

# Efficacy of bedrock erosion by subglacial water flow

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**Abstract.** Bedrock erosion by sediment-bearing subglacial water remains little-studied, however the process is thought to contribute to bedrock erosion rates in glaciated landscapes and is implicated in the excavation of tunnel valleys and the incision of inner gorges. We adapt physics-based models of fluvial abrasion to the subglacial environment, assembling the first model designed to quantify  
5 bedrock erosion caused by transient subglacial water flow. The subglacial drainage model consists of a one-dimensional network of cavities dynamically coupled to one or several Röthlisberger channels (R-channels). The bedrock erosion model is based on the tools and cover effect, whereby particles entrained by the flow impact exposed bedrock. We explore the dependency of glacial meltwater erosion on the structure and magnitude of water input to the system, the ice geometry and the sediment  
10 supply. We find that erosion is not a function of water discharge alone, but also depends on channel size, water pressure and on sediment supply, as in fluvial systems. Modelled glacial meltwater erosion rates are one to two orders of magnitude lower than the expected rates of total glacial erosion required to produce the sediment supply rates we impose, suggesting that glacial meltwater erosion is negligible at the basin scale. Nevertheless, due to the extreme localization of glacial meltwater  
15 erosion (at the base of R-channels), this process can carve bedrock (Nye) channels. In fact, our simulations suggest that the incision of bedrock channels several centimetres deep and a few meters wide can occur in a single year. Modelled incision rates indicate that subglacial water flow can gradually carve a tunnel valley and enhance the relief or even initiate the carving of an inner gorge.

## 1 Introduction

20 Textbook descriptions of glacial erosion detail mechanisms of abrasion and quarrying, but mention erosion by subglacial meltwater as a potential, unquantified, additional incision mechanism (e.g., Bennett and Glasser, 2009; Anderson and Anderson, 2010). This imbalance reflects the deficiency in our understanding of the latter. In fact, subglacial meltwater loaded with sediment has been inferred to carve meter-scale channels in bedrock (e.g., Glasser and Bennett, 2004), often called Nye channels  
25 (N-channel; Weertman, 1972) and kilometre-scale tunnel valleys (e.g., Ó Cofaigh, 1996; Glasser and

Bennett, 2004; Jørgensen and Sandersen, 2006; Dürst Stucki et al., 2010; Kehew et al., 2012), and more recently its role has been invoked as a necessary mechanism in the carving and deepening of inner gorges (Dürst Stucki et al., 2012; Jansen et al., 2014).

Although the ability of subglacial water flow to flush subglacial sediment is well established  
30 (e.g., Humphrey and Raymond, 1994; Gurnell et al., 1996; Hallet et al., 1996; Willis et al., 1996; Alley et al., 1997; Koppes and Hallet, 2002, 2006; Riihimaki et al., 2005; Koppes and Montgomery, 2009), there has been little work quantifying subglacial sediment transport (Creys et al., 2013), and no work on bedrock erosion by subglacial meltwater. Most studies of glacial erosion that use measurements of proglacial sediment yield rely on the hypothesis that subglacial meltwater flow  
35 is the most important process removing sediment from the glacier bed (e.g., Gurnell et al., 1996; Koppes and Hallet, 2002, 2006; Orwin and Smart, 2004; Riihimaki et al., 2005). In numerical models of glacial erosion, sediment transport by subglacial water flow is usually neglected (e.g., MacGregor et al., 2000, 2009; Brocklehurst and Whipple, 2002; Tomkin and Braun, 2002; Anderson et al., 2006; Herman and Braun, 2008; Egholm et al., 2009, 2011a, b, 2012; Herman et al., 2011; Beaud  
40 et al., 2014), often under the assumption that sediment is removed instantaneously. Bedrock erosion resulting from subglacial water is likewise neglected. A better understanding of the processes that lead to the formation of tunnel valleys or inner gorges is also important for the evaluation of deep geological repositories for nuclear waste in regions facing a potential future glaciation (e.g., Iverson and Person, 2012).

45 Subglacial water flows through two main types of drainage systems: distributed and channelized. In numerical models (see Flowers (2015) for a review), a distributed drainage system is typically represented by a network of connected cavities (e.g., Lliboutry, 1968; Iken, 1981; Kamb, 1987; Schoof, 2010; Bartholomaeus et al., 2011), a macroporous sheet of sediment (e.g. Clarke, 1996; Creys and Schoof, 2009) or a water film (e.g., Weertman, 1972; Le Brocq et al., 2009). These representations  
50 reflect field observations of an increase in water pressure with discharge (e.g. Iken and Bindschadler, 1986; Nienow et al., 1998). A channelized drainage system is most often described by a single Röthlisberger channel or a network thereof (e.g., Röthlisberger, 1972; Nye, 1976; Flowers et al., 2004; Kessler and Anderson, 2004; Schoof, 2010; Hewitt et al., 2012). Water velocities are relatively high in the channelized system and, under steady-state conditions, water pressure decreases with increasing  
55 discharge (Röthlisberger, 1972). Conduits carved both in sediment and ice, so-called canals (e.g., Walder and Fowler, 1994; Ng, 2000; Kyrke-Smith and Fowler, 2014), have properties closer to those of a distributed system as pressure often increases with discharge (Walder and Fowler, 1994). In winter, the distributed drainage system evacuates most basal water, and water pressures tend to be relatively high. As surface melt becomes significant, the water input becomes too large for the  
60 distributed drainage system alone, water pressure increases and R-channels start to form. Once an efficient drainage system is established, meltwater is routed relatively quickly downstream and baseline water pressures are generally lower than in winter, with large daily fluctuations. As surface melt

decreases, channels are reduced in size by ice creep and may eventually close. We hypothesize that this cycle may have a significant effect on bedrock erosion by subglacial meltwater flow (e.g., Willis et al., 1996; Swift et al., 2005).  
65

Three main processes produce bedrock erosion in rivers: abrasion, macro-abrasion and quarrying (e.g., Whipple et al., 2000, 2013). Abrasion is the result of particles entrained by the flow (saltating or in suspension) colliding with the bedrock and is governed by the tools and cover effect, whereby particles (i.e., tools) entrained by the flow impact exposed bedrock but can also shield it if they are immobile (i.e., cover; e.g., Whipple et al., 2000, 2013; Sklar and Dietrich, 2001, 2004; Turowski et al., 2007; Lamb et al., 2008; Cook et al., 2013; Scheingross et al., 2014). Macro-abrasion and quarrying both result from dislodgement of blocks and require a relatively high joint density in the bedrock. Macro-abrasion occurs when blocks are dislodged as a result of the impact of moving particles, while quarrying is the result of dislodgement by pressure gradients caused by water  
75 flow (Whipple et al., 2000, 2013). Over highly jointed bedrock, quarrying and macro-abrasion can produce large canyons under extreme flow conditions (e.g., Bretz, 1969; Lamb and Fongstad, 2010; Baynes et al., 2015). In this study, we limit our analysis to abrasion (Whipple et al., 2000, 2013; Cook et al., 2013) and use well-established models to estimate the erosion due to total sediment load (Lamb et al., 2008) and saltating load only (Sklar and Dietrich, 2004).

Alley et al. (1997) suggest that the sediment transport capacity of an R-channel is most affected by changes in water discharge and hydraulic potential gradient and further posit that in most cases, the hydraulic potential gradient increases downstream (due to steep ice-surface slopes close to the terminus) so that the transport capacity should also increase. Dürst Stucki et al. (2012) calculate an erosional potential of subglacial water based on the hydraulic potential gradient under a valley glacier and find that the erosional potential increases toward the terminus and could explain the deepening of inner gorges during a glaciation (e.g., Jansen et al., 2014). Both studies are, however, quite speculative regarding the processes behind subglacial meltwater erosion. Creyts et al. (2013) were the first to couple subglacial water flow in a distributed drainage system, basal refreezing and sediment transport in a numerical model to explore the evolution of bed slopes adverse to ice flow close  
85 to the terminus. They demonstrate strong feedbacks between sediment deposition / entrainment and hydraulic conditions as well as the importance of daily fluctuations in water input on the sediment flux.

In proglacial studies of seasonal sediment yield, hysteresis between sediment and water discharge is often observed (e.g., Willis et al., 1996; Swift et al., 2005; Mao et al., 2014). It is usually attributed  
95 to changes in sediment availability, tapping of new areas of the bed by the developing drainage system, or increased mobilization caused by sudden changes in the subglacial hydraulic system (e.g., Willis et al., 1996). An event during which sediment transport peaks before discharge is usually defined as clockwise hysteresis and is interpreted as the manifestation of an unlimited sediment source (e.g., Mao et al., 2014). The opposite is true for counter clockwise hysteresis. In a study of

100 bedload transport by a proglacial stream in the Italian Alps, Mao et al. (2014) identify a transition  
throughout the melt-season from hysteresis dominated by clockwise events to hysteresis dominated  
by counter clockwise events. The authors infer that this transition is due to the activation of different  
sediment sources across the drainage basin.

