



Modeling the Subglacial Sediment System of the Finnish Lake District Ice Lobe During Deglaciation

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Abstract. The systematic connection of glacial conditions in ice models with subglacial geomorphological observations has been limited by an inability to model subglacial sediment processes. The Finnish Lake District Ice Lobe (FLDIL) presents an opportunity to apply new model approaches in a situation of relative simplicity, with a well-preserved sedimentary record of its past subglacial hydrology and ice flow. With a recent model of the FLDIL subglacial hydrology as driver, we derive a sediment system model ensemble using the Graphical Subglacial Sediment Transport model (GraphSSeT). Model scenarios analyse the impact of varying sedimentary conditions, and resolve spatial and temporal variations in basal sediment thickness, sediment flux rate, grain size and detrital provenance. Our results show the development of a supply-limited system within 10 years characterised by strong seasonal cycles of winter gains from bed erosion, spring and summer losses from the mobilisation of basal sediment and autumn gains from deposition. Modelled at-outlet grain size also varies seasonally and would yield clastic varves, if deposited in a proglacial lake. The results define a submarginal zone of basal sediment depletion extending 40-60 km back from the terminus, in line with the modern-day sediment thickness. The mobilisation of an extensive blanket of sediment from this submarginal zone is proposed to form the Salpausselkä II ice marginal complex. Our model approach provides a template for the validation of subglacial hydrology models against sedimentary observables, opening a path to employ such constraints to study hard-to-observe modern and past subglacial hydrology, and ice conditions.

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1 Introduction

A well-constrained knowledge of the dynamic evolution of sediments in the subglacial environment is a significant factor to understand the development of the cryosphere (Andresen et al., 2024; Overeem et al., 2017), and the associated impacts on landscapes (Hepburn et al., 2024; Boulton et al., 2007), seascapes (Kirkham et al., 2024) and ecological and ocean systems (Cape et al., 2019; Meire et al., 2017; Chu et al., 2009). A significant challenge to developing this knowledge to a high level

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is the ability to quantitatively model the behavior of sediments under ice with sufficient spatio-temporal resolution to capture dynamic processes. The ability to accurately simulate geological and geomorphological observables is key to unlocking our ability to use sediments to constrain hard-to-observe subglacial hydrology systems under major ice sheets and in the past (Licht and Hemming, 2017; Ojala et al., 2022; Lunkka, 2023).

25 Building on foundational concepts (Shreve, 1972; Röthlisberger, 1972; Walder and Fowler, 1994; Boulton et al., 2007) and earlier modeling approaches (Hewitt, 2011; Flowers, 2015; Werder et al., 2013), recent years have seen the integration of subglacial hydrology with ice sheet model codes, e.g. the Ice-sheet and Sea-level System Model (Larour et al., 2012) and a revolution in the application of subglacial hydrology models at-scale with more realistic forcings and higher spatial and temporal resolution (Ehrenfeucht et al., 2023; Dow et al., 2022; Hepburn et al., 2024). In turn, these developments in subglacial
 30 hydrology models have opened the door to better understand the impact of hydrology on sediments and thus new potential to connect the cryosphere with its sedimentary consequences (Delaney et al., 2019, 2023; Delaney and Adhikari, 2020; Aitken et al., 2024).

A recently developed approach to this problem, the Graphical Subglacial Sediment Transport (GraphSSeT) model represents the subglacial hydrology system as a graph, permitting flexible characterisation of network properties, and the ability to track
 35 sediment mobilisation, transport, and deposition driven by channelised water flow (Aitken et al., 2024). As well as resolving sediment flux, GraphSSeT tracks key sediment characteristics such as grain size and detrital provenance, that are routinely observed in the sedimentary record (Licht and Hemming, 2017; Vorren, 1977; Lunkka et al., 2021; Hovikoski et al., 2023).

GraphSSeT has previously been applied to suites of synthetic models (Aitken et al., 2024) but its capacity to resolve more complex systems has not been comprehensively tested. This work applies this approach to a realistic catchment-scale simulation
 40 of the Finnish Lake District Ice Lobe of the Fennoscandian Ice Sheet (Hepburn et al., 2024). This region is a compelling test case due to its relatively simple ice sheet structure, its dynamic seasonal hydrology, and widespread, well-exposed and well-studied morpho-lithogenetic landforms and sedimentary observables with which to compare the findings (Lunkka et al., 2021; Palmu et al., 2021). A model-ensemble is used to develop a knowledge of the subglacial sedimentary system factoring in variations in sedimentary initial conditions. The results are compared to the subglacial and proglacial landscape and its
 45 sediment characteristics.

2 The Finnish Lake District Ice Lobe

The Finnish Lake District Ice Lobe (FLDIL) is a significant component of the former Fennoscandian Ice Sheet (FIS). The FIS is remarkable for the formation of a series of large ice lobes during the Late Weichselian deglaciation, particularly in association with the Younger Dryas stadial (12.9-11.7 cal. ka), the last cold period marking the transition from the Late Pleistocene to the
 50 Holocene (Palmu et al., 2021; Alley et al., 2003; Naughton et al., 2023). Geomorphological and geochronological evidence indicate that, while the main deglaciation of the FIS was paused during the Younger Dryas, the ice margin exhibited significant and complex behaviour involving rapid oscillations, retreats and re-advances across many areas of Fennoscandia (Lunkka et al., 2021; Hughes et al., 2023). The migration of the ice lobe margins was evidently diachronous, varying in both time and space

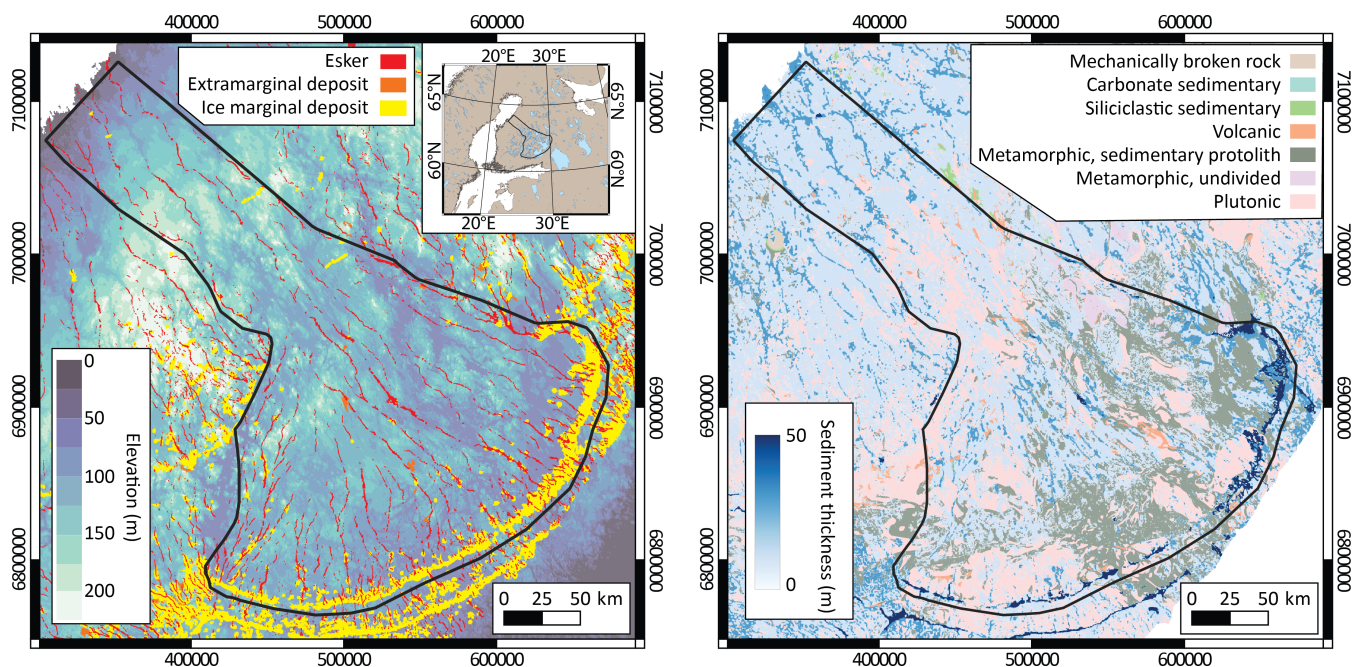


Figure 1. Overview map of the FLDIL region showing a) surface topography (Danielson and Gesch, 2011) and glacial features (GTK Finland, 2025c), b) surficial sediment thickness (GTK Finland, 2025d) and exposed bedrock geology, shown where sediment is absent (GTK Finland, 2025b). Coordinate system is EUREF-FIN EPSG:3067.

(Saarnisto and Saarinen, 2001; Lunkka et al., 2021; Hughes et al., 2023). Regions that experienced substantial ice-margin re-
 55 advances during the Younger Dryas preserve some of the best-developed landforms in the FIS area, such as the Salpausselkäs ice-marginal complexes in southern Finland.

The Salpausselkäs were formed at the ice margin in front of the Baltic Sea Ice Lobe and the FLDIL, which terminated in the Baltic Sea Basin during deglaciation (Boulton et al., 2001; Lunkka et al., 2021). They consist of two sub-parallel main ridges, the Salpausselkä I and the Salpausselkä II, and the more scattered Salpausselkä III, which occurs only in the Baltic Sea Ice
 60 Lobe area. Today, the Salpausselkäs stand out from their surrounding landscape as elevated arcs (100 to 150 m a.s.l.), hundreds of kilometres long, and comprising end moraine ridges, subaquatic fans and extra-marginal glaciofluvial deltas (Palmu et al., 2021; Lunkka, 2023).

Current understanding suggests that, in the FLDIL area, the Salpausselkä I was formed at around 12.5 cal. ka, with a maximum age of 13.3 ± 0.9 cal. ka (Saarnisto and Saarinen, 2001; Svendsen et al., 2004; Lunkka et al., 2021). The age of the
 65 Salpausselkä II is better constrained, as it is associated with the sudden drainage of the Baltic Ice Lake into the North Atlantic via the Kattegat-Skagerrak region at 11.6-11.7 cal. ka (Saarnisto and Saarinen, 2001; Johnson et al., 2022; Regnéll et al., 2025). This drainage occurred soon after the ice margin began to retreat north-northwest from the Salpausselkä II, providing an estimated age of 11.6-12.0 cal. ka for its formation (Saarnisto and Saarinen, 2001; Lunkka et al., 2021). Evidence for the



drainage includes a series of ice-contact delta pairs, located ~5 km up-ice from the Salpausselkä II, with a 26-28 m difference in delta-plain elevation (Lunkka, 2023). This abrupt event closely marks the termination of the Younger Dryas and the onset of the Holocene Epoch, and was followed by a collapse to an ice-free state in Fennoscandia by ~9 cal. ka (Kleman et al., 1997).

