 in the drumlin-free zone P suggests a relatively strong coupling between the ice and the bed and—if indeed our model is correct—excessive drainage of the bed that would result in bed strengthening. Such conditions may have hindered the drumlin formation in zone P, in contrast with zone N where water surplus at the ice/bed contact due to the upwelling from the substratum could have facilitated drumlin formation by weakening the topmost subglacial deposits leading to their deformation and erosion, and/or by causing basal de-coupling and erosion by pressurized water flows moulding the bed (see below).

6. Discussion

The results of our simulations are consistent with other studies showing a strong impact of ice sheets on groundwater flow pattern, velocities, fluxes, and penetration depths (see reviews in Person et al., 2012; Piotrowski, 2006). It is now well established that during glacial/interglacial cycles, terrains overridden by ice sheets experienced large-scale, repeated reorganization of groundwater flow systems (Boulton &

Caban, 1995; le Brocq et al., 2009; Lemieux et al., 2008a, 2008b; Person et al., 2007; Piotrowski, 1997a; Piotrowski et al., 2009; Provost et al., 2012), whereby high water pressures in deep aquifers inherited from glaciations may persist for thousands of years after the ice sheet retreat (Bense & Person, 2008; Person et al., 2007). In previous studies (Hermanowski & Piotrowski, 2009; Piotrowski et al., 2009) we showed that because of the generally low hydraulic conductivity of the bed only a fraction of basal meltwater could have been evacuated from the Odra lobe as groundwater flow and the rest drained mainly through subglacial channels that are abundant in this area. The present simulations suggest a spatially complicated, time-dependent flow system strongly related to the position of ice margin. In comparison with the interglacial time, we observe a reversal of groundwater flow direction (now generally to the north and under the ice sheet generally to the south), a much deeper circulation forced by the hydraulic heads high up in the glacier, and much faster flow velocities. During the ice advance a rather predictable pattern of groundwater flow is evident around the ice margin, whereby upwelling occurs in front of the glacier and recharge occurs in a narrow zone behind the ice margin. This indicates that most areas overridden by the ice sheet first experienced strong groundwater expulsion and then, once under the ice, groundwater recharge within a zone of up to about 20 km moving time-transgressively with the advancing ice. This system has a high hydraulic gradient given by a steep potentiometric surface close to the ice margin potentially capable of completely refilling old groundwater in shallow, highly permeable aquifers.

The mosaic of upwelling and sinking subglacial groundwater flows during the ice advance reflects changes in basal meltwater pressures and most likely also changes in the amount of water at the ice/bed interface. Given the pivotal role of basal water in the ice movement mechanism (Christoffersen et al., 2014) and formation of specific deposits and landforms, most areas of the Odra ice lobe probably experienced multiple shifts in the mode of subglacial processes, primarily between basal decoupling and recoupling. A similarly complex distribution of processes can be envisaged at any time slice in different places at the ice/bed interface. This diversity of subglacial processes suggested by the pattern of groundwater flow yielded by our simulations is consistent with the time-transgressive and space-transgressive models of basal conditions envisaged by Narloch et al. (2015), Piotrowski et al. (2004), and Tylman et al. (2013) and is supported by the genetic and structural complexity of subglacial deposits such as traction tills (Evans, 2018; Evans et al., 2006).

Odra ice lobe marks the terminal part of a major land-terminating ice stream of the Scandinavian Ice Sheet. Contrary to marine-based ice streams, OPIS has no modern analogues to better constrain its movement mechanisms. However, this study, in accord with Hermanowski and Piotrowski (2009) and Piotrowski et al. (2009), indicates that fast ice flow in the terminal part of this ice stream was facilitated by heavily pressurized subglacial water lubricating the bed and enhancing basal sliding. In particular, areas of groundwater upwelling would have contributed most to reducing basal friction and increasing ice flow velocities; in extreme cases a widespread decoupling of ice from the bed can be considered in areas of thick low-permeability deposits (glaciolacustrine silt and clay) such as in the western part of SDF (Dobracka, 1981). Besides the generally fine-grained nature of the Odra lobe substratum (model layer 2), meltwater storage at the ice/bed interface would have been facilitated by relatively high basal melting rates as the area lies in the zone of positive geothermal heat anomaly (Szuman et al., 2018). Ubiquitous meltwater at the ice/bed interface in the study area would be consistent with the study by Shackleton et al. (2018), suggesting that the OPIS area and its up-ice continuation was one major route of subglacial water drainage during the Last Glacial Maximum.