Evidence of the erosional action of subglacial meltwater flow is widespread in formerly glaciated  
105 regions and appears in the form of N-channels, tunnel valleys and inner gorges. Tunnel valleys are  
large (few hundred metres to kilometres wide and up to tens of kilometres long) channel-like fea-  
tures found within the limits of former continental ice sheets (e.g., Ó Cofaigh, 1996; Glasser and  
Bennett, 2004; Jørgensen and Sandersen, 2006; Dürst Stucki et al., 2010; Kehew et al., 2012) or in  
Antarctica (e.g., Denton and Sugden, 2005; Rose et al., 2014) in substrata varying from loosely con-  
110 solidated sediment to bedrock. Their formation is attributed to the action of pressurized subglacial  
meltwater, and three particular mechanisms have been proposed (e.g., Ó Cofaigh, 1996; Glasser and  
Bennett, 2004; Kehew et al., 2012): (1) sediment creep toward preferential groundwater flow paths,  
(2) carving by sediment-loaded subglacial water flow or (3) erosion caused by large subglacial water  
floods.

115 Inner gorges are narrow canyons incised at the bottom of an otherwise U-shaped valley (e.g.,  
Montgomery and Korup, 2011), and are found extensively in formerly glaciated mountain ranges  
like the Alps (e.g., Montgomery and Korup, 2011; Dürst Stucki et al., 2012). The origin of inner  
gorges was originally entirely attributed to postglacial fluvial erosion, although Montgomery and  
Korup (2011) conclude that such features persist through repeated glaciations instead of being reset  
120 by glacial erosion. For example, Valla et al. (2010) find that fluvial incision rates on the order of a  
centimetre per year occurred during at least the past four thousand years in a gorge in the French  
Western Alps. Recently Dürst Stucki et al. (2012) showed that pressurized water flow is necessary  
to explain the longitudinal profile of an inner gorge in the foothills of the Alps, and Jansen et al.  
(2014) infer from cosmogenic nuclide exposure that the timing of carving of seven inner gorges in  
125 the Baltic Shield matches the timing of glacial cover. Inner gorges are therefore most likely the com-  
bined product of fluvial erosion during interglacial periods and subglacial meltwater erosion during  
glacial periods, although the importance of fluvial versus glacial conditions is probably dependent  
on surrounding topography. Nye channels, tunnel valleys and inner gorges share some characteristics  
suggesting a common genetic origin, although the specific combination of processes responsible for  
130 their evolution may differ slightly.

We implement a one-dimensional (1-D) model of subglacial water flow in which a network of  
cavities is dynamically coupled to one or a few channels. We then compute the shear stress exerted  
on the bed and use it to compute transport stages and bedrock erosion rates by abrasion caused by  
particle impacts. We compare the results of models that treat erosion by saltating particles only (Sklar  
135 and Dietrich, 2004) and erosion by both the saltating and suspended particles (Lamb et al., 2008).

We use the word “erosion” only to describe the carving of bedrock while “transport”, “mobilization” or “entrainment” refer to the movement of unconsolidated sediments.

We first perform steady-state simulations with the channelized drainage system alone to demonstrate basic model behaviour. We investigate the role of ice geometry, surface melt and sediment supply. We then introduce the coupled (distributed and channelized) hydraulic system and water forcing with sub-seasonal fluctuations to test the importance of transients in the subglacial drainage system. Finally, we discuss the implications of our results for the formation of N-channels and consequently of tunnel valleys over bedrock and the persistence of inner gorges through repeated glaciations. Our specific research questions are: 1) What are the major controls on subglacial meltwater erosion? 2) How important is subglacial meltwater erosion compared to overall glacial erosion? 3) Can ordinary seasonal melt processes lead to subglacial bedrock channel incision (and potentially the formation of an incipient tunnel valley or persistence of an inner gorge)? 4) What are the implications of the water flow regime in channels for hysteresis and sediment transport?

## 2 Modelling approach

### 2.1 Subglacial water flow

We use a 1-D model of coupled cavity and channelized drainage styled after previous models (e.g., Werder et al., 2013), with the numerical distensibility parameter of Clarke (2003) and the water exchange term of Hewitt and Fowler (2008). The system of equations describing R-channels admits both a wave-like solution and the solution for water flow which introduces numerical stiffness. To circumvent this problem, Clarke (2003) proposed to treat water as a slightly compressible fluid and he introduced a numerical distensibility parameter,  $\gamma$ . The water exchange term is required as the channelized and distributed drainage systems are otherwise defined as two independent systems (Flowers et al., 2004; Hewitt and Fowler, 2008).

#### 2.1.1 Channelized drainage

As in most subglacial drainage models, we assume the channelized system to be a network of semi-circular R othlisberger channels (R othlisberger, 1972; Nye, 1976). The channels are assumed to be linear (along the  $x$  coordinate) and parallel to one another if more than one is considered. The conservation of water mass is given by (Clarke, 2003):

$$-\gamma S \frac{\partial \phi_{\text{ch}}}{\partial t} = \frac{\partial Q_{\text{ch}}}{\partial x} + \frac{\Xi - \Pi}{L} \left( \frac{1}{\rho_i} - \frac{1}{\rho_w} \right) - v_{\text{cc}} - (\dot{b}_{\text{ch}} + \varphi)W, \quad (1)$$

where  $S$  is the cross-sectional area of the R-channel,  $\phi_{\text{ch}}$  is the hydraulic potential,  $Q_{\text{ch}}$  is the water discharge,  $\rho_w$  and  $\rho_i$  are the densities of water and ice respectively,  $v_{\text{cc}}$  is the rate of creep closure of the channel walls,  $\dot{b}_{\text{ch}}$  is a water source term to the channel,  $\varphi$  is the exchange rate between the distributed and channelized drainage systems,  $W$  is the width of the catchment drained by the R-

channel,  $\gamma$  is a numerical compressibility parameter (Clarke, 2003),  $x$  is the coordinate along the  
 170 flow line and  $t$  is time. Note that we assume changes in bed topography with time are small enough  
 to be neglected. Therefore  $\partial\phi_b/\partial t = 0$  and  $\partial p_{ch}/\partial t = \partial(\phi_{ch} - \phi_b)/\partial t = \partial\phi_{ch}/\partial t$ , where  $p_{ch}$  is the  
 water pressure in the channel and  $\phi_b$  the hydraulic potential at the bed. The evolution of channel  
 cross-sectional area  $S$  with time is

$$\frac{\partial S}{\partial t} = \frac{\Xi - \Pi}{\rho_i L} - v_{cc}, \quad (2)$$

175 where  $\Xi$  is the dissipation of potential energy,  $\Pi$  is the energy to maintain the water at the pressure  
 melting point,  $L$  is the latent heat of fusion, thus  $(\Xi - \Pi)/(\rho_i L)$  is the rate of channel opening by  
 viscous heat dissipation. We refer the reader to the Supplement for the description of  $\Xi$ ,  $\Pi$ ,  $v_{cc}$  and  
 $Q_{ch}$ .

### 2.1.2 Distributed drainage

180 The distributed drainage system is treated as a network of connected cavities (Kamb, 1987) under-  
 going turbulent flow. Assuming that water can be stored englacially or subglacially, the conservation  
 of water mass can be written (Werder et al., 2013)

$$\frac{e_v}{\rho_w g} \frac{\partial \phi_{ca}}{\partial t} + \nabla \cdot \mathbf{q}_{ca} + v_o - v_c - (\dot{b}_{ca} - \varphi) = 0 \quad (3)$$

185 where  $\phi_{ca}$  is the hydraulic potential in the cavity network,  $e_v$  is the englacial void ratio,  $\mathbf{q}_{ca}$  is the  
 water flux in the cavity network,  $g$  is the gravitational acceleration,  $v_o$  and  $v_c$  are the opening and  
 closure rate, respectively,  $\dot{b}_{ca}$  is a water source term to the cavity network. The evolution of the  
 average cavity height  $h_{ca}$  with time is:

$$\frac{\partial h_{ca}}{\partial t} = v_o - v_c. \quad (4)$$

We refer the reader to the Supplement for the description of  $v_o$ ,  $v_c$  and  $\mathbf{q}_{ca}$ .

### 190 2.1.3 Coupled channelized and distributed drainage

To couple the two drainage systems we use a term to describe water exchange between cavities and  
 channels as a function of their respective pressure differences (e.g., Flowers et al., 2004; Hewitt and  
 Fowler, 2008):

$$\varphi = k_{ex}(N_{ch} - N_{ca}), \quad (5)$$

195 where  $k_{ex}$  is an exchange coefficient (Table 1) and  $N_{ch}$  and  $N_{ca}$  are the effective pressures in the  
 channelized and cavity systems, respectively (see Supplement). The system formed by Eqs. (1), (2),  
 (3) and (4) is solved with the MATLAB pdepe solver.