The final phase of the FLDIL evolution has been the focus of recent studies aimed at understanding how the self-organisation of the subglacial hydrological network evolved and systematically generated landscape features such as murtoos (Ojala et al., 2019, 2021) and subglacial meltwater routes (Ahokangas et al., 2021; Dewald et al., 2022), which are linked to the transition from distributed to channelised subglacial hydrology (Hepburn et al., 2024).

The FLDIL is a compelling example with which to study both the subglacial hydrological system (Hepburn et al., 2024) and the associated sedimentary evolution, due to several characteristics: First, topographic relief is overall low and flat (Fig. 1a), lacking deeply incised glacial valleys or fjords, and therefore provides little topographic focusing of ice flow; second, the advance and subsequent ice margin retreat from the Salpausselkä II over the model region was relatively short-lived; third, there is an abundance of well-preserved glacial geomorphology including eskers and ice-marginal deposits (Fig. 1a); finally, these sediments and the bedrock geology beneath are well-mapped with high quality and openly available datasets, including sediment thickness maps (GTK Finland, 2025d) and grain-size analyses and geochronology studies (Vorren, 1977; Lunkka et al., 2021).

For this study we frame our work with the following research questions:

1. Can the sub- to pro-glacial sediment dynamics of the FLDIL be well represented in a hydrology-forced sediment system model?
2. Can the sediment system model explain geological observables, including subglacial sediment thickness, and the characteristics and spatial distribution of ice-marginal deposits?
3. Can the comparison of geological observables with the sediment model outcomes help to better constrain ice sheet and subglacial hydrology system evolutions?

3 Methods

3.1 GlaDS and GraphSSeT

Our approach is grounded in the combination of a Glacier Drainage System (GlaDS) subglacial hydrology model (Werder et al., 2013), implemented in the Ice-sheet and Sea-level System Model (ISSM) (Larour et al., 2012) applied as forcing to the GraphSSeT sediment system model.

GlaDS resolves hydrological flow in two coupled components; an inefficient (sheet) flow represented by a linked-cavity network (Walder and Fowler, 1994), and an efficient channelised flow, represented by semi-cylindrical channels, called Röthlisberger-channels, or R-channels (Röthlisberger, 1972). In the finite-element-method (FEM) model the sheet-flow is resolved over the elements while the channelised flow is resolved along element edges. Coupling between these components is achieved by equalising the pressure between the channels and the sheet (Werder et al., 2013).



In GlaDS, the R-channels evolve in time through a thermodynamic balance between expansion through melting versus closure by viscous ice-creep (Werder et al., 2013). The opening rate is driven by heat-transfer from turbulent water flow and changes in the pressure-melting point, and varies with changes in the hydraulic potential gradient, the rate of water flow, and the water pressure (see eqs. 15 and 16 in Werder et al. (2013)). The closing rate is driven by ice deformation proportional to N^n where N is effective pressure, and n is the exponent in Glen's flow law, typically $n = 3$. In this study, the ice geometry is fixed, therefore channel evolution reflects the changing water flow and basal water pressure. With increased meltwater supply, mass conservation demands faster flow, in response to which channels will expand; in contrast decreasing water supply will cause slower flow, and channels will contract. Consequently, the conditions of pressurised high-velocity water flow needed to transport sediment may be quite variable through both space and time.

Taking a GlaDS model result as input, GraphSSeT translates the FEM mesh into a directed acyclic graph incorporating element edges and nodes, through which the channelised hydrology network is represented (Aitken et al., 2024). Edges carry the information for channelised flow including water flow rate, channel cross-sectional area and hydraulic potential gradient. Nodes carry the information of the ice geometry including bed elevation, ice thickness, effective pressure and basal ice velocity. In GraphSSeT elements are not included and the role of hydrologic sheet flow in sediment transport is not considered explicitly.

The GraphSSeT model (Aitken, 2024) involves several processes computed in order. First, the sediment transport capacity is derived based on the basal shear stress τ which is proportional to the square of the mean water velocity u_w in the channel. The transport capacity, in this case using the formulation of Engelund and Hansen (1967), is proportional to $\tau^{5/2}$ and the net effect is that transport capacity varies proportional to u_w^5 . The transport capacity is grain size dependent, and for Engelund and Hansen (1967) is linearly inverse-proportional to d_{50} , the median grain size of sediment at the bed. Other factors are small influences on transport capacity (Aitken et al., 2024). In the second step, sediment mobilisation from and demobilisation to the basal sediment layer is calculated. Here we consider the balance of sediment transport capacity versus sediment supply from upstream and local bed erosion, and the sediment available in the basal sediment layer (Delaney et al., 2019). If sediment supply from upstream is less than the transport capacity, this will lead to mobilisation from the basal sediment layer, so long as sediment is available. On the other hand, sediment supply exceeding transport capacity will lead to demobilisation to the basal sediment layer. The third step considers transport at network-scale, which is managed to maintain the following conditions: sediment volume is always conserved; transport capacity is never exceeded; reasonable sediment velocity limits are never exceeded. In the last stage statistical distributions of grain size and detrital provenance are tracked on the network through volumetric-mixing equations. Detritus tracking is entirely passive, however, the evolving grain size can have significant impacts on transport capacity and so model evolution, where transport-capacity limited (Aitken et al., 2024).

3.2 Model Design and Implementation

For the model forcing we use the subglacial hydrology model of Hepburn et al. (2024, 2023). Hepburn et al. (2024) ran a FLDIL-GlaDS simulation for 10,000 days (27.3 years) to describe subglacial drainage within the FLDIL at the end of the Younger Dryas stadial (~ 12 cal. ka). The model geometry comprised an initially parabolic ice surface that was allowed to relax over 10,000 years, and a representative basal topography created by subtracting sediment thicknesses from a modern

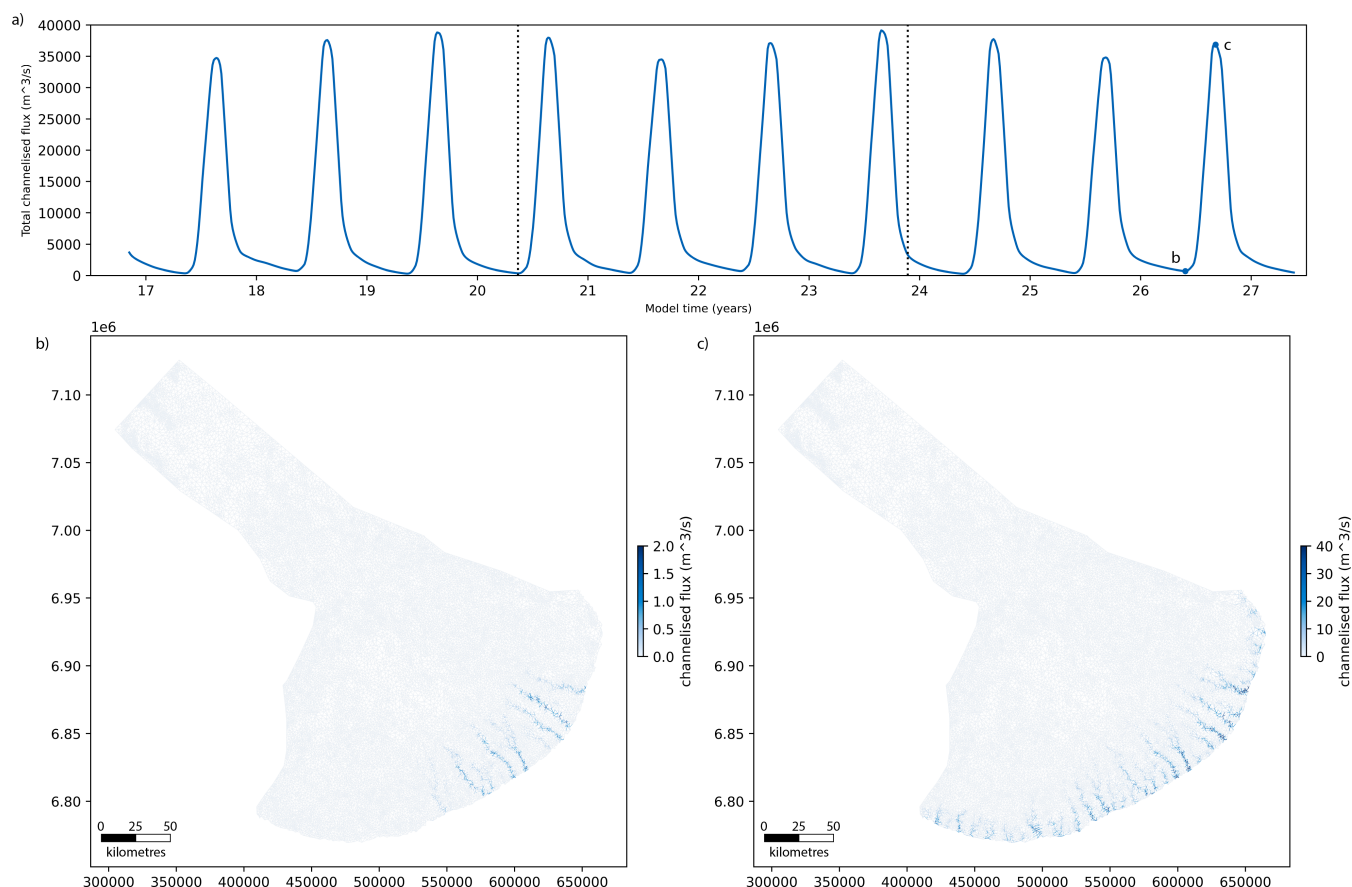


Figure 2. The hydrology model forcing showing a) integrated channelised water flux through the last 10.5 years of the FLDIL-GlaDS model run (Hepburn et al., 2024). Dotted lines indicate beginning, middle and end legs b) channelised water flux at end-winter conditions (26.4 y) c) channelised water flux in peak flow conditions (26.7 y)

135 terrain. The FLDIL-GlaDS model was forced using a modern precipitation record for the region, a modern temperature record
 depressed by 15°C , and an elevation dependant lapse rate of $7.5^{\circ}\text{C km}^{-1}$. A quadratic function was used to estimate monthly
 temperature variability (Wake and Marshall, 2015), and a positive degree day scheme used to calculate total monthly melt rates
 (van den Broeke et al., 2010). Melt was integrated within randomly distributed catchment areas and routed to the bed via 2,500
 moulins. A spatially uniform basal velocity of 150 m a^{-1} was imposed, and the FLDIL-GlaDS model ran with an adaptive
 140 timestep between 90 seconds and one hour. For input to the GraphSSeT model, we use the last 10.5 years of the model run.
 To reduce compute load, the 10.5 year model evolution is divided into beginning, middle, and end segments each of 3.5 years
 duration (Fig. 2). For each, GlaDS model outputs are extracted at ~ 4 day intervals and for each, the channel water flux and
 channel area were spatially filtered to reduce abrupt changes in conditions between edges and ensure network connectivity
 (Fig. 2).