Despite the vast literature on drumlins and subglacial groundwater, there are to our knowledge nearly no studies attempting to combine in a single generic model the formation of drumlins and quantification of groundwater flow through the bed in a drumlin field (cf. Rattas & Piotrowski, 2003). Our numerical simulation comprising a drumlin field offers a possibility to consider drumlin formation in the context of groundwater flow dynamics. In the parallel study of SDF, Hermanowski et al. (2019) documented the internal composition of several drumlins and analyzed the spatial characteristics of over 1,300 individual drumlins together with the geomorphological pattern of the entire drumlin field. Based mainly on the occurrence of diverse and from place to place different deposits (various till facies and sorted sediments) and their structural arrangements (undeformed to heavily glaciotectionized; geological contacts truncated by landform surfaces), it was concluded that the drumlin deposits mainly predate the drumlinizing event. Consequently, it

was suggested that the drumlins are mainly remnants of an antecedent landscape carved by some combination of direct glacial erosion and erosion by subglacial meltwater flows. The occurrence of distinct meltwater features such as scour marks, tunnel channels, and eskers underlines the significance of water flows in the shaping of this drumlin field.

The results of the present study are consistent with the erosional origin hypothesis of Stargard drumlins by Hermanowski et al. (2019). Although there is no obvious systematic relationship at the scale of the entire study area between the location of drumlins and the groundwater flow patterns and fluxes, such correlation is apparent in the south-eastern part of the study area. Two largest areas of consistent upward (N in Figure 9d) and downward (P in Figure 9d) flow of groundwater host the most prominent drumlins and nearly no drumlins at all, respectively. It is difficult to prove that this pattern reflects a causal relationship, yet it may be suggested that—if the numerical model is correct—groundwater dissipation down into the bed in area P reduced water pressure at the ice/bed interface and increased the strength of the bed deposits. Accordingly, little erosion occurred there which is reflected by the scarcity of drumlins. Conversely, groundwater upwelling and discharging at the ice/bed interface in area N would have reduced the strength of subglacial deposits facilitating glacial erosion and, ultimately, weakened the basal contact and cause decoupling opening room for erosion by turbulent subglacial meltwater drainage, all resulting in carving the bed into drumlin shapes by removing material from the interdrumlin areas (Hermanowski et al., 2019). Because area N started to form as an area of consistent upward flow of groundwater early during ice overriding, it is possible that the drumlin-forming process was also initiated rather early.

Our simulations do not consider the impact of permafrost on groundwater drainage, although frozen ground most probably occurred in front of the advancing Weichselian ice sheet in the central European Lowland (Mojski, 2005). Its spatial extent and thickness are very difficult to estimate, which is reflected by the lack of agreement between reconstructions based on field data (Vandenberghe et al., 2014) and numerical modeling (Kitover et al., 2016). In northern Germany and the Netherlands the LGM permafrost could have been up to ~140 m thick (Delisle, 1998). Recently, relict permafrost was discovered in an area of a negative geothermal anomaly in NE Poland at the depth of 357 m (Szewczyk & Nawrocki, 2011) and the LGM permafrost thickness was estimated at ~600 m (Szewczyk, 2017). However, the present study area lies within a positive geothermal heat anomaly (Szuman et al., 2018) where the permafrost thickness must have been significantly lower. The frozen ground was probably discontinuous due to taliks under perennial water bodies (Delisle, 1998; McEwen & de Marsily, 1991) and maybe confined to low-permeability sediments while aquifers remained unfrozen (cf. Lemieux et al., 2016). In NW Poland, permafrost also occurred under the marginal part of the ice sheet itself, most likely as patches separated by thawed ground (Mojski, 1993; Szuman et al., 2013). Subglacial permafrost exerts a strong impact on ice sheet dynamics and the processes below the ice/bed interface (Waller et al., 2012). In the context of our study, this impact would have been twofold. First, it would have mechanically strengthened the bed reducing erosion and deformation. Second, because of the drop of hydraulic conductivity by several orders of magnitude in relation to unfrozen sediment (Williams & Smith, 1991) the permafrost would have strongly limited shallow groundwater flow (Bense & Person, 2008; Cutler et al., 2000; Piotrowski, 1997b) and force deep circulation beneath the frozen ground (Edmunds, 2001; Mooers, 1990). The elevated pressure of meltwater on the up-ice sides of the permafrost wedge would have increased the chances of basal decoupling and ice flow acceleration there. However, because the spatial and temporal characteristics of the permafrost beneath the ice sheet cannot be rigorously constrained, frozen ground has not been included in the modeling, which must be considered a limitation.