**Table 1.** Summary of hydrological model parameters (see Supplement for extended list).

Parameter	Description	Value
$M$	Number of grid points	101
$dx$	Cell size	500 m
$dt$	Time step	600 s
$W$	System width	1000 m
$X_L$	Glacier length	50 km
$\rho_w$	Water density	$1000 \text{ kg m}^{-3}$
$\rho_i$	Ice density	$910 \text{ kg m}^{-3}$
$L$	Latent heat of fusion	$3.34 \times 10^5 \text{ J kg}^{-1}$
$n$	Flow-law exponent	3
$k_{\text{ex}}$	Exchange coefficient	$5 \times 10^{-13} \text{ m}^{3/2} \text{ kg}^{-1/2}$
$e_v$	Englacial void ratio	$1 \times 10^{-3}$
$\gamma$	Numerical compressibility	$10^{-9} \text{ Pa}^{-1}$

#### 2.1.4 Shear stress partitioning

Following Clarke (2003) we write the shear stress exerted on the bed by subglacial water flow as:

$$200 \quad \tau_b = \frac{1}{8} f_b \rho_w u^2, \quad (6)$$

where  $f_b$  is the Darcy-Weisbach friction coefficient for the bed and  $u$  is the depth-averaged water flow velocity defined as  $u = Q_{\text{ch}}/S$  in a channel and  $u = q_{\text{ca}}/h_{\text{ca}}$  in the network of cavities. Details pertaining to the shear stress calculation are described in the Supplement.

#### 2.1.5 Subglacial water flow model simplifications

205 We implement the model in a 1-D continuum such that the drainage system is assumed to be fully connected. Distributed and channelized drainage systems are assumed to be saturated. The bed is impermeable and undeformable. We also neglect the routing of supra- and englacial water, the effect of particles or refreezing on water flow constriction and the feedbacks between sliding speed and water pressure.

#### 210 2.1.6 Boundary and initial conditions

Atmospheric pressure defines the downstream boundary condition for both the network of cavities and the R-channels:  $\phi_{\text{ch}}(x = X_L, t) = \phi_b + 1000 \text{ Pa}$ . For most simulations, a no-flux condition is applied at the upstream boundary such that  $Q_{\text{ch}}(x = 0, t) = 0$ ,  $q_{\text{ca}}(x = 0, t) = 0$ . Otherwise, for the simulation in which discharge is constant throughout the domain, we apply a Neumann boundary  
 215 condition at the upstream node of the channel:  $Q_{\text{ch}}(x = 0, t) = 18 \text{ m}^3 \text{ s}^{-1}$ . We use a one-year model

spin-up for all the transient simulations as we find no significant difference in results using one- and two-year spin-ups.

## 2.2 Erosion by saltating and total load

In river reaches where bedrock fracture density is low (i.e. blocks are larger than 1–2 metres in size), abrasion is the primary erosional process (Whipple et al., 2000, 2013). Particles that are mobilized by the flow and move by saltation or in suspension can impact the bed, and the energy released upon impact can cause erosion. To erode its bed, a river therefore requires tools (particles available for transport) and an exposed bed.

### 2.2.1 Saltation erosion model (SEM) (Sklar and Dietrich, 2004)

The rate of erosion caused by a saltating load can be expressed as:

$$\dot{e}_{\text{salt}} = I_r V_i F_e, \quad (7)$$

where  $I_r$  is the rate of particle impact with the bed,  $V_i$  is the volume of bed material removed upon impact and  $F_e$  the fraction of the bed exposed. The impact rate ( $I_r$ ) is a function of the number of particles and their saltation trajectories. The more particles and the shorter the saltation length, the higher the impact rate. The volume removed upon impact ( $V_i$ ) is a function of the energy released on impact and therefore of the particle mass, its speed normal to the bed and the physical properties of the bed and the particle. The fraction exposed ( $F_e$ ) determines the areal extent of bedrock vulnerable to erosion and represents the cover effect. All the quantities described here are given per unit width. Values of  $V_i$ ,  $I_r$  and  $F_e$  are computed as:

$$V_i = \frac{\pi \rho_s D^3 w_{si}^2 Y}{6 k_v \sigma_T^2}, \quad (8)$$

where  $\rho_s$  is the sediment density,  $D$  is the particle diameter,  $w_{si}$  is the particle velocity upon impact,  $Y$  is the Young's modulus of the bedrock,  $k_v$  is an empirical rock erodibility coefficient and  $\sigma_T$  is the rock tensile strength;

$$I_r = \frac{6 q_s}{\pi D^3 L_s}, \quad (9)$$

where  $q_s$  is the sediment supply rate and  $L_s$  is the hop length of particles;

$$F_e = 1 - \frac{q_s}{q_{tc}}, \text{ for } q_s \leq q_{tc}, \quad (10)$$

where  $q_{tc}$  is the sediment transport capacity. Substituting Eqs. (8), (9) and (10) into (7) and simplifying yields:

$$\dot{e}_{\text{salt}} = \frac{\rho_s q_s w_{si}^2 Y}{L_s k_v \sigma_T^2} \left( 1 - \frac{q_s}{q_{tc}} \right). \quad (11)$$



245 Note that when  $q_s/q_{tc} > 1$  the erosion rate becomes  $\dot{e}_{\text{salt}} = 0$  because the bed is completely covered with sediment. Constants and parameters are listed in Table 2, and the formulations of  $L_s$  and  $w_{\text{si}}$  are described in the Supplement. We follow Sklar and Dietrich (2004) in using the Fernandez-Luque and van Beek (1976) formulation of transport capacity:

$$q_{tc} = 5.7(rgD^3)^{1/2}(\tau^* - \tau_c^*)^{3/2}, \quad (12)$$

250 where  $r = \rho_s/\rho_w - 1$  is the buoyant density of sediment and  $\tau_c^*$  the critical value of the Shields stress. The Shields stress  $\tau^*$  is computed as:

$$\tau^* = \frac{\tau_b}{(\rho_s - \rho_w)gD}. \quad (13)$$

### 2.2.2 Total load erosion model (TLEM) (Lamb et al., 2008)

The total load erosion model (TLEM) is based on the original model of Sklar and Dietrich (2004) and extended by Lamb et al. (2008) to account for the impact of particles in suspension close to the bed. The total load erosion rate is calculated as:

$$\dot{e}_{\text{tot}} = \frac{A_1 \rho_s Y}{k_v \sigma_T^2} c_b w_{i,\text{eff}}^3 \left(1 - \frac{q_b}{q_{tc}}\right), \quad (14)$$

where  $A_1 < 1$  is a coefficient to account for the upward lifting of particles close to the bed by turbulent eddies (here  $A_1 = 0.36$ ),  $c_b$  is the near-bed sediment concentration,  $w_{i,\text{eff}}$  is the effective impact velocity, and  $q_b$  is the volumetric flux of sediment transported as bed-load. The calculation of these quantities is explained in the Supplement. By analogy with the saltation-erosion model,  $c_b$  is similar to the impact rate  $I_r$ ,  $w_{i,\text{eff}}$  to the volume removed upon impact  $V_i$ , and  $(1 - q_b/q_{tc})$  to the fraction of the bed exposed ( $F_e$ ). Similarly to the SEM, when  $q_b/q_{tc} > 1$  the erosion rate becomes  $\dot{e}_{\text{tot}} = 0$  because the bed is completely covered with sediment. Constants and parameters used for the TLEM are the same as in the SEM and are listed in Table 2.

265 The major differences between the TLEM and the SEM are that: (1) the effective impact velocity  $w_{i,\text{eff}}$  (TLEM) increases monotonically with transport stage as it accounts for the effect of turbulence causing particle collisions with the bed whereas  $w_{\text{si}}$  tends to 0 at large transport stages (and therefore  $\dot{e}_{\text{salt}}$  approaches 0) as particles trajectories become parallel to the bed, and (2) the fraction of the bed exposed in the TLEM ( $1 - q_b/q_{tc}$ ) uses the bed load flux ( $q_b$ ) as contributing to the cover. This means that as the transport stage increases, the fraction of bedrock exposed in the TLEM is higher than in the SEM, provided a significant portion of the sediment is transported in suspension.

### 2.2.3 Erosion model simplifications

We calculate erosion rates with a constant particle diameter ( $D = 60$  mm) and assume that this particle size is representative of the median grain size in an R-channel. This choice is consistent with Riihimaki et al. (2005) who report a median grain size corresponding to very coarse gravel in the

**Table 2.** Summary of parameter values for the saltation (Sect. 2.2.1) and total load (Sect. 2.2.2) erosion models.

Parameter	Description	Value
$A_1$	Coefficient for upward lifting of particles <sup>†</sup>	0.36
$D$	Diameter of particles	60 mm
$\sigma_T$	Rock tensile strength <sup>†</sup>	7.0 MPa
$Y$	Young’s modulus of bedrock <sup>†</sup>	$5.0 \times 10^4$ MPa
$k_v$	Rock erodibility coefficient <sup>†</sup>	$1.0 \times 10^6$
$\tau_c^*$	Critical Shields stress <sup>†</sup>	0.03
$\rho_s$	Sediment density	$2650 \text{ Kg m}^{-3}$
$q_{s,\text{ref}}$	Ref. sediment supply per unit width (Steady state – transient)	$3.6 - 9.1 \times 10^{-3} \text{ m}^2 \text{ s}^{-1}$

<sup>†</sup>Parameter values from Lamb et al. (2008)

proglacial area of Bench Glacier, Alaska. We assume that the bedrock has a uniform resistance to erosion, erosion occurs uniformly across the width of a channel, abrasion is the main erosion mechanism and that the cover effect is linear. A linear cover effect means that as long as the bed is partially exposed, newly deposited particles are assumed to cover bedrock rather than previously deposited sediment, hence the exponent of one on the “cover” fraction  $q_s/q_{tc}$  (cf. Turowski et al., 2007). We also neglect sediment transport and the effect of downstream fining.