145 The GraphSSeT geology model involves an active basal sediment layer, accessible to hydrology-forced mobilisation, and an underlying bedrock released to the basal sediment layer by glacial erosion. In the standard implementation, the basal sediment layer is variable in thickness between zero and an imposed upper limit (1 m here). In this work, we introduce a mixed-bed mode where thick sediment accumulations are available for mobilisation, but we retain the maximum limit for demobilisation, necessary to avoid potential for runaway sediment accumulations. In the standard implementation, the sediment cover of the area (Fig. 1) is considered as bedrock, requiring glacial erosion, while in the mixed bed model it is considered as basal sediment, able to be mobilised by hydrology. Glacial erosion potential is defined with a quadratic formula (see; Herman et al., 2021) using the basal velocity of 150 m a^{-1} from the GlaDS model, giving erosion potential of $\sim 6 \text{ mm a}^{-1}$.

For each ~ 4 -day GlaDS model output interval, the GraphSSeT implementation runs with constant forcing. GraphSSeT time steps were between one hour and one day duration, adapting to the basal sediment height (H), targeting $\max(dH/dt) = H/2$. At the end of the GlaDS interval, the graph is reformed, including any changes in the directions of edges and/or locations of outlet nodes and updating the forcing parameters. State variables including sediment thickness, active sediment flux density, grain size and detrital properties are transferred to the new graph. Summary model outputs are derived at the end of each GlaDS model output interval, and full model outputs are exported for selected intervals identified at inflections in the total channel flux (Fig. 2).

160 3.2.1 Model Scenarios and Initial Conditions

We tested three different model scenarios with different initial conditions that represent different possible glacial-sedimentological conditions. These scenarios are selected to test key controls on sediment dynamics, and to assess ability to explain observations from the FLDIL region.

1. Low- H : this scenario considers a low initial sediment thickness. This scenario represents a sediment-free former glacial margin, with only a brief period of subglacial sediment accumulation. Sediment thickness is initialised using a 20-year steady state run in winter hydrology conditions, yielding a maximum thickness of $\sim 11 \text{ cm}$.
2. High- H : this scenario considers a higher initial sediment thickness. In comparison to the low- H scenario, this scenario has a more prolonged period of subglacial sediment accumulation. A 150-year steady state run in winter hydrology conditions yields a maximum initial thickness of $\sim 60 \text{ cm}$.
3. Mixed-bed: this scenario considers spatially variable sediment coverage, representing the sedimentary remnants of a previous glaciation with similar geometry. The initial thickness begins with the modern-day sediment thickness and is then initialised with a 20 year steady-state run as before.

3.2.2 Ensemble Trees

Our model design uses ensemble trees (Fig. 3) to develop statistical robustness of the outcome while reducing the compute cost. For each initial condition we have a separate tree, each with three subtrees. The main subtree is the ‘default’ mode, with

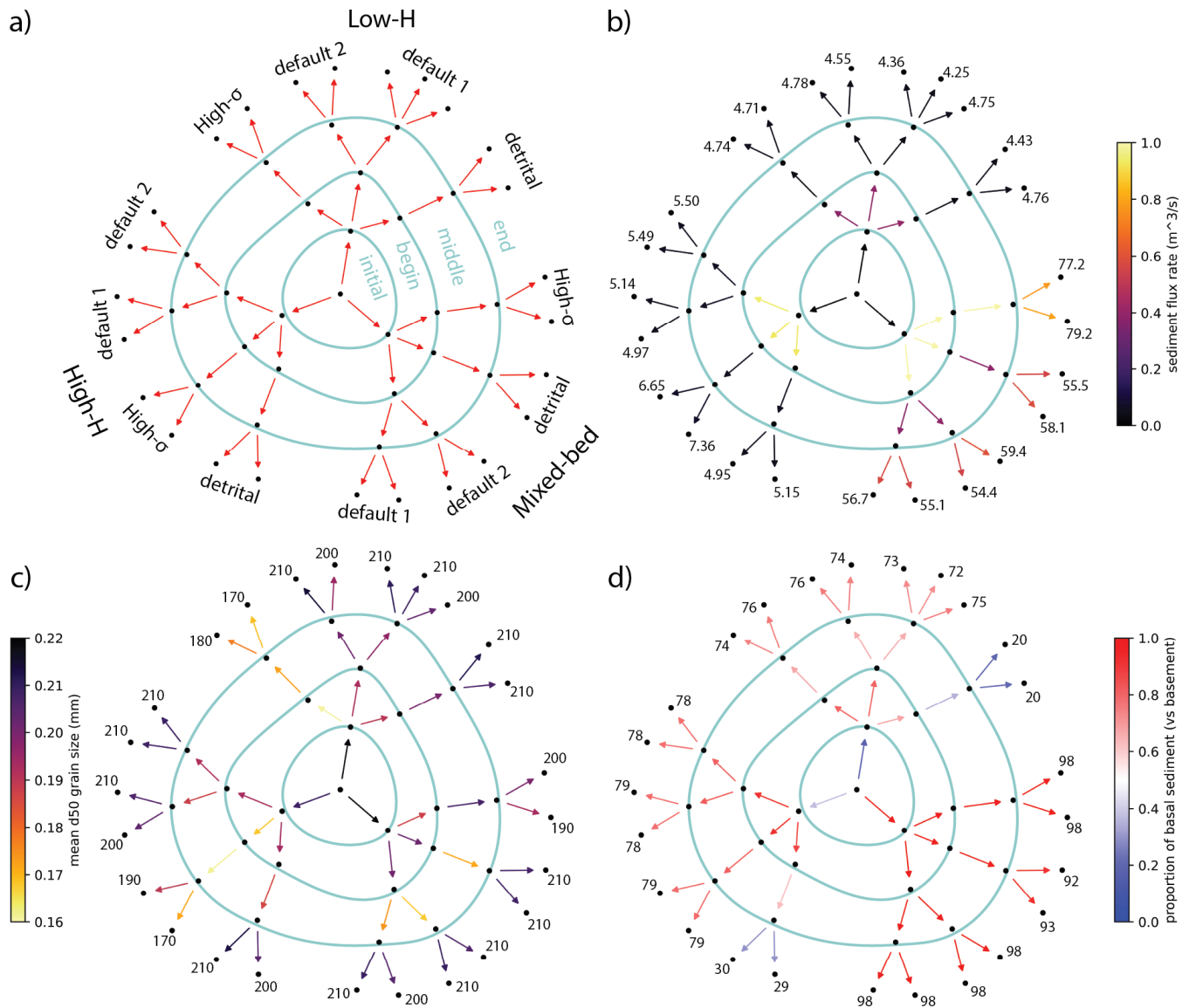


Figure 3. Ensemble structure and summary results including a) structure of the ensemble trees b) time-averaged sediment flux rate for each model leg, labeled values in hundredths of m^3s^{-1} . c) time-averaged and volume-weighted mean d_{50} (median grain size) across all outlets for each model leg, labeled values in μm . d) volumetric proportion of basal sediment for each model leg, labeled values in %. Summary data in Tables A2, A3, and A4.



Table 1. Selected model parameters

Parameter	Default	High- σ	Detritus
Max H (m)	1.0	1.0	1.0
Erode limit (m)	0.75	0.75	0.75
Mean grain size (ϕ)	2.19	2.19	2.19
Grain size (σ)	0.76	1.52	0.76
Detritus Mode	‘normal’	‘normal’	‘bedrock’

two subordinate subtrees using a) the ‘bedrock’ detrital tracking mode and b) increased grain size variability (Table 1). The ‘bedrock’ detritus mode does not affect sediment dynamics - the difference is that the original ‘bedrock’ source characteristic is retained after cycling through the basal sediment layer, whereas in the ‘normal’ detritus mode the provenance is reset to ‘basal’ when deposited. High grain size variability, in principle, will provide a more dynamic model evolution due to its impact on transport capacity. For each subtree a cascade of runs is completed for each leg, yielding at least four results for the ‘default’ subtree, and two for the others (Fig. 3).

4 Results

The results for the beginning and middle legs show high and highly variable volumetric discharge (Fig. 3). These earlier legs represent the period during which the initial condition has not reached a dynamic equilibrium with the seasonal hydrology forcing. For the low- H scenario, the ‘end’ leg model runs show a near-zero net change in sediment thickness and represents a balance between winter gains and summer losses. The high- H runs are similar, but with slightly higher overall discharge and a slight decline in basal sediment thickness. In both these cases, the system is strongly limited by the supply of sediment through erosion. For the mixed bed model, discharge-rates stabilise, but at much higher rates and with a sustained decline in sediment thickness enabled by the much higher availability of sediments. Here we focus on the result of the end leg.

4.1 Low- H scenario

The ensemble result for the low- H scenario (Fig. 4) shows some key features of the sediment system. The sediment discharge rate is low, peaking at $\sim 0.23 \text{ m}^3\text{s}^{-1}$. During the melt season a systematic pattern is seen: The early melt season sees mobilisation of basal sediment and a rapid increase in discharge. This early stage shows, in some runs, a distinct peak in grain size, while in others this peak is absent (Fig. 4a). This peak can be attributed to the high transport capacity of high-pressure flow in developing channels coupled with the availability of previously-deposited sediment.