Apart from a limited number of cases where isotopic and geochemical data can be used to constrain numerical simulations of groundwater flow under past ice sheets (e.g., Grasby & Chen, 2005; McIntosh et al., 2012; Remenda et al., 1994; Sterckx et al., 2018; Vaikmäe et al., 2001), such models are inherently difficult to validate and thus the results should be treated with caution (Flowers, 2015). The simplifying assumptions related to the boundary conditions increase the degree of uncertainty. Previous studies typically used a specified head boundary prescribed to the uppermost subglacial sediment layer (e.g., Bense & Person, 2008; Person et al., 2003, 2007; Piotrowski, 1997a, 1997b), a specified flux boundary at the ice/bed interface (e.g., Breemer et al., 2002; Carlson et al., 2007) or a combination of the two (e.g., Lemieux et al., 2008a, 2008b, 2008c; Provost et al., 1998). As shown by Person et al. (2012) all these approaches have advantages

and disadvantages. The specified head boundary applied in this study has the advantage of avoiding using the groundwater recharge/discharge rates at the ice/bed interface, which can strongly vary in time and space depending on the basal melting rates and input of surface meltwater through englacial passages. On the other hand, the weakness of the specified head condition is that it requires knowledge of the position of the potentiometric surface, which may vary across the whole thickness of the glacier (the latter may be estimated from preconsolidation characteristics of the subglacial deposits; Piotrowski & Kraus, 1997). On the whole, however, the uncertainties related to fixed head boundaries are likely less than those embedded in the fixed head boundaries (Iverson & Person, 2012).

In order to test the robustness of our simulations, we performed a sensitivity analysis of the maximum ice extent scenario in the transient flow model, which showed that an increase of all hydraulic conductivity values by a factor of 10 in z direction results in the increase of groundwater discharge at the ice forefield by 21%. Doubling of all conductivities in z direction results in discharge increase by 8.6%, whereas reduction by a half decreases the discharge by 16.9%. Greater changes are observed if all hydraulic conductivities in x and y directions are increased tenfold, in which case groundwater discharge at the ice margin increases by about 60%. The analysis shows that the models are sensitive to changes in the hydraulic conductivity of the bed, but since our estimates of this parameter are based on in-depth analyses of extensive geological and hydrogeological data, we consider uncertainties in our simulations resulting from potentially incorrect conductivity values as relatively small.

7. Conclusions

Numerical experiments presented in this study suggest a profound reorganization of subglacial groundwater flow in the Odra ice lobe area in comparison with the modern situation. In general, a major flow reversal from the south-to-north direction at present to the north-to-south direction under the ice sheet is observed, caused by a high hydraulic gradient imposed by the sloping ice sheet profile. Subglacial groundwater flows much faster (up to $\sim 5 \times 10^{-5}$ m/s) and deeper (up to ~ 200 m) under the ice sheet than in a nonglacial time, whereby the flow dynamics are driven by the hydraulic heads high up in the glacier. A complex, time-dependent groundwater flow field is generated with intervening areas of upward and downward oriented flows, whereby the same areas can experience shifts in the flow directions depending on the position relative to the advancing ice margin. The most dynamic flows occur in the vicinity of the ice margin: behind it basal meltwater is preferentially pressed into the bed, whereas in the ice forefield it upwells and discharges on the ground surface. This creates a hydraulic system likely capable of refilling shallow aquifers as the ice sheet moves across the landscape. Fast ice flow could have been facilitated by water surplus at the ice/bed interface where groundwater upwelling occurs and where a thick aquitard is present immediately beneath the ice sheet.

Our simulations evaluated in the context of subglacial landforming processes suggest a possible influence of groundwater flow on the formation of Stargard drumlins. Although averaged over the whole ice advance groundwater discharge/recharge at the ice/bed interface in the drumlin field area is approximately balanced, the flows changed significantly in time. During the early stages of ice advance approximately up to the southern fringe of the drumlin field, groundwater fluxes are increasing with upwelling dominating. The upwelling continues in the main area of the drumlin field until the end of ice advance. Consistent with the hypothesis that Stargard drumlins are primarily a product of glacial erosion (Hermanowski et al., 2019), this suggests that the drumlin forming mechanism may have been facilitated by highly pressurized groundwater discharging at the ice/bed interface through reducing the strength of basal deposits and ultimately de-coupling the ice from the bed followed by meltwater erosion.

Data Accessibility

Groundwater flow modeling has been performed with commercial software VisualMODFLOW 2011.1 and FEFLOW 6.1, both subject to licensing restriction. Access to all original geological and hydrogeological data used in this study and required to replicate our simulations may be obtained upon request (subject to noncommercial use agreement) from the archives of the Polish Geological Institute (<https://www.pgi.gov.pl/en/>), Poland; Landesamt für Bergbau, Geologie und Rohstoffe (LBGR) Brandenburg, Germany (<https://lbgr.brandenburg.de/cms/detail.php/622564>); and Landesamt für Umwelt, Naturschutz und Geologie

Mecklenburg-Vorpommern, Germany (<https://www.lung.mv-regierung.de/>). Essential data used in this study including input files for numerical models (lithology, hydrogeological parameters, thicknesses of layers and stratigraphy) are available from the Pangaea repository at <https://doi.org/10.1594/PANGAEA.902690>.

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