### 3 Modelling strategy and rationale

The model outlined above involves numerous variables and parameters leading to feedbacks. Applications of the model of erosion by abrasion in rivers (e.g. Sklar and Dietrich, 2004, 2006, 2008; Turowski et al., 2007; Lamb et al., 2008; Chatanantavet and Parker, 2009; Nelson and Seminara, 2011; Egholm et al., 2013) show that the primary dependencies are: the transport stage, the relative sediment supply and the hydraulic potential gradient. Numerical modelling studies of subglacial water flow emphasize the importance of the frequency and amplitude of the water input forcing and of the ice and bed geometry (e.g., Flowers, 2008; Creyts and Schoof, 2009; Schoof, 2010; Hewitt, 2013; Werder et al., 2013; Beaud et al., 2014), although few studies have discussed the shear stress on channel walls (e.g., Clarke, 2003). In an effort to identify the key variables, parameters and feedbacks, we start with simple experiments and build up the complexity. The simulations are separated in two subsections: (1) steady state decoupled simulations of a channelized drainage system only (Table 3) and (2) transient simulations of a coupled channelized and distributed drainage system (Table 4).

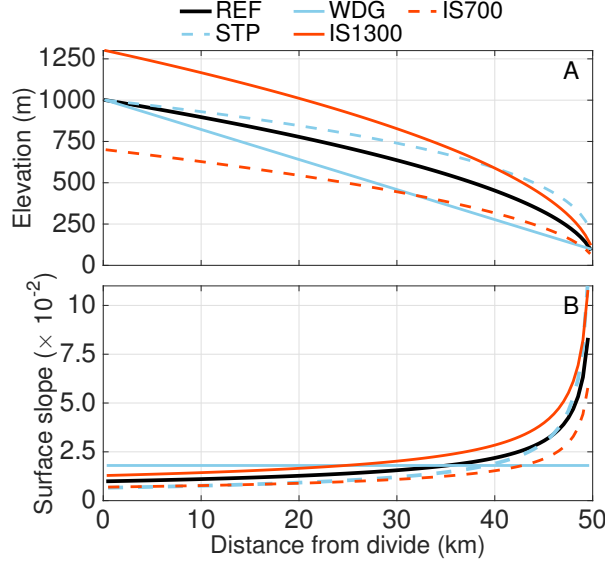
In the steady state simulations (Table 3), we first assess the erosion pattern resulting from a constant discharge along the glacier bed. Then we introduce a water forcing that increases with decreasing ice-surface elevation and test the effect of ice geometry and sediment supply. Additional  
300 experiments assessing the effect of water flow through a network of cavities, water input, sediment size, and channel wall and bed roughness are shown in the Supplement. In the transient simulations (Table 4) we first analyze the role of a synthetic melt season, then use water forcing following realistic melt seasons, test the role of ice geometry, channel density and sediment supply.

As per the theory on which the SEM (Sklar and Dietrich, 2004) and TLEM (Lamb et al., 2008) are  
305 based, sediment supply, through the tools and cover effects, exerts a major control on erosion rates and patterns by water flow beneath glaciers. Although it is possible to estimate sediment supply to rivers (e.g., Gurnell et al., 1996; Willis et al., 1996; Kirchner et al., 2001; Orwin and Smart, 2004; Sklar and Dietrich, 2004; Riihimaki et al., 2005; Turowski et al., 2009; Mao et al., 2014), no method exists to quantify subglacial sediment supply or transport. In glacierized catchments, measurements  
310 are usually made in the proglacial stream, relatively close to the terminus (e.g., Warburton, 1990; Gurnell et al., 1996; Willis et al., 1996; Orwin and Smart, 2004; Riihimaki et al., 2005; Swift et al., 2005; Mao et al., 2014), where water flow is subaerial and potentially influenced by channel dynamics between the glacier terminus and the measurement station (Warburton, 1990; Orwin and Smart, 2004; Mao et al., 2014). In the steady-state simulations we impose a sediment supply that leads to an  
315 interesting and diverse range of simulations illustrating most processes and their feedbacks, whereas in the transient simulations we choose a sediment supply that leads to a sediment yield at the last node equivalent to a few millimetres of glacial erosion per year.

We use a flat bed and a parabolic surface for all but two geometries: STP and WDG (Fig. 1). The  
STP geometry aims at reproducing the steep front of an advancing ice sheet, while the WDG geom-  
320 etry has a wedge shape that resembles the profile of a thinning and retreating ice sheet margin. We assume that water input is distributed uniformly (no moulins) along the bed except for the simulation S\_MOULIN. For almost all simulations we fix the sediment supply per unit width; the total sediment supply therefore increases with channel size. This is similar to the assumption of a uniform till distribution across the width of the glacier in which the channelized water flow sources its sediment.

## 325 4 Results

All quantities related to the erosion model are given per unit width of flow. The cross-sectional area of a subglacial channel ( $S$ ), thus its width ( $W_{\text{ch}} = 2\sqrt{2S/\pi}$ ), changes more rapidly with distance along the flow path than a typical subaerial river. We need to account for these changes when displaying the results, and therefore introduce three quantities: the total transport capacity  
330  $Q_{\text{tc}} = q_{\text{tc}}W_{\text{ch}}$  ( $\text{m}^3\text{s}^{-1}$ ), the total erosion computed with the TLEM  $E_{\text{tot}} = \dot{e}_{\text{tot}}W_{\text{ch}}$  ( $\text{m}^2\text{s}^{-1}$ ) and the total erosion computed with the SEM  $E_{\text{salt}} = \dot{e}_{\text{salt}}W_{\text{ch}}$  ( $\text{m}^2\text{s}^{-1}$ ).



**Figure 1.** (A) Different ice geometries considered (see Tables 3 and 4) and (B) corresponding ice-surface slopes. The geometries REF, IS1300 and IS700 are parabolas such that the surface elevation is given by  $z_s(x) = z_{s,\max}\sqrt{x}$ , where  $z_{s,\max}$  is the maximum ice-surface elevation and  $x \in [1, 0]$ . In order to obtain the steeper front in STP we use a cubic root instead of the square root:  $z_s(x) = z_{s,\max}x^{1/3}$ . The wedge-like geometry WDG is defined by a straight line with the same elevation change as REF.

**Table 3.** Summary of steady state simulations. For simulations in which meltwater input is a function of ice-surface elevation  $z_s$ , we compute  $f(z_s(x)) = \dot{b}_{ss,\max} \times (1 - (z_s(x) - z_{s,\min})/z_{s,\max})$ , where  $\dot{b}_{ss,\max} = 8.5 \times 10^{-7} \text{ m s}^{-1}$  is the maximum melt water input rate to the channelized drainage system, and  $z_{s,\min}$  and  $z_{s,\max}$  are respectively the minimum and maximum ice-surface elevations. Note that  $\dot{b}_{ss,\max} = 8.5 \times 10^{-7} \text{ m s}^{-1}$  corresponds to 7.6 cm of ice melt per day assuming  $\rho_i = 910 \text{ kg m}^{-3}$ . The reference sediment supply used for the steady state simulations is:  $q_{s,\text{ref}} = 3.6 \times 10^{-3} \text{ m}^2 \text{ s}^{-1}$  (Table 2).

Simulation	Purpose	Forcing	Difference from reference run	Section
S_MOULIN	R-channel only	$Q_{\text{ch}}(x=0, t) = 4.25 \text{ m}^3 \text{ s}^{-1}$	Localized input upstream boundary	4.1.1
S_REF	Reference	$\dot{b}_{\text{ch,ref}}(x, t) = f(z_s(x))$		4.1.2
S_WDG	Ice geometry	$\dot{b}_{\text{ch}}(x, t) = \dot{b}_{\text{ch,ref}}$	Constant ice-surface slope $d\phi_0/dx = \text{cst}$	4.1.2
S_STP	Ice geometry	$\dot{b}_{\text{ch}}(x, t) = \dot{b}_{\text{ch,ref}}$	Steeper and thicker terminus	4.1.2
S_1300	Ice geometry	$\dot{b}_{\text{ch}}(x, t) = \dot{b}_{\text{ch,ref}}$	$z_{s,\max} = 1300 \text{ m}$	4.1.2
S_700	Ice geometry	$\dot{b}_{\text{ch}}(x, t) = \dot{b}_{\text{ch,ref}}$	$z_{s,\max} = 700 \text{ m}$	4.1.2
S_SSP	Sediment supply	$\dot{b}_{\text{ch}}(x, t) = \dot{b}_{\text{ch,ref}}$	$q_s = q_{s,\text{ref}}/20 - q_{s,\text{ref}} \times 25$	4.1.3

**Table 4.** Summary of transient simulations. In this series of simulations the basal sliding speed is  $w_b = 5 \text{ m a}^{-1}$  and the reference sediment supply is  $q_{s,\text{ref}} = 9.1 \times 10^{-3} \text{ m}^2 \text{ s}^{-1}$  (Table 2). Water is fed to the network of cavities rather than the channel, hence  $\dot{b}_{\text{ch}} = 0 \text{ m s}^{-1}$ , and the function of surface elevation is that of the steady state simulations (Table 3).