The melt season continues with, at peak discharge, reduced grain size, indicating selective transport from upstream, and a greater proportion of directly bedrock-derived sediment. Post peak-discharge, the grain size recovers to the mean value, indicating a declining supply from upstream, and by autumn a dominance of bedrock-derived sediment at a low volume. For this scenario the high- σ mode does not have increased discharge rate, although the sediments are markedly finer. This

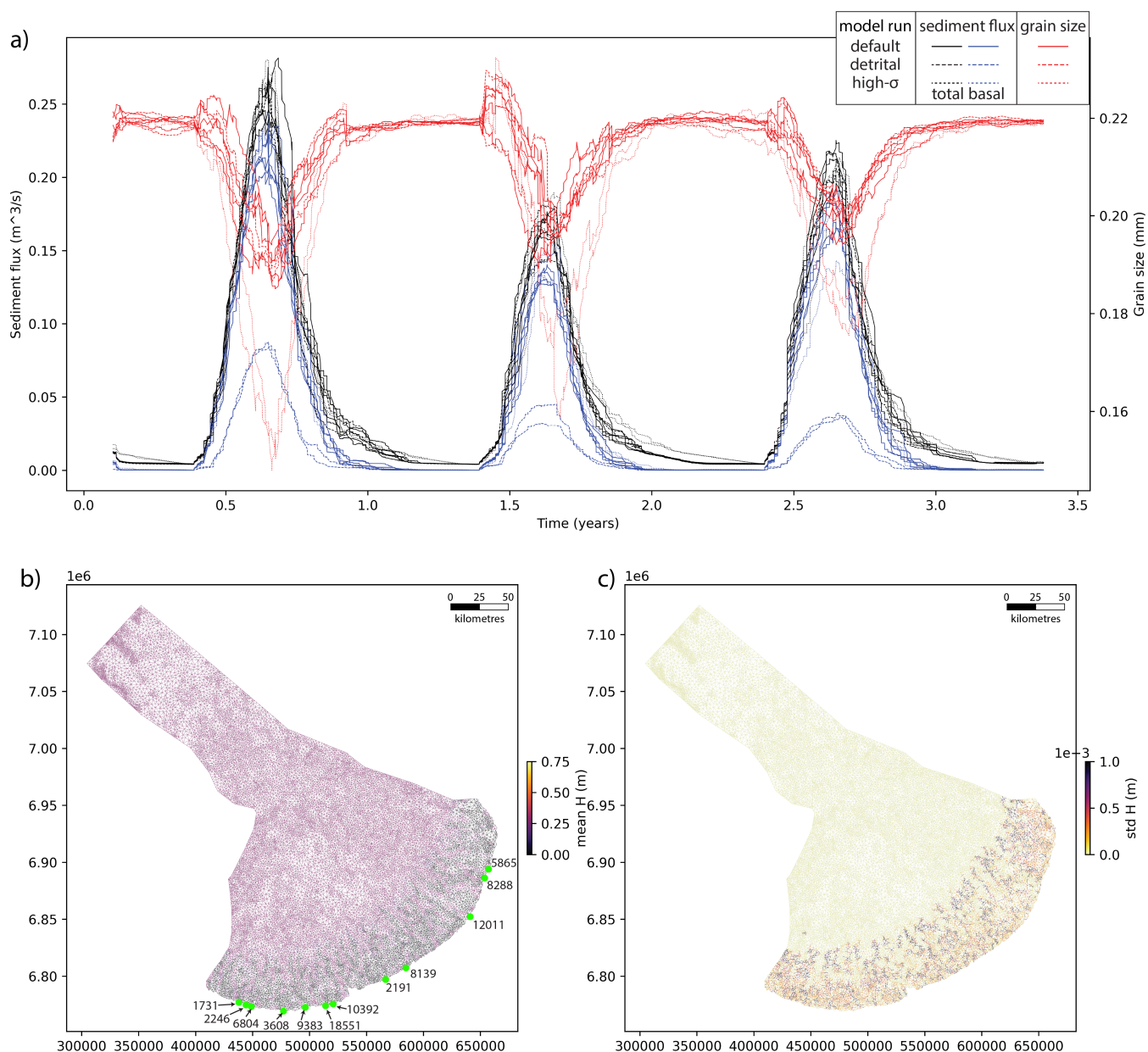


Figure 4. a) The end-leg results for each member of the low- H ensemble tree showing total sediment flux, basal sediment flux and volume-weighted average d_{50} . Dashed lines are detrital models, dotted lines high- σ models. b) ensemble mean and c) standard deviation in sediment thickness at the end of the model run. Labeled green dots indicate the locations of outlet nodes shown in Figs. 8 and 9.



200 indicates that the low- H scenario is strongly supply-limited. The detrital models all have a significantly lower volume derived from the initial basal sediment layer, which indicates $\sim 2/3$ of sediment has experienced at least one cycle through the basal sediment layer.

The remaining basal sediment at the end of the model runs shows a systematic depletion of sediment in channels, extending back ~ 60 km from the ice margin (Fig. 4 b). Variation in the sediment thickness among the ensemble members is low but is
 205 highest in the upper reaches of subglacial channels (Fig. 4 c).

4.2 High- H scenario

Results for the high- H scenario are fairly similar to the low- H scenario, but with a higher discharge rate and a higher proportion of sediment derived from the basal sediment layer (Fig. 5). In this case the high- σ mode does have slightly higher discharge rate at peak flow indicating a greater role for sediment transport capacity in controlling supply from the upper part
 210 of the channelised region. Correspondingly, the detrital models have approximately half the proportion derived from basal sediment, indicating less recycling through the basal sediment layer.

As before, the remaining basal sediment is systematically depleted in channels, extending back ~ 60 km from the ice margin (Fig. 5b). Variation in the sediment thickness among the ensemble members remains low, and is highest in the upper reaches of subglacial channels (Fig. 5c).

215 4.3 Mixed-bed scenario

For the mixed-bed scenario, the results are markedly different (Fig. 6). Most notably, the sediment discharge rate is approximately ten times higher than for the low- H and high- H scenarios, and sediment is derived almost entirely from the basal sediment layer. Median grain size does still reduce during peak flow, but by much less, and has a less distinct relationship with discharge-rate. The high- σ mode has markedly higher discharge rate at peak flow indicating a more significant role for
 220 transport capacity in the upstream region. Sediment discharge remains far below bulk sediment transport capacity. The detrital models indicate that barely any sediment has been cycled through the basal sediment layer.

The remaining basal sediment is systematically depleted in channels, reaching zero in many places, but with regions of higher thickness patchily preserved (Fig. 6b). Variation in the sediment thickness among the ensemble members, while still low, is overall higher and more varied than the other scenarios due to the variable initial sediment thickness (Fig. 6c).

225 5 Discussion

These results highlight some key features of the subglacial sediment system dynamics beneath the FLDIL.

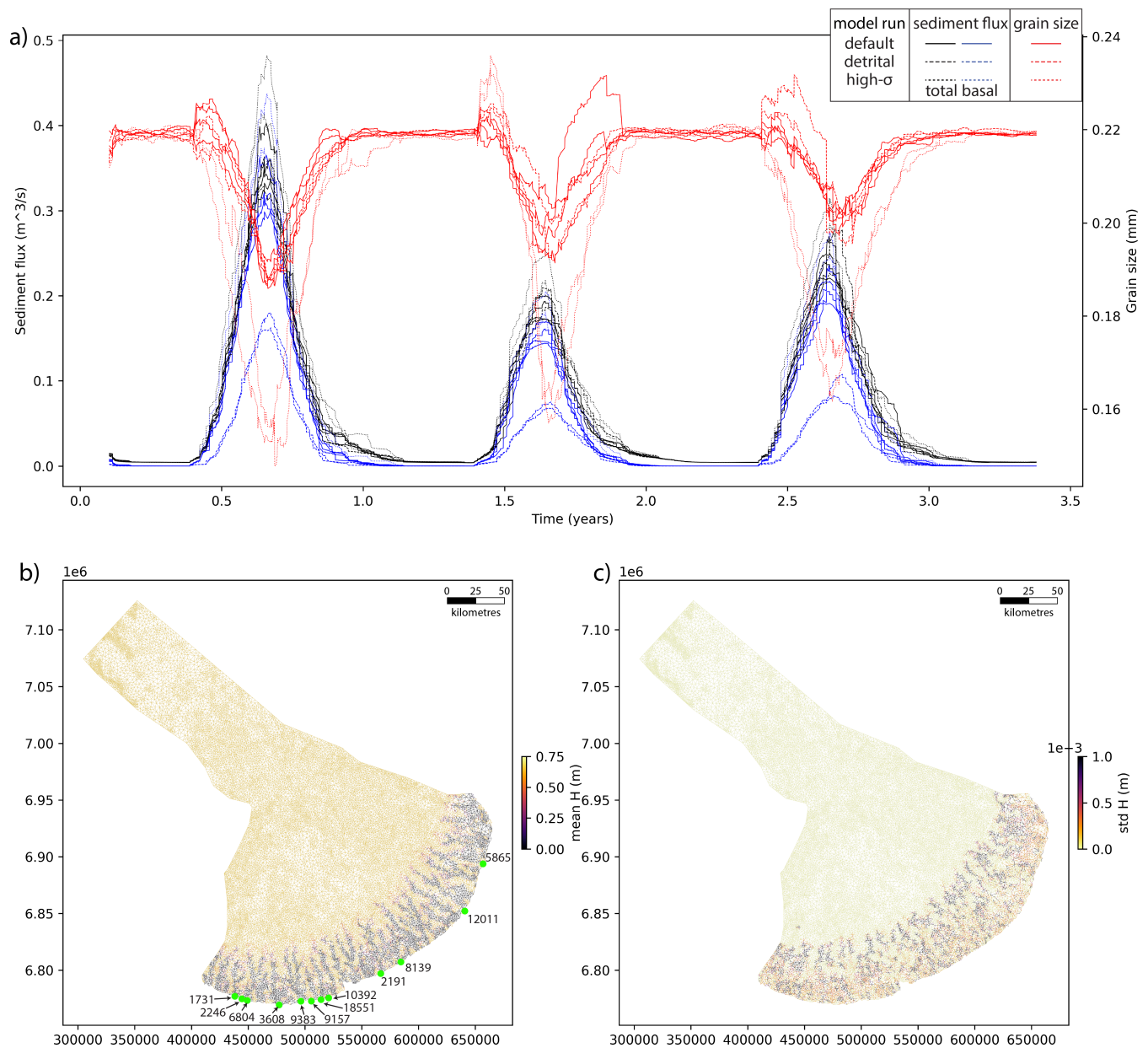


Figure 5. a) The end-leg results for each member of the high- H ensemble tree showing total sediment flux, basal sediment flux and volume-weighted average d_{50} . Dashed lines are detrital models, dotted lines high- σ models. b) and c) ensemble mean and standard deviation in sediment thickness at the end of the model run. Labeled green dots indicate the locations of outlet nodes shown in Figs. 8 and 9.