Simulation	Purpose	Description	Section
T_REF	Reference transient	Reference geometry and synthetic water input	4.2.1
T_2007	Realistic forcing	2007 surface-melt time series	4.2.2
T_2008	Realistic forcing	2008 surface-melt time series	4.2.2
T_1300	Ice geometry	$z_{s,\text{max}} = 1300 \text{ m}$	4.2.3
T_700	Ice geometry	$z_{s,\text{max}} = 700 \text{ m}$	4.2.3
T_W500	Drainage catchment width	$W = 500 \text{ m}$ , i.e. 2 R-channels	4.2.4
T_W333	Drainage catchment width	$W = 1000/3 \text{ m}$ , i.e. 3 R-channels	4.2.4
T_W250	Drainage catchment width	$W = 250 \text{ m}$ , i.e. 4 R-channels	4.2.4
T_SSP/4	Sediment supply	$q_s = q_{s,\text{ref}}/4$	4.2.5
T_SSP/2	Sediment supply	$q_s = q_{s,\text{ref}}/2$	4.2.5
T_SSPOPT	Sediment supply	Optimized erosion; $q_s = 0.6 \times q_{tc}$	4.2.5

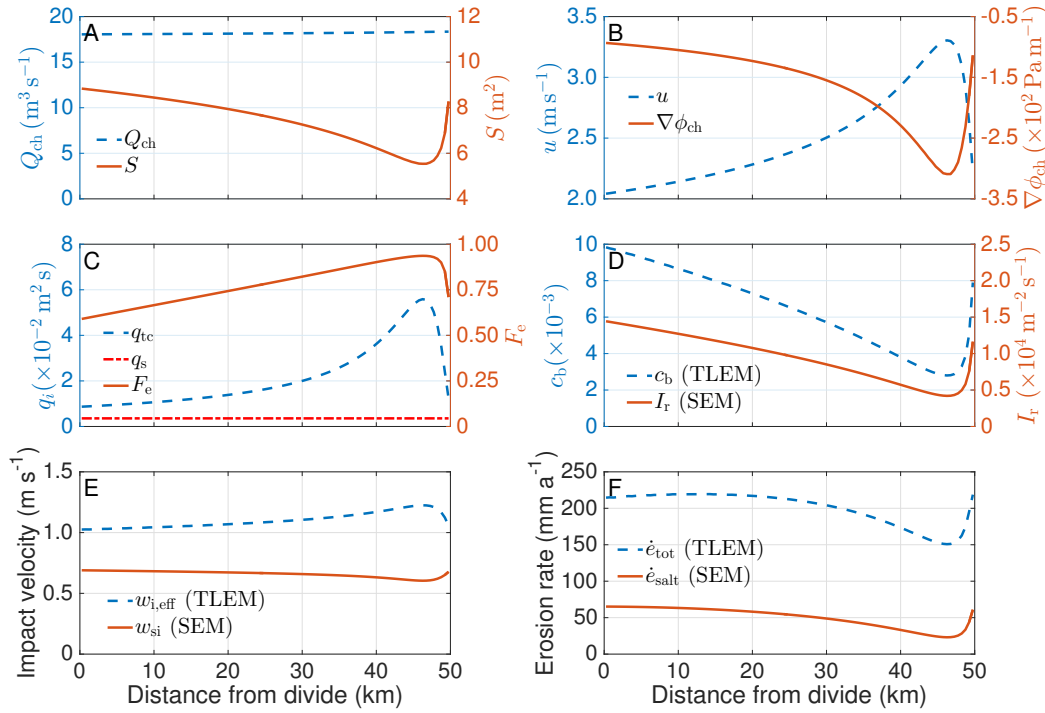
#### 4.1 Steady state decoupled simulations

We examine basic steady-state behaviour of key model variables such as R-channel cross-sectional area ( $S$ ), transport capacity ( $q_{tc}$ ), impact velocity ( $w_{si}$  and  $w_{i,\text{eff}}$ ), transport stage ( $\tau^*/\tau_c^*$ ) and relative sediment supply ( $q_s/q_{tc}$ ). In the steady state simulations, the drainage system is composed of a single R-channel. Simulations are terminated once dependent variables ( $S$  and  $\phi_{\text{ch}}$ ) reach steady state. In this series of simulations, unless stated otherwise, we impose the sediment supply  $q_s = 3.6 \times 10^{-3} \text{ m}^2 \text{ s}^{-1}$  to produce an interesting and diverse range of model behaviour.

##### 4.1.1 R-channel with constant discharge

In the S\_MOULIN simulation we use the reference glacier geometry (Fig. 1, REF) and the water input is imposed at the uppermost node only, i.e. as if a moulin were feeding the system. This permits us to drive the system with a constant water discharge and to analyse the resulting relation between discharge, channel cross-sectional area, velocity, instantaneous erosion rates and transport capacity patterns.

Although discharge is constant (Fig. 2a), the cross-sectional area of the R-channel changes along the profile (Fig. 2a) due to the ice geometry (Fig. 1). Over the first 46 km, the cross-sectional area decreases in response to the steepening hydraulic potential gradient (Fig. 2b), the latter being a function of ice-surface slope (Fig. 1). Close to the terminus (last 4 km of the profile), the ice thins significantly, the hydraulic potential gradient shallows (by a factor 3) and the cross-sectional area increases. The



**Figure 2.** Relationship between the hydraulic and erosion variables in a steady state R-channel with constant discharge (Table 3, S\_MOULIN; Sect. 4.1.1). Components of the TLEM and SEM models are also compared. (A) R-channel discharge,  $Q_{ch}$  and cross-sectional area  $S$ ; (B) Velocity,  $u$  and gradient in hydraulic potential,  $\nabla\phi_{ch}$ ; (C) Sediment transport capacity,  $q_{tc}$ , prescribed sediment supply rate,  $q_s$  and fraction of bed exposed,  $F_e$ ; (D) Near-bed sediment concentration,  $c_b$  (TLEM) and impact rate,  $I_r$  (SEM); (E) Impact velocity in the TLEM,  $w_{i,eff}$  and the SEM,  $w_{si}$ ; (F) Erosion rate calculated with the TLEM,  $\dot{e}_{tot}$  and the SEM,  $\dot{e}_{salt}$ .

350 average water velocity assumes the opposite pattern (Fig. 2b). Because the grain size is kept constant,  $q_{tc} \propto (\tau^* - \tau_c^*)^{3/2}$ , the rate of sediment transport (Fig. 2c) is amplified relative to the velocity ( $q_{tc}$  drops by a factor 6 while  $u$  is reduced by about 50%); both have maxima at 46 km and decrease sharply near the terminus. In this simulation the sediment transport capacity is always larger than the supply rate (Fig. 2c), exposing most of the bed ( $F_e > 0.5$ ).

355 In both the TLEM and SEM, near-bed sediment concentration and impact rate are described as a function of the sediment supply and the hop trajectory of a particle, so they are similar (Fig. 2d). As the velocity increases (km 0–46) the sediment is transported faster and further from the bed and the near-bed sediment concentration and impact rate decrease.

Impact velocities (Fig. 2e) vary depending on the model. In the SEM the impact velocity tends  
 360 toward zero as the hop length increases and the particles approach transport in suspension, which leads to a local minimum around km 46. The TLEM accounts for the effect of turbulent eddies on

the trajectory of particles close to the bed, thus the effective impact velocity is commensurate with the velocity and peaks around km 46.