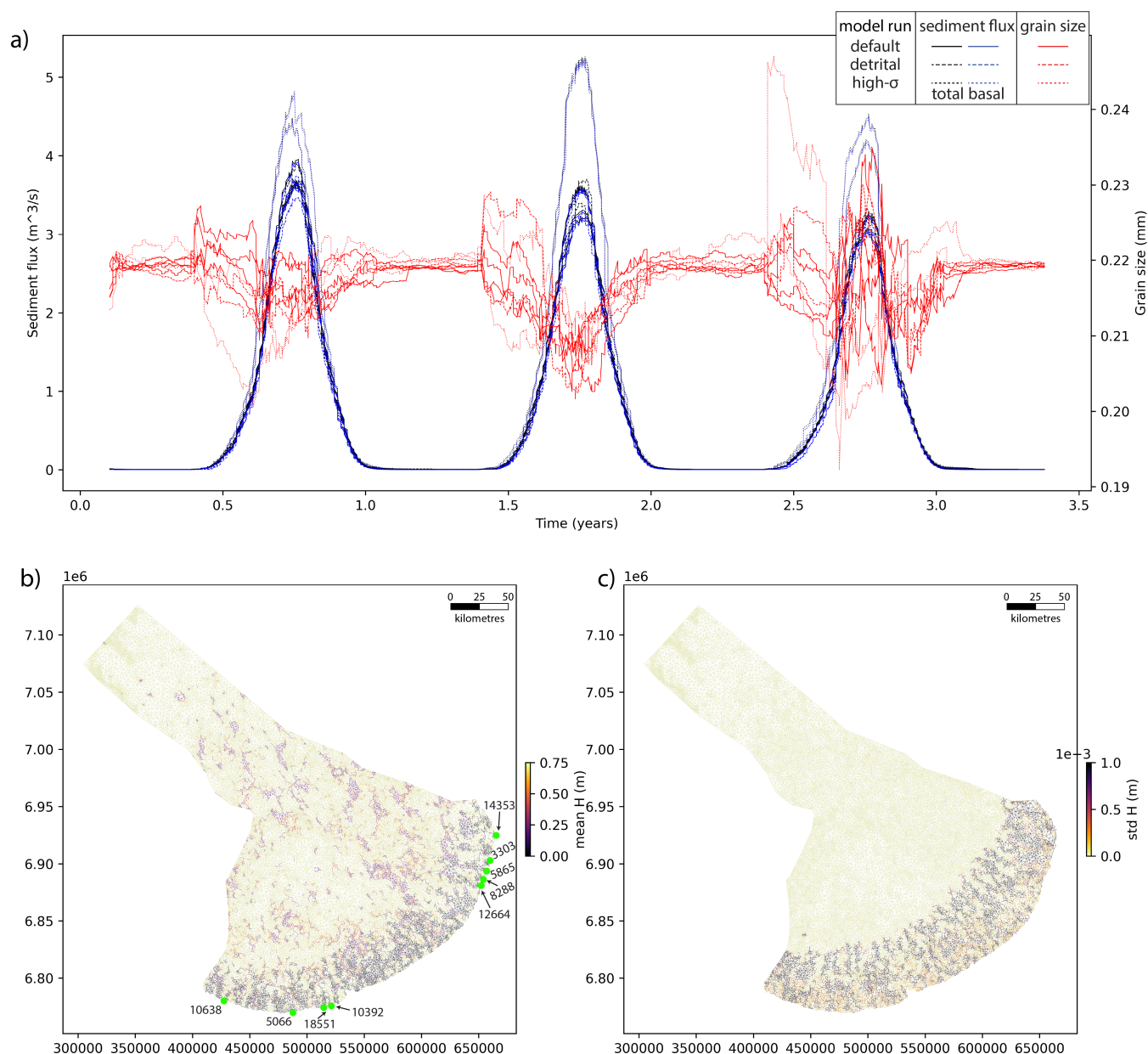


Figure 6. a) The end-leg results for each member of the mixed-bed ensemble tree showing total sediment flux, basal sediment flux and volume-weighted average of d_{50} . Dashed lines are detrital models, dotted lines high- σ models. b) and c) ensemble mean and standard deviation in sediment thickness at the end of the model run. Labeled green dots indicate the locations of outlet nodes shown in Figs. 8 and 9.

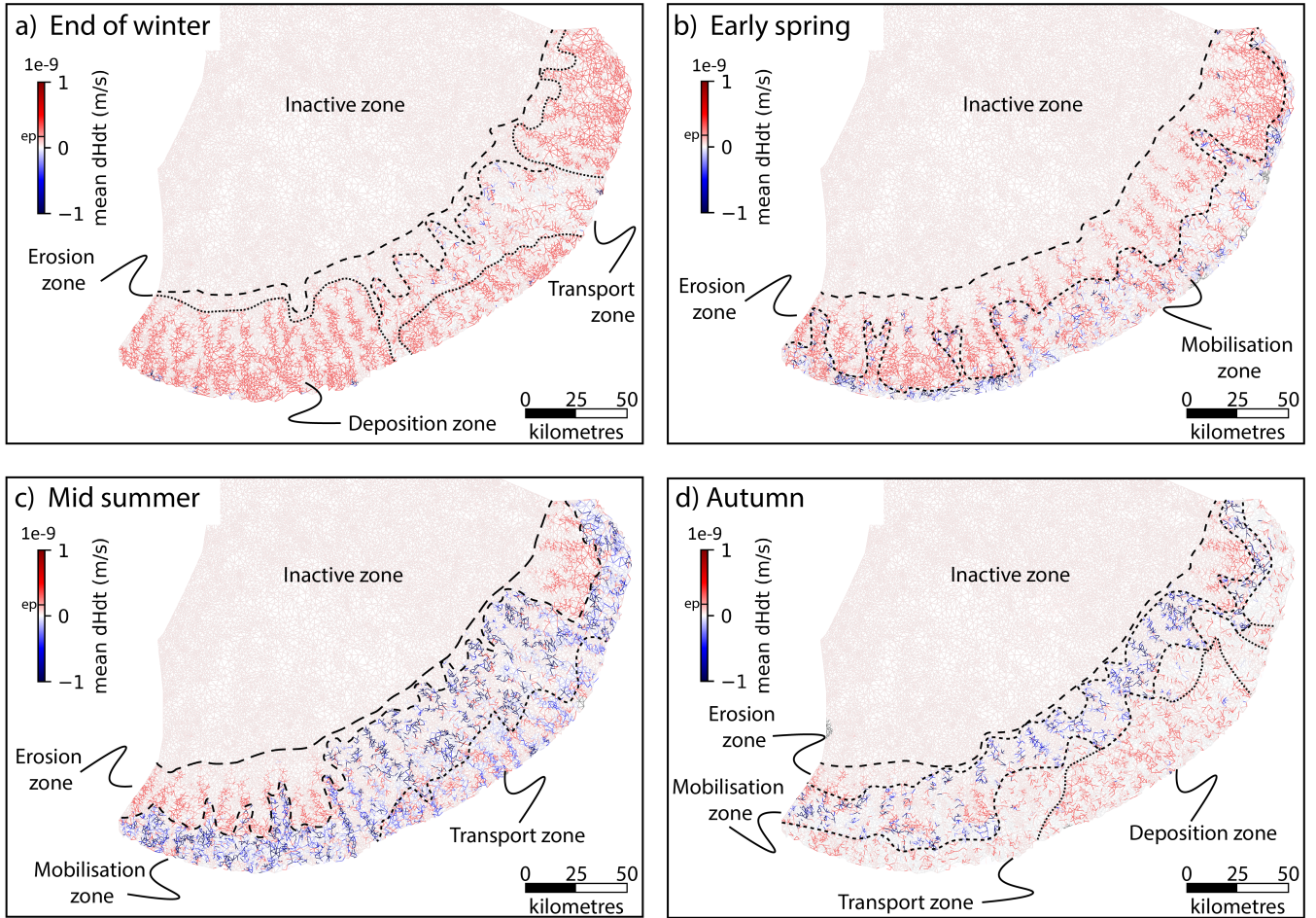


Figure 7. Ensemble mean of dH/dt across a melt season for the high- H scenario showing a) end of winter b) early spring (onset of melt season) c) mid-summer (peak hydrology flow) d) autumn (end of melt season). Interpreted zones of the sedimentary system are shown

5.1 Seasonal to interannual sediment evolution

The results for the low- H scenario, and to a lesser extent the high- H scenario map out a strongly supply-limited regime where discharge during the melt season is limited by the erosion rate (summed over the year). In these conditions, the seasonal cycle shows a clear zonation. Here we describe the seasonal characteristics of the high- H scenario (Fig.7a).

At the end of winter (Fig.7a) the system is dominated by three zones: an inactive zone in the upstream catchment (lacking channels) where $dH/dt \approx 0$; a small erosional zone with $0 < dH/dt \leq \dot{\epsilon}$ ($\dot{\epsilon}$ is erosion potential); a transport zone with variable dH/dt and a frontal depositional zone with $dH/dt > \dot{\epsilon}$. Spring (Fig. 7b) sees a different zonation characterised by the development of a mobilisation front. Upstream of the mobilisation front, dH/dt is positive, while downstream, channels have negative dH/dt as residual sediment is mobilized. Initially, the mobilisation front is close to the ice margin, but then recedes, and by



summer is $\sim 20\text{--}50$ km back from the ice margin (Fig. 7c). This is in line with the processes considered for the deposition of eskers (Núñez Ferreira et al., 2025; Mäkinen, 2003) and murtoos and murtoo-like landforms (Ojala et al., 2019; Mäkinen et al., 2023). A lower transport-zone with both positive and negative dH/dt is emergent in summer downstream from the mobilisation front. The back-propagation of the mobilisation front is consistent with the elevation-dependent positive-degree day forcing applied in FLDIL-GlaDS (Hepburn et al., 2024), which, together with ice-thickness exceeding 1 km, dictates the upstream extent of channelisation. Continuing into autumn, the mobilisation front has reached the upper limit of the channel extents, and has narrowed as mobile sediment is depleted downstream. A deposition zone is reformed near the ice margin with positive dH/dt (Fig. 7d). The low- H scenario model (Fig. A1) has the following differences attributed to lower sediment availability: the inactive zone sees $dH/dT \approx \dot{\epsilon}$, mobilisation in the mobilisation front is more localised, and the deposition in front of the mobilisation front is reduced. The mixed-bed scenario (Fig. A2) demonstrates the same behaviours, albeit, with increased intensity and spatial variability.