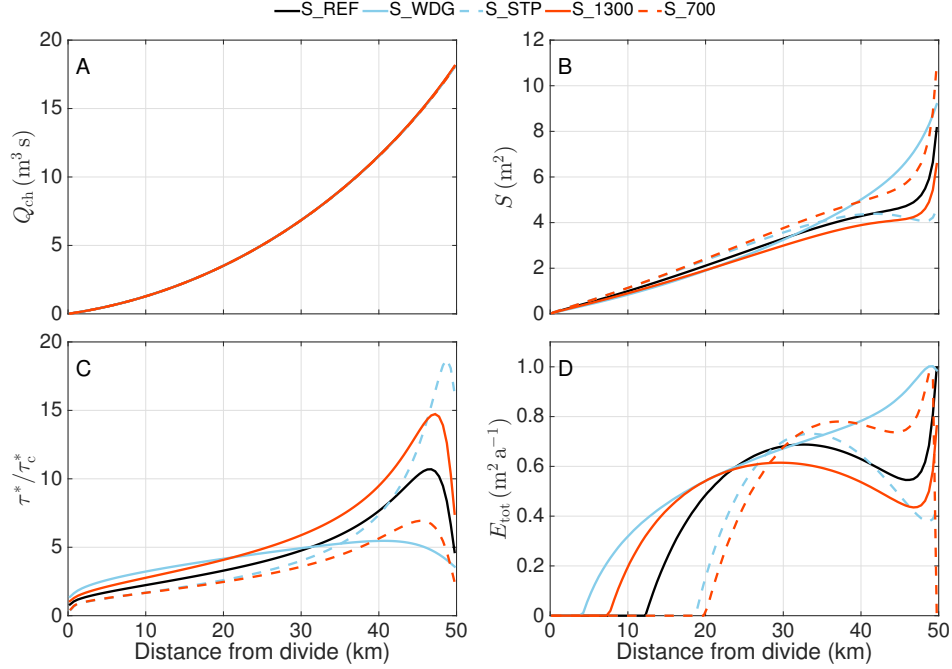
Erosion rates (Fig. 2f) in both the SEM and TLEM show a minimum at 46 km and a maximum close to the terminus. The minimum and maximum correspond respectively to the minimum and sharp rise in near-bed sediment concentration or impact rate. Even under a constant discharge, ice surface slope and channel size produce a peak in velocity just upstream of the terminus. At this peak, the flow has the power to lift particles far enough from the bed to reduce erosion in both the SEM and TLEM. Erosion rates are higher with the TLEM than the SEM due to the difference in impact velocities when the sediment transport regime approaches suspension. For the transport stages we obtain, impact velocities with the TLEM are consistently higher than with the SEM. Moreover for relatively large transport stages, more of the bed is exposed in the TLEM: when some fraction of the total load travels in suspension:  $q_s > q_b$ .

#### 4.1.2 Ice geometry

The surface slope of a glacier is a first-order control on subglacial water flow (Eqs. (1)–(2)). We experiment with thicker ice (IS1300, Fig. 1), thinner ice (IS700, Fig. 1) and steeper ice-surface slopes close to the terminus (STP, Fig. 1), all compared to the reference ice geometry (REF, Fig. 1). Shallower surface slopes close to the terminus are tested with a wedge-like geometry of constant surface slope (WDG, Fig. 1). All these simulations (Fig. 3; see Table 3) employ the surface melt profile of S\_REF and therefore yield nearly identical discharge profiles (Fig. 3a). The water input is a function of the ice-surface elevation (Table 3), hence the discharge is largest close to the terminus.

The thinner the ice close to the terminus the larger the channel (Fig. 3b) and the lower the transport stage (Fig. 3c). In the case of S\_STP, the combination of a particularly steep hydraulic potential gradient and relatively thick ice near the terminus inhibits channel enlargement over the last 5–10 km as observed in other simulations (Fig. 3b). Since the discharge profile is the same for the simulations presented, the smaller the channel, the faster the flow and hence the larger the transport stage. The maximum transport stage for the simulation with the steepest terminus (S\_STP,  $\tau^*/\tau_c^* \approx 18.5$ , Fig. 3c) is almost four times larger than that for the simulation with the shallowest terminus (S\_WDG,  $\tau^*/\tau_c^* \approx 5.5$ , Fig. 3c; see Fig. 1 and Table 3). Over the first 25 km of the profile the simulation with a constant ice-surface slope (S\_WDG) shows the steepest hydraulic potential gradient, creating comparatively higher transport stages than other models.

Erosion begins around km 4 for S\_WDG (Fig. 3d), and around km 19–20 for S\_STP and S\_700 (Fig. 3d), because for the prescribed sediment supply, the bed becomes exposed for transport stages  $\tau^*/\tau_c^* \gtrsim 2.5$ . All simulations but S\_WDG have a local maximum in total erosion ( $E_{tot}$ , Fig. 3d) between km 30 and 37 and a local minimum between km 45 and 49. The sediment supply per unit width is constant, therefore the relative sediment supply decreases as the transport stage increases and thus the erosion rate per unit width decreases when  $q_s/q_{tc} < 0.5$  because the number of tools decreases.



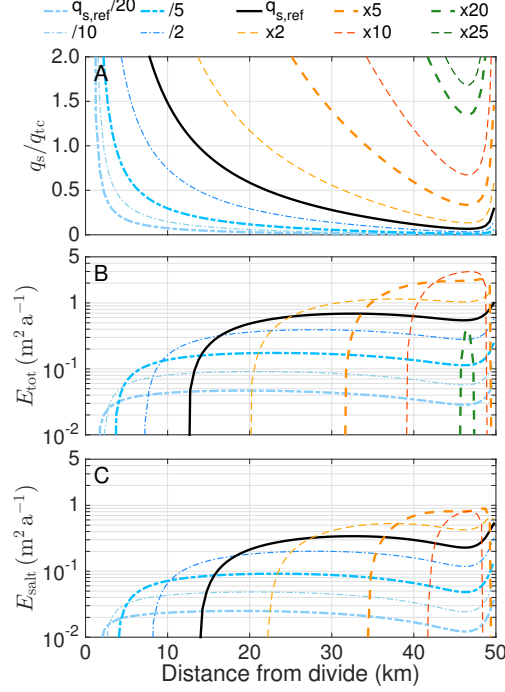
**Figure 3.** Comparison of steady-state simulations with varying ice geometries, reference water input and drainage through a single R-channel (Table 3). (A) Discharge in the channel,  $Q_{\text{ch}}$  (all curves overlap); (B) Channel cross-sectional area,  $S$ ; (C) Transport stage,  $\tau^*/\tau_c^*$ ; (D) Total erosion computed with the TLEM ( $E_{\text{tot}}$ ).

For simulation S\_WDG the increase in channel size compensates for the small decrease in erosion rate per unit width (not shown) because the transport stage remains relatively low ( $\tau^*/\tau_c^* < 6$ ). Over the last few kilometres of the profile, total erosion increases again for the simulations in which transport stage drops to moderate values (S\_REF and S\_1300). Total erosion drops sharply at the terminus for S\_700 as the transport stage drops below 2.5. When the transport stage remains relatively high ( $\tau^*/\tau_c^* > 15$ ) the number of tools remains low, as does total erosion (S\_STP, Fig. 3c and 3d). Ice geometry exerts a primary influence on transport stage and erosion patterns via its control on hydraulic potential gradients and channel size. Simulation S\_WDG yields the most erosion, despite relatively low transport stages, illustrating the importance of the tools effect (e.g., Sklar and Dietrich, 2006).

### 4.1.3 Sediment supply

Tools and cover compete so that both a lack and an overabundance of tools hinder erosion. Sklar and Dietrich (2004) and Lamb et al. (2008) have shown that erosion peaks for a given flow regime at an optimum relative sediment supply. We thus investigate how varying sediment supply ( $q_s = 1.8 \times 10^{-4}$  to  $8.9 \times 10^{-2} \text{ m}^2 \text{ s}^{-1}$ ) affects the rates and patterns of subglacial meltwater erosion, while the subglacial hydraulic regime remains that of S\_REF (Fig. 3; Table 3).





**Figure 4.** Influence of sediment supply rate  $q_s$  (S\_SSP, Table 3) on total erosion with the TLEM and SEM for the same hydrology as the reference simulation (S\_REF, Table 3). The legend applies to all panels. (A) Relative sediment supply ( $q_s/q_{tc}$ ); (B) Total erosion rate computed with the TLEM ( $E_{tot}$ ); (C) Total erosion rate computed with the SEM ( $E_{salt}$ ). Note the logarithmic scale on the y-axis of (B) and (C).

For sediment supply rates  $q_s \leq q_{s,ref} \times 2$  (Fig. 4a) the patterns of total erosion with the TLEM and SEM (Fig. 4b and 4c) show the same features as in Fig. 3: no erosion in the uppermost part of the profile followed by a local maximum around mid-profile, a local minimum around km 46–47 and a sharp rise in the last 3 km. For larger sediment supply  $q_s \geq q_{s,ref} \times 5$ , the patterns change and have a single maximum only. If the relative sediment supply satisfies  $q_s/q_{tc} > 0.3$  (Fig. 4a) the number of tools remains high enough for total erosion to increase with transport stage ( $\tau^*/\tau_c^*$ ) as the increase in channel size compensates for the small drop in erosion per unit width  $\dot{e}_{tot}$  when  $q_s/q_{tc} < 0.5$ .

Erosion occurs over 48 km of the bed (S\_SSP,  $q_{s,ref}/20$ ) in both the SEM and TLEM for the lowest sediment supply, whereas the bed is almost completely shielded for the second largest sediment supply (S\_SSP,  $q_{s,ref} \times 20$ ) and erosion occurs only between km 46 and 48 in the TLEM (Fig. 4). The SEM treats the whole sediment supply as participating to the cover effect, and the bed is shielded as soon as  $q_s/q_{tc} \leq 1$ . The TLEM discriminates between transport as bedload and in suspension, therefore if  $q_b < q_s$  the bed can remain partially exposed even for a relative sediment supply  $q_s/q_{tc} \geq 1$ . The erosion window is therefore larger for the TLEM (Fig. 6 in Lamb et al., 2008). The fact that

erosion rates computed with the TLEM are higher than those computed with the SEM is inherent to the model formulation (Lamb et al., 2008).