Sediment discharge, and sediment properties are tracked at outlet nodes allowing us to analyse a detailed time series of sediment output for specific channels across the model ensemble (Fig. 8). In the low- H and high- H scenarios, we see significant variations in volume output, grain size and detrital character within each melt season. The melt season often begins with a sharp peak in volumetric discharge rate, during which relatively coarse grain sizes (medium-coarse sand) are seen in some cases. Early-season detritus comes wholly from remobilisation of the basal sediment layer. Through the melt season the volumetric sediment discharge rate typically declines, the grain size reduces (to fine or very fine sand) and the proportion derived from the basal sediment reduces. This systematic behaviour has parallels with the diverse sedimentary settings observed in murtoos in the region, previously linked to variability in water velocity and sediment load (Hovikoski et al., 2023; Ojala et al., 2022). This behaviour is fairly consistent along the ice sheet margin, with no obvious spatial trend in the onset timing or duration of the sediment flux event (Fig. 8). Both the low- H and high- H scenarios indicate a situation where sediments deposited during winter are remobilised in spring. Discharge is ultimately limited by supply, first as the local basal sediment is depleted, and then by waning supply from upstream.

The mixed-bed scenario is different, comprising broadly symmetric high-flux events during the melt season, with sediment discharge rate peaking in line with peak hydrology flow. Grain-size in this scenario is more erratic, and the detrital provenance during high flux events is almost entirely derived from the basal sediment layer (Fig. 8). In this scenario the more consistent availability of sediment for some outlets leads to a dichotomy of behaviour in winter: Some outlets show ongoing dominance of basal sediment with fine (but highly variable) grain size (e.g., nodes 9598, 5066) albeit at low discharge rate. In contrast, more sediment-starved outlets are dominated by bedrock detritus with grain size close to the mean (e.g., node 14353) (Fig. 8).

Individual outlets can show significant interannual variability with alternating high- and low-flux seasons, while others show a more consistent series (Fig. 8). High-flux seasons have a greater dominance of basal sediment detritus compared to low-flux seasons (Fig. 8). With an erosion-rate limited system, such variations may occur where there is insufficient basal sediment accumulated over winter to sustain concurrent high-flux seasons. Moreover, the hydrology forcing is not consistent year-to-year and channels have variable extents, and sometimes are not formed for one or several years, and then re-established. This hydrologic variability impacts supply as ‘new’ regions of the bed are exposed to sediment mobilising conditions. The more

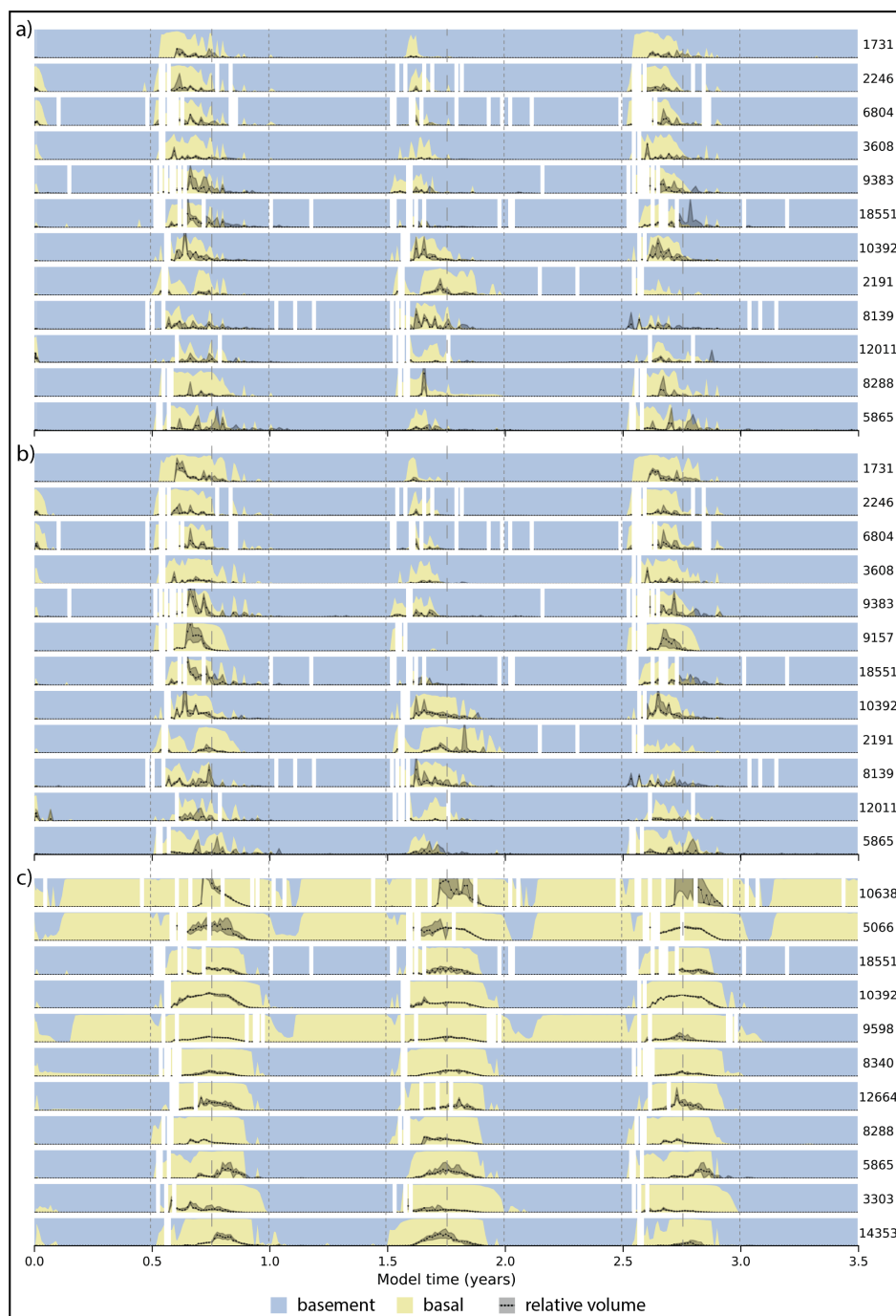


Figure 8. Time-series for major outlet nodes in the ‘end’ period of a) low- H model ensemble, b) high- H model ensemble and c) Mixed-bed model ensemble. For each we exclude the high σ branch. The charts show in colours the proportional detrital output, and the ensemble minimum, median and maximum discharge (relative) volume in black. Nodes are ordered west to east, for node locations see Figures 4, 5 and 6



severe interannual variability seen in the low- H and high- H scenarios relative to the mixed-bed scenario points to variable access to the basal sediment layer as the main factor driving interannual variability.

In detrital provenance, (Fig. 9) the discharge in all cases predominantly reflects the local basement rocks in the vicinity of the outlet, with relatively little variation seen. On the basis of these major units, it would be difficult to distinguish between our scenarios, although major changes in the margin location or hydrological structure (for example, focusing of flow towards the north or south) could be observed. In some cases, minor units are seen only during the melt season, such as the volcanic units seen at nodes 6804, 2246 and 1731 (for locations, see Fig. 4). These units, if observed in detritus, could act as markers of the extent and location of channelized flow.

5.2 Outcomes

This work set out to address three research questions, for which we may now review the progress made:

1. Can the sub- to pro-glacial sediment dynamics of the FLDIL be well represented in a hydrology-forced sediment system model?

The sediment system was adequately represented for timescales of weeks to several years, yielding internally consistent results across ensemble members in each scenario and between the low- H and high- H scenarios. The seasonal behaviours of the sediment system were particularly well resolved and generate variations in sediment yield and grain size characteristics that compare well to sedimentary observations from both murtoos and eskers (Hovikoski et al., 2023; Ojala et al., 2022; Mäkinen, 2003). The interannual record is more limited due to the length of the model run, however, the observed variations in sediment characteristics are consistent with ‘flow switching’ under varying hydrological forcing as a driver of the geomorphological complexity seen in nature (Palmu et al., 2021; GTK Finland, 2025c).

2. Can the sediment system model explain geological observables, including subglacial sediment thickness, and the characteristics and spatial distribution of ice-marginal deposits?

The sediment system model yields geologically realistic outputs that are consistent with the formation of the observed subglacial and proglacial landforms. These include the preservation of the preglacial landscape in the ice sheet interior, with a sediment-depleted submarginal-zone, and extensive ice-marginal deposits. The ice-marginal deposits are dominated by sand, without marked variations in volume along the margin, but with spaced outlets of major channels giving localised depocentres. These depocentres and the gaps between them, bear similarity to the patterns of glaciofluvial landforms (GTK Finland, 2025c) and deposits observed in the Salpausselkä II (Fig. 1).

For grain size, we resolve a systematic annual signal of a coarse spring onset, fining upwards into summer indicating an influx of finer-grained transported material. In a proglacial lake setting such as Salpausselkä II this would likely be represented by an ice-contact delta (Lunkka, 2023) and clastic varves composed of fine-grained silt and clay deposited at the margin and in a proglacial basin in front of the retreating glacier (Zolitschka et al., 2015). This coarser-to-finer grain size variation is accompanied by a change in detritus from reworked basal sediment grading to freshly eroded sediment



Figure 9. Time-series of detrital output for major outlet nodes in the detrital branch models showing a) low- H , b) high- H , and c) mixed-bed. Nodes for each model ordered from west to east (for locations see Figures 4, 5 and 6)



from bedrock. In some cases this change is marked by lithologies that are not seen during low-flux periods, and these could be viable targets for detrital provenance analysis.

With a model run of ten years' duration the pattern of sediment depletion, preservation and deposition is effectively represented with clear similarity to the observed distribution. The model resolves an intensive sediment depletion zone for ~50 km behind the ice margin, and behind this, preservation of pre-existing sediment. These are consistent with the pattern seen in sediment thickness data (Fig. 1) and with the interpreted glacial evolution, which is very dynamic and fast-changing (Lunkka et al., 2021; Ojala et al., 2019, 2022; Mäkinen et al., 2023). Some contrasts exist: for example we do not resolve linear regions of thinner sediment extending for ~150 km. These likely formed as a result of a focused channel network at some stage. FLDIL-GlaDS model testing shows such long channels are difficult to generate with reasonable forcing, and are not consistent with the margin location advanced to the Salpausselkä II (Hepburn et al., 2024). A model incorporating time-transgressive ice-margin evolution would potentially be able to resolve some of these features. Although in-channel fluvial deposition is resolved, the model does not yield eskers because virtually all the sediment deposited during autumn and winter is remobilised in the next melt season (Fig.7). Coarse-sand and gravel grain size is a key feature of esker formation during channel expansion in the early melt season (Mäkinen, 2003) and these coarse-grained deposits are much less susceptible to remobilisation. GraphSSeT does not consider englacial processes and subglacial fluvial transport capacity for gravel grain sizes and larger is limited in the modeled scenarios, so this coarse-grained sediment is not represented.