An interesting conclusion arising from Fig. 4 is that total erosion is significant over most of the bed (Fig. 4b and 4c) when the sediment supply rate is relatively low ( $q_s \leq q_{s,\text{ref}} \times 2$ ; Fig. 4a). Total erosion becomes more localized at higher relative sediment supply rates. Changing the particle diameter ( $D$ ), instead of sediment supply ( $q_s$ ), leads to similar changes in relative sediment supply and therefore in erosion patterns (see Supplement). The transport stages calculated are large enough to produce significant differences between the SEM and TLEM for large relative sediment supplies. Hereafter, we focus on the TLEM only in the results as it is more appropriate for the flow conditions encountered beneath glaciers.

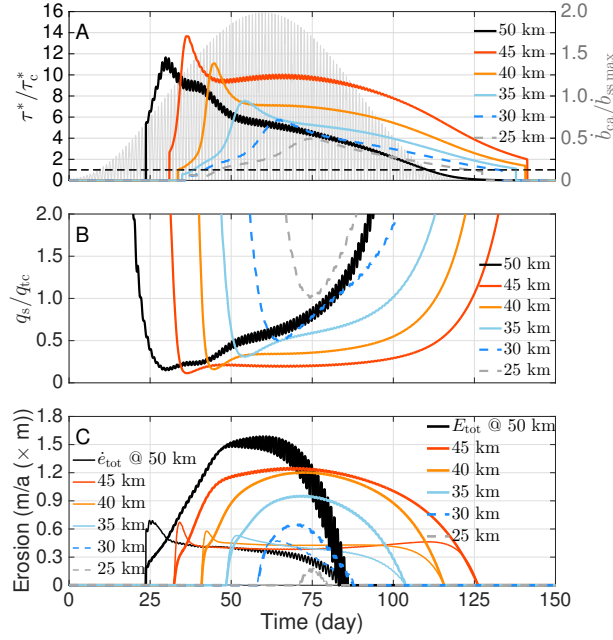
## 4.2 Transient coupled simulations

Arguably, most erosion and sediment transport in fluvial systems occur during flood events (e.g., Whipple et al., 2000, 2013; Kirchner et al., 2001; Sklar and Dietrich, 2004; Lamb et al., 2008; Turowski et al., 2009; Lamb and Fongstad, 2010; Cook et al., 2013). In glacial environments large daily variations in melt-water input to the glacier bed can be likened to periodic flooding (e.g., Willis et al., 1996). We perform a series of transient simulations with a cavity network coupled to an R-channel (Eqs. (1), (2), (3) and (4)) to explore how the transience in subglacial water flow is affected by changes in ice thickness, surface melt and sediment supply, and how this transience impacts instantaneous and annually integrated erosion rates. In this series of simulations, we choose the sediment supply ( $q_{s,\text{ref}} = 9.1 \times 10^{-3} \text{m}^2 \text{s}^{-1}$ ) such that the modelled sediment yield ( $\int \min(q_s(X_L)W_{\text{ch}}(X_L), Q_{\text{tc}}(X_L))dt$ ) corresponds to an inferred basin-wide erosion rate of a few millimetres of erosion per year (e.g., Gurnell et al., 1996; Hallet et al., 1996; Koppes and Montgomery, 2009).

### 4.2.1 Reference model

The reference model uses a synthetic forcing in the form of water supply to the distributed system ( $\dot{b}_{\text{ca}}$ , Eq. (3); sinusoid with a period of 120 days on which we superimpose daily fluctuations) and the reference glacier geometry (Fig. 1, REF; Table 4, T\_REF). The sediment supply rate per unit width is assumed to be uniform along the bed. With this simple test we explore how the transience in water input and response of the subglacial drainage system affect transport stage and erosion rate. Modelling subglacial water flow through coupled distributed and channelized drainage systems has been described extensively in recent literature (see Table 3 in Flowers, 2015), so we omit a discussion of the drainage system itself and focus instead on how subglacial hydrology affects transport stage and erosion.

Since the sediment supply is fixed, transport stage (Fig. 5a) and relative sediment supply (Fig. 5b) are anti-correlated. The time transgression in transport stage and relative sediment supply is a result of the upglacier incision of R-channels. Once a channel is well developed, water pressure decreases



**Figure 5.** Timeseries at six different distances from the divide (km 50, 45, 40, 35, 30 and 25) for simulation T\_REF (Table 4). (A) Transport stage ( $\tau^*/\tau_c^*$ ) and normalized meltwater input ( $b_{ca}/b_{ss\ max}$ ) at the terminus (light grey). The dashed line represents the threshold for sediment motion:  $\tau^*/\tau_c^* = 1$ . (B) Relative sediment supply  $q_s/q_{tc}$ . (C) Erosion per unit width  $\dot{e}_{tot}$  ( $m\ a^{-1}$ , thin lines) and total erosion  $E_{tot}$  ( $m^2\ a^{-1}$ , thick lines) computed with the TLEM.

in the channel, as does water velocity and transport stage (Fig. 5a). Daily fluctuations are only detectable close to the terminus, where a channel is relatively well developed. Similar to what we find in the steady state simulations, the largest transport stages are found a few kilometres up-glacier from the terminus. The bed remains shielded over the first 25 km from the ice divide. The time window during which erosion occurs ( $q_s/q_{tc} \lesssim 1 - 1.5$ ) is longest at km 45 and decreases up- and down-glacier. The size of the R-channel at the terminus (km 50) after day 80 becomes large enough that the transport stage (Fig. 5a) is insufficient to maintain an exposed bed. Further upstream from km 45, the hydraulic potential gradient shallows and the discharge decreases so that the bed is exposed for a shorter time.

Erosion per unit width ( $\dot{e}_{tot}$ , Fig. 5c) peaks when the relative sediment supply satisfies  $0.25 \leq q_s/q_{tc} \leq 0.4$ , i.e. at the onset of R-channel formation, even before the peak in transport stage for the lowermost 15 km (km 35–50 in Fig. 5). The peak in erosion is followed by relatively constant values and eventually an abrupt drop. At the terminus, erosion ceases due to low shear stress in the relatively large channel, while upstream erosion ceases due to declining water supply. Note that erosion can

occur after the melt season ends (day 120) at km 45 because water remains stored englacially and subglacially (Eq. (3)). Further up-glacier (km 25–30) the maximum erosion per unit width coincides with the minimum in relative sediment supply, as the latter remains larger than 0.5 and enough tools are available close to the bed.

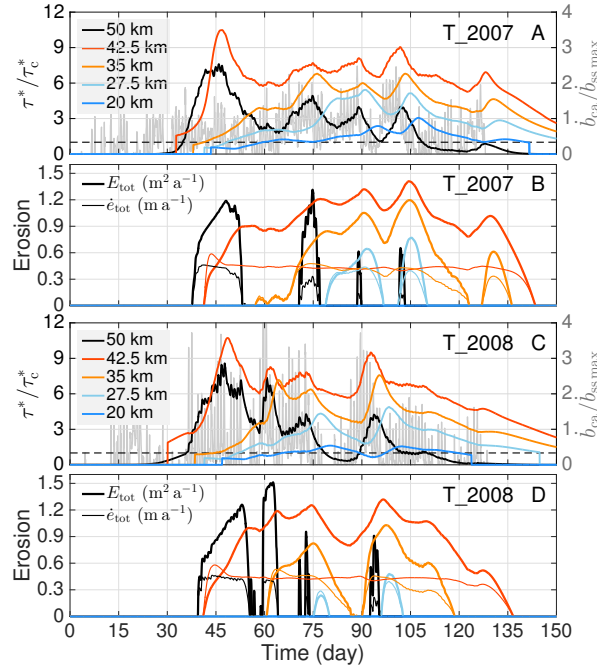
480 Patterns of total erosion ( $E_{\text{tot}}$ , Fig. 5c, thick lines) all peak between day 65 and 75, except at the terminus (km 50). The initial peak in erosion per unit width occurs while the channel is small, thus total erosion is largely controlled by channel size over most of the record. Simulations with transient meltwater input highlight the role of channel size in controlling transport stages and erosion close to the terminus.

#### 485 4.2.2 Surface melt

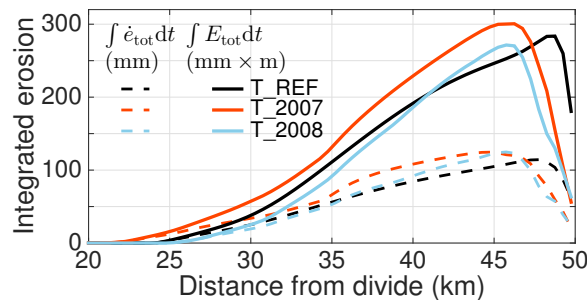
We vary the amount of water reaching the bed to explore the differences between using synthetic and realistic melt records. The realistic melt records come from the ablation area of an unnamed glacier in the Saint Elias Mountains, Yukon, Canada, in 2007 and 2008 (Wheler et al., 2014). We scale the 2007 melt record (T\_2007, Table 4) so that the total volume of water is identical to the synthetic  
490 input. The melt timeseries from 2008 (T\_2008, Table 4) is then scaled such that the ratio of 2007 to 2008 melt volumes is preserved. This test is intended to highlight the importance of total melt volume and the temporal structure of meltwater input.