The Salpausselkä II comprises the major sedimentary deposit for the FLDIL configuration of the Younger Dryas ice sheet. For Salpausselkä II we estimate the volume of sand at ~30 km³ (see Appendix A). Geological data suggest the Salpausselkä II formed within an estimated timeframe of 500 to 1000 years (Lunkka et al., 2021; Joakim Donner, 2010). This implies an average sediment discharge rate of ~0.95 to 1.90 m³s⁻¹, although formation was likely in discrete episodes of more rapid deposition, and with imperfect preservation (Lunkka, 2023; Palmu et al., 2021). To build up the Salpausselkä II volume at the rates of discharge in our models for the low-*H* and high-*H* scenarios, would take ~20 cal. ka and ~16 cal. ka respectively - such rates of discharge are clearly insufficient. A model run at transport capacity (i.e. without any limits imposed on sediment supply) has discharge rates of ~60 m³s⁻¹, requiring just 15 to 20 years to supply the volume for the Salpausselkä II. The mixed-bed scenario needs ~1500 years to supply the required volume for Salpausselkä II and is the most plausible model of our series of models. The implication is that formation of the Salpausselkä II was critically controlled by the limited availability of sediment and involved the re-mobilisation of a sediment cover not dissimilar to today's, but likely more extensive in the now depleted sub-marginal zone.

3. Can the comparison of geological observables with the sediment model outcomes help to better constrain ice sheet and subglacial hydrology system evolutions?

The geomorphology, stratigraphy, grain size and detrital signatures of glacial sediments are commonly used as diagnostic tools for reconstructing glacial systems (Licht and Hemming, 2017; Ojala et al., 2019; Lunkka, 2023; Lunkka et al., 2021). As well as tracking volumes and thicknesses, the model results resolve two significant features that can be used to



constrain interpretations. First is the systemic co-variation of fining-up grain size and decreasing maturity of the detrital provenance during high-flux events, giving criteria for identification of these events in sediment cores. The second is the identification of ‘indicator’ detrital units (see; Aitken and Urosevic, 2021) found only in the fresh detritus during high flux events. These may prove good targets for detrital fingerprinting as they are spatially discrete, geologically distinct and clearly tied to the subglacial hydrology conditions.

5.3 Future directions

Future work may look to refine the approach and resolve some of the deficits identified in this work.

The choice of the initial conditions was very significant for model evolution, in particular, the thickness and distribution of pre-existing sediment cover was in this case the critical factor controlling sediment supply. For accurate volume calculations, this initial condition must be known well, which is challenging for both modern subglacial systems and past examples. A strong control on initial conditions requires robust knowledge of the ice sheet and its subglacial hydrology, constrained by the subglacial environment and its past history and the characteristics of glacial deposits. We show for the FLDIL, where these factors are reasonably well known, the forward-model approach allows us to reject unsuitable initial conditions. For less well constrained examples, data-constrained model ensembles may be sought using inverse-methods that couple geological and geophysical data with the model outputs.

The realistic, but relatively narrow grain size distribution used here (Vorren, 1977) focuses this study on sands, consistent with the majority ($\sim 80\%$) of sediment observed in the region, but does not allow to analyse the origins of the extensive gravel deposits. In particular, we do not test if they are subject to significant fluvial transport, and over what timescales this might occur.

Interannual variations were detected for many outlets, however, the 10-year model run, after the ‘burn in’ period did not allow to resolve this over multiple cycles. A longer model run could allow this to be more fully tested, and to investigate more fully the interaction between variable hydrology forcing and sediment availability.

Eskers and sediment-free zones extending hundreds of kilometers beneath the ice are not a feature of our model. Such features are inconsistent with a model domain fixed at the Salpausselkä II, and would instead require a time-transgressive margin evolution (see; Mäkinen, 2003) over longer time periods than considered here. Additionally, longer-term transport of coarse-grained sediment by englacial processes are likely important in order to provide a consistent source of coarse-grained sediment to the ice margin (Mäkinen, 2003).

6 Conclusions

This work applied the GraphSSeT sediment system model with forcing from a recent GlaDS model of the FLDIL for three different scenarios. Our modelling resolves, for these scenarios, a realistic pattern of subglacial sediment depletion, preservation, and deposition that matches the sediment thickness patterns observed in the subglacial and proglacial landscape of the FLDIL. The model robustly predicts seasonal cyclicity in the sediment supply, transport and deposition system with charac-



teristics that are clearly linked to the hydrological forcing. The sediment characteristics predicted by the model, while not
 370 designed to explain specific features, have volumes, grain size distributions, and seasonal evolutions that match observations
 from glaciofluvial landforms such as murtoos and eskers (Hovikoski et al., 2023; Mäkinen, 2003; Ojala et al., 2021). The
 model is able to explain the supply of sediment for Salpausselkä II with a supply-limited system involving a distributed sedi-
 ment cover similar to but more extensive than today's, at the same time excluding erosion-rate limited and transport-capacity
 limited models. The model has allowed a stronger comparison between numerical models of the past ice-sheet and hydrology
 375 system, and key observations in the sedimentary record, strengthening the links already made (Hepburn et al., 2024; Ojala
 et al., 2022), and lending increased confidence to interpretations of the FLDIL in the Younger Dryas stadial. Looking to the
 modern Earth, models such as this can be used to constrain sediment-based interpretations of glacial conditions in Antarctica
 and Greenland and the associated impacts on landscapes and ecological and ocean systems now and into the future.

Code and data availability. GraphSSeT code available at <https://github.com/al8ken/GraphSSeT> and (Aitken, 2024). GlaDS model data from
 380 Hepburn et al. (2024) available at <https://zenodo.org/records/8344208>. GraphSSeT model outputs, figures and videos for this study available
 at <https://doi.org/10.5281/zenodo.17204682>. Geological and Geomorphological data from GTK (<https://hakku.gtk.fi/en>)

Appendix A

Calculation of the volume of the Salpausselkä II, and timeframes for its deposition, used data sets freely available from the
 GTK Finland web-server (GTK Finland, 2025c, d, a) and was generated with the following process.

385 Ice marginal deposits (from GTK Finland (2025c)) were clipped within the model domain. For each feature, the mean
 sediment thickness (from GTK Finland (2025d)) was calculated using zonal statistics. Multiplying this by feature area yields
 the estimated volume of the Salpausselkä II (within the model domain) of 38.5 km^3 .

For the ice-marginal deposit features a spatial join was performed with the sand and gravel resources data (from GTK Finland
 (2025a)). Across the joined dataset the relative proportions of hiekka (sand), sora (gravel), and murskavatta (larger rocks) were
 390 calculated, yielding 78.0 % sand, 20.2% gravel and 1.8% larger rocks. Our model does not capture gravel transport so we
 assess the total sand volume in Salpausselkä II as 78% of 38.5 km^3 , i.e. 30 km^3 .

From the total sediment yield of our model runs we can calculate the time required to supply this sand volume (Table A1).
 We sum total sediment yield from 'beginning', 'middle' and 'end' legs to yield the residual volume needed to reach the total
 sand volume in the Salpausselkä II. The time required to reach the target volume is estimated by extrapolating the sediment
 395 discharge rate from the 'end' model leg until the required volume is reached.



	Infinite-till	low-H	high-H	Mixed-bed
‘Begin’ volume (m ³)	6.83×10^9	4.29×10^7	1.03×10^8	9.80×10^8
‘Middle’ volume (m ³)	6.94×10^9	3.60×10^6	6.63×10^6	6.42×10^7
‘End’ volume (m ³)	6.50×10^9	5.03×10^6	6.24×10^6	6.84×10^7
Residual volume (m ³)	9.73×10^9	2.99×10^{10}	2.99×10^{10}	2.89×10^{10}
dV/dt (m ³ a ⁻¹)	1.86×10^9	1.44×10^6	1.78×10^6	1.95×10^7
Time required (a)	5.24	2.08×10^4	1.68×10^4	1.48×10^3

Table A1. Total volume discharge and estimate of time required to reach the total sand volume in Salpausselkä II. The infinite-till model assumes that basal sediment is always available and represents a purely transport-capacity limited scenario.

Name	Volume (km³)	Mean flux (m³s⁻¹)	Mean grainsize (μm)	Proportion basal (%)
‘begin’ period				
Baseline_LH_b1	0.0433	0.393	193	80.2
Baseline_LH_hs_b1	0.0415	0.376	161	80.1
Baseline_LH_det_b1	0.0440	0.399	191	67.4
‘mid’ period				
Baseline_LH_b1m1	0.00359	0.0324	196	65.8
Baseline_LH_b1m2	0.00356	0.0322	201	65.9
Baseline_LH_hs_b1m1	0.00381	0.0344	172	67.2
Baseline_LH_det_b1m1	0.00342	0.0309	200	34.8
‘end’ period				
Baseline_LH_b1m1e1	0.00524	0.0475	203	75.2
Baseline_LH_b1m1e2	0.00469	0.0425	209	71.9
Baseline_LH_b1m1e3	0.00480	0.0436	211	72.8
Baseline_LH_b1m2e1	0.00502	0.0455	195	74.4
Baseline_LH_b1m2e2	0.00527	0.0478	214	75.6
Baseline_LH_hs_b1m1e1	0.00520	0.0471	169	75.6
Baseline_LH_hs_b1m1e2	0.00523	0.0474	178	74.3
Baseline_LH_det_b1m1e1	0.00525	0.0476	207	20.5
Baseline_LH_det_b1m1e2	0.00458	0.0443	212	19.7

Table A2. List of model runs and key output parameters for the low- H scenario model ensemble.



Name	Volume (km ³)	Mean flux (m ³ s ⁻¹)	Mean grainsize (μm)	Proportion basal (%)
‘begin’ period				
Baseline_HH_b1	0.104	0.944	193	92.1
Baseline_HH_hs_b1	0.102	0.927	168	92.2
Baseline_HH_det_b1	0.103	0.931	194	87.1
‘mid’ period				
Baseline_HH_b1m1	0.00659	0.0595	192	81.0
Baseline_HH_b1m2	0.00704	0.0636	188	82.5
Baseline_HH_hs_b1m1	0.00644	0.0582	161	81.2
Baseline_HH_det_b1m1	0.00645	0.0583	186	62.9
‘end’ period				
Baseline_HH_b1m1e1	0.00607	0.0550	209	78.2
Baseline_HH_b1m1e2	0.00606	0.0549	208	77.8
Baseline_HH_b1m2e1	0.00549	0.0497	203	77.8
Baseline_HH_b1m2e2	0.00567	0.0514	205	78.5
Baseline_HH_hs_b1m1e1	0.00733	0.0665	190	78.6
Baseline_HH_hs_b1m1e2	0.00812	0.0736	172	78.9
Baseline_HH_det_b1m1e1	0.00546	0.0495	213	29.5
Baseline_HH_det_b1m1e2	0.00568	0.0515	204	29.5

Table A3. List of model runs and key output parameters for the high-*H* scenario model ensemble.



Name	Volume (km ³)	Mean flux (m ³ s ⁻¹)	Mean grainsize (μm)	Proportion basal (%)
‘begin’ period				
Baseline_MB_b1	0.985	8.92	201	99.6
Baseline_MB_hs_b1	0.976	8.85	188	99.6
Baseline_MB_det_b1	0.980	8.88	201	98.1
‘mid’ period				
Baseline_MB_b1m1	0.0417	0.376	174	99.3
Baseline_MB_b1m2	0.0432	0.389	168	99.3
Baseline_MB_hs_b1m1	0.129	1.17	207	99.2
Baseline_MB_det_b1m1	0.0425	0.383	172	97.0
‘end’ period				
Baseline_MB_b1m1e1	0.0626	0.567	209	98.0
Baseline_MB_b1m1e2	0.0608	0.551	205	98.0
Baseline_MB_b1m2e1	0.0655	0.594	206	98.0
Baseline_MB_b1m2e2	0.0601	0.544	209	98.0
Baseline_MB_hs_b1m1e1	0.0874	0.792	193	97.9
Baseline_MB_hs_b1m1e2	0.0852	0.772	200	97.8
Baseline_MB_det_b1m1e1	0.0642	0.581	208	92.7
Baseline_MB_det_b1m1e2	0.0612	0.555	208	91.6

Table A4. List of model runs and key output parameters for the Mixed-Bed scenario model ensemble.

Name	Volume (km ³)	Mean flux (m ³ s ⁻¹)	Mean grainsize (μm)	Proportion basal (%)
Baseline_IT_b1	6.83	61.9	219	100
Baseline_IT_b1m1	6.94	62.8	219	100
Baseline_IT_b1m1e1	6.50	58.9	219	100

Table A5. List of model runs and key output parameters for the infinite till model run. In infinite till mode GraphSSeT assumes basal sediment is always available and therefore represents a purely transport limited case. Grain size is fixed at the mean of the population distribution

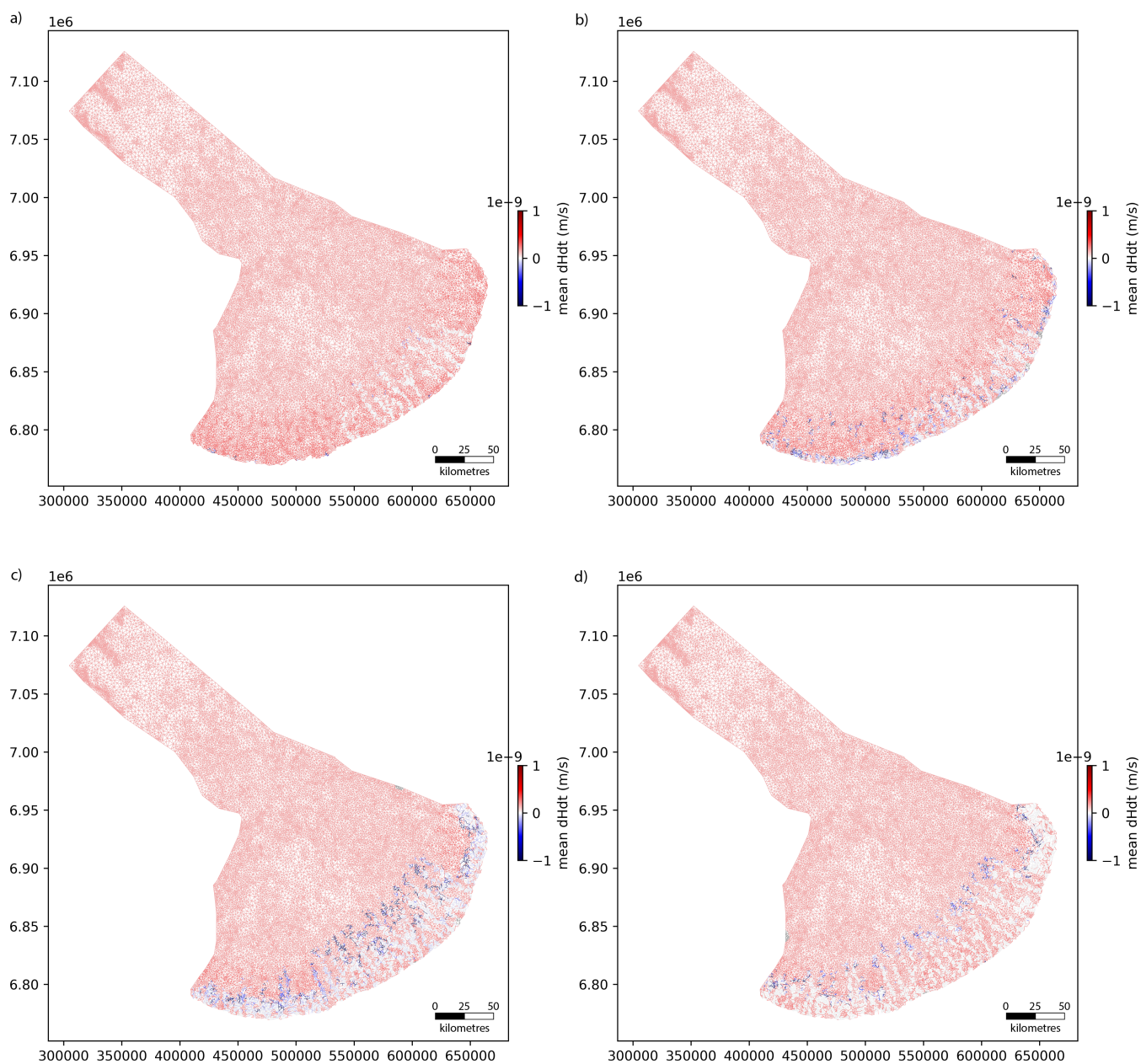


Figure A1. Ensemble mean of dH/dt across a melt season for the low- H scenario showing a) end of winter b) early spring (onset of melt season) c) mid-summer (peak hydrology flow) d) autumn (end of melt season)

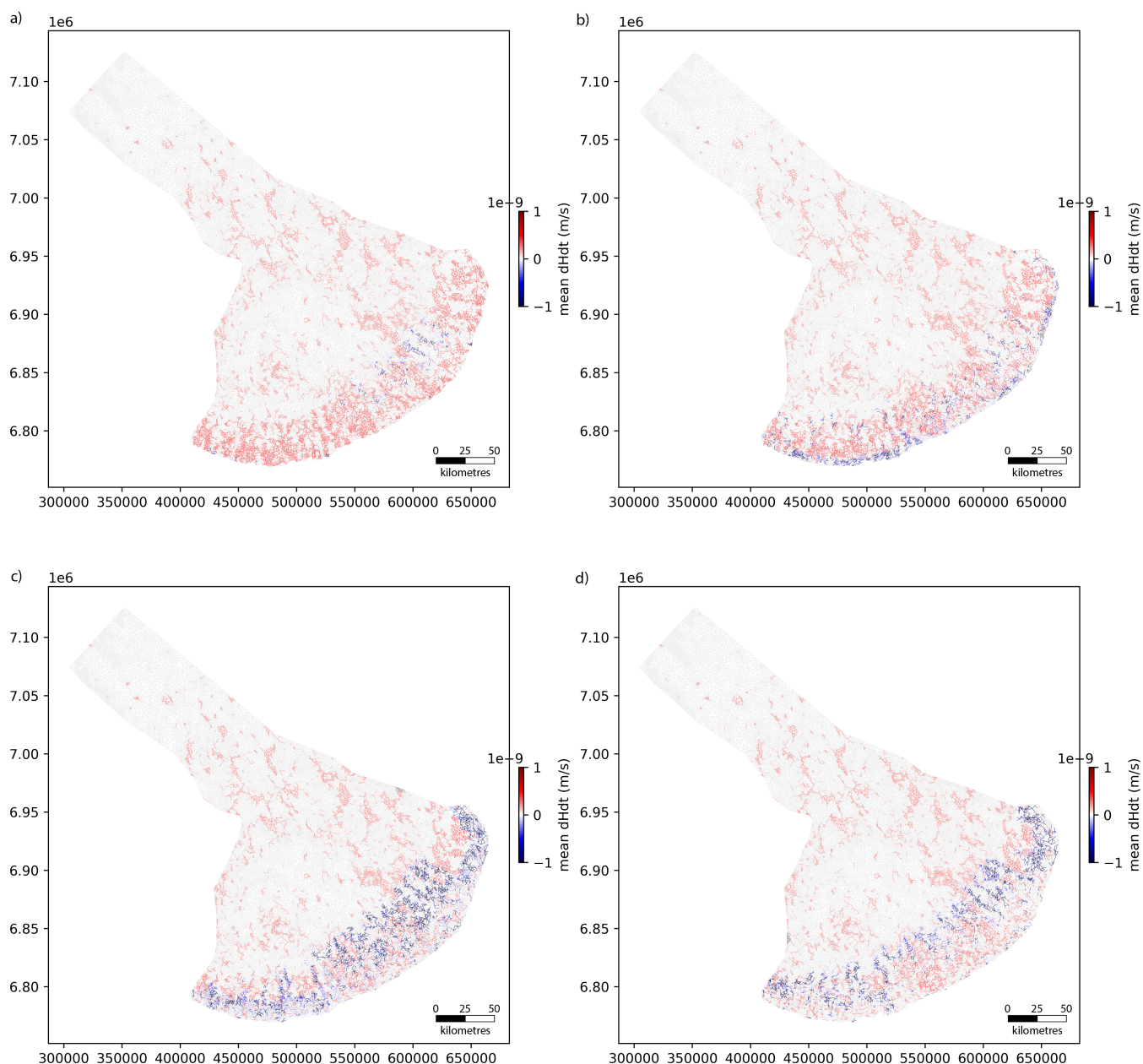


Figure A2. Ensemble mean of dH/dt across a melt season for the mixed-bed scenario showing a) end of winter b) early spring (onset of melt season) c) mid-summer (peak hydrology flow) d) autumn (end of melt season)

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Competing interests. None to declare

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