When we apply the realistic forcing from 2007 (T\_2007), the transport stage exhibits four to five peaks (Fig. 6a) at km 35, 42.5 and 50. At these three locations, the first peak occurs once enough  
495 water is supplied to the bed to form a channel (Fig. 6a; after 45 days at 42.5 km and 50 km, and after 75 days at 35 km). The subsequent peaks in transport stage (after day 60 at 42.5 km and 50 km and after day 80 at 35 km) follow periods of high melt. After day 45, the transport stage remains highest at 42.5 km because creep closure prevents R-channels from becoming too large and water discharge is high enough to maintain high velocities. At km 20 and 27.5 it takes about 60 days for the first  
500 peak in transport stage to occur. The subsequent peaks at these two locations (around day 90, 105 and 129) lag high melt periods even further.

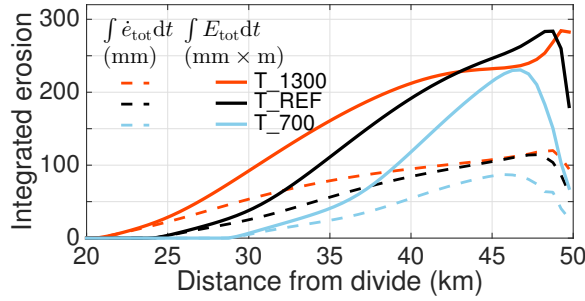
Given the prescribed sediment supply rate, the bed is only exposed at transport stages larger than 3.5. Erosion (Fig. 6b;  $\dot{e}_{\text{tot}}$  and  $E_{\text{tot}}$ ) at km 27.5 and 50 therefore only occurs during peaks in transport stage; at km 20 the bed is always covered. Erosion rate per unit width ( $\dot{e}_{\text{tot}}$ , Fig. 6b) plateaus at  
505 moderate transport stages, thus it remains relatively constant once the bed is partially exposed. On the other hand, total erosion ( $E_{\text{tot}}$ , Fig. 6b) peaks with transport stage (Fig. 6a). Similar results are obtained for T\_2008 (Fig. 6c and 6d), where a different melt timeseries is employed. While the amplitudes of the fluctuations of meltwater input are a few times larger in T\_2008 than T\_2007 (Fig. 6a and 6c), the total melt in T\_2008 is about 80% that of T\_2007. These realistic meltwater forcings  
510 produce episodic variations in transport stage and erosion rate that suggest multi-day fluctuations in meltwater input are important.



**Figure 6.** Timeseries of transport stage and erosion rates for two realistic water input timeseries (Table 4, T\_2007 and T\_2008) at five different distances from the divide (km 20, 27.5, 35, 42.5 and 50). (A) Transport stage ( $\tau^*/\tau_c^*$ ) and normalized meltwater input ( $\dot{b}_{ca}/\dot{b}_{ss\max}$ ) at the terminus (light grey) for simulation T\_2007. (B) Erosion rate per unit width  $\dot{e}_{tot}$  (thin lines) and total erosion  $E_{tot}$  (thick lines) for simulation T\_2007. (C) Transport stage ( $\tau^*/\tau_c^*$ ) and normalized meltwater input ( $\dot{b}_{ca}/\dot{b}_{ss\max}$ ) at the terminus (light grey) for simulation T\_2008. (D) Erosion rate per unit width  $\dot{e}_{tot}$  (thin lines) and total erosion  $E_{tot}$  (thick lines) for simulation T\_2008. The dashed lines in (A) and (C) represent the threshold for sediment motion:  $\tau^*/\tau_c^* = 1$ .



**Figure 7.** Comparison of time integrated erosion per unit width ( $\int \dot{e}_{tot} dt$ ) and total erosion ( $\int E_{tot} dt$ ) for different water inputs. The first 20 km of the profile are not shown because the bed is alluviated and erosion rates are negligible.



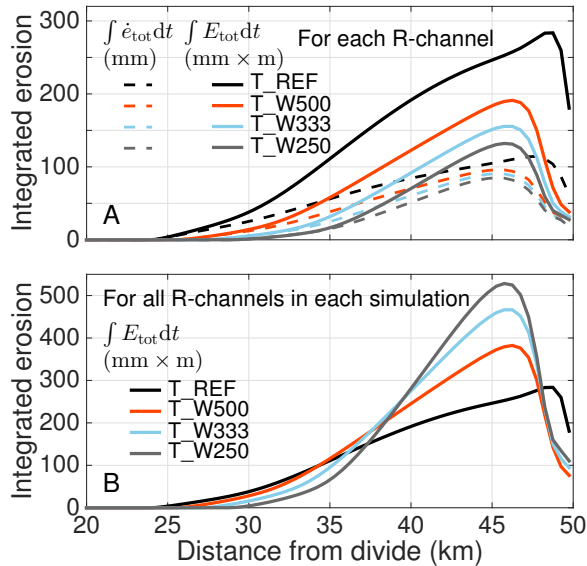
**Figure 8.** Comparison of time integrated erosion per unit width ( $\int \dot{e}_{\text{tot}} dt$ ) and total erosion ( $\int E_{\text{tot}} dt$ ) for different ice geometries. The first 20 km of the profile are not shown because the bed is alluviated and erosion rates are negligible.

These multi-day variations in water input also lead to a succession of channel enlargement events (Fig. 6a and 6c), represented by multiple peaks in transport stage ( $\tau^*/\tau_c^*$ ). On the time scale of several days, creep closure near the terminus is low enough that a channel is sustained between the melt events, leading to an up-glacier migration of relatively large transport stages and integrated erosion. Thus, the pressure in the channel close to the terminus, and hence transport stage, is low. This results in the integrated total erosion ( $\int E_{\text{tot}} dt$ ) being about three times lower for realistic inputs (Fig. 7, T\_2007 and T\_2008) than for the synthetic one (Fig. 7, T\_REF) at the terminus. For the same total water input (T\_REF and T\_2007), the realistic melt-season produces more erosion averaged over the glacier bed than the synthetic input ( $8 \times 10^{-2}$  mm for T\_2007 versus  $\sim 7 \times 10^{-2}$  mm for T\_REF).

### 4.2.3 Ice geometry

Studies of sediment yield from glacierized catchments (e.g., Hallet et al., 1996; Koppes and Hallet, 2002, 2006; Koppes and Montgomery, 2009) conclude that glaciers are more erosive during retreat than during advance due to the amount of meltwater production. Glacier thinning (or thickening) during a phase of retreat (or advance) will also impact the development of the subglacial drainage system and hence its ability to flush sediments and erode the bed. In this series of model tests we hold the sediment supply fixed and vary the glacier geometry by changing the maximum ice thickness (Fig. 1, T\_REF, T\_1300, T\_700, Table 4), while the water input remains the same. As we have already described the principal mechanisms responsible for fluctuations in erosion rates in previous sections, we now focus on annually integrated erosion.

For all ice geometries tested in Fig. 8 (T\_1300, T\_700 and T\_REF, Table 4), significant erosion only occurs downglacier of km 20. The thicker the ice, the further upglacier significant erosion (both  $\int \dot{e}_{\text{tot}} dt$  and  $\int E_{\text{tot}} dt$ ) occurs (up to km 21 for T\_1300 and km 29 for T\_700). In these tests, thicker ice also means steeper surface slopes (Fig. 1). Since water input is identical for these simulations,



**Figure 9.** Influence of drainage catchment width ( $W$ ) on time integrated erosion. (A) Time integrated erosion per unit width ( $\int \dot{e}_{tot} dt$ ) and total erosion ( $\int E_{tot} dt$ ) for an individual R-channel. (B) Time integrated total erosion ( $\int E_{tot} dt$ ) summed over all R-channels in each simulation. The first 20 km of the profile are not shown because the bed is alluviated and erosion rates are negligible.

steeper surface slopes lead to faster water flow and the possibility of initiating sediment motion further up glacier. At the terminus almost four times as much erosion occurs for T\_1300 than T\_700 because thick ice prevents the growth of a large channel.

#### 4.2.4 Subglacial drainage catchment width of a channel

540 Hydraulic properties of the distributed drainage system determine the density of channels that form (e.g., Werder et al., 2013). The smaller the channel spacing, the lower the discharge in a single channel. A smaller channel, at equilibrium, yields larger water pressures, so we expect that more water would be evacuated through the cavity network. In this test we fix the total glacier width at 1000 m and allow two, three or four channels to form such that channel catchment widths ( $W$ ) are, 545 respectively, 500, 333 and 250 m (T\_W500, T\_W333, T\_W250; Table 4).

If we consider the erosion in a single R-channel per simulation (Fig. 9a), the smaller the drainage catchment width, the smaller the discharge, and the smaller the time integrated erosion (Fig. 9a;  $\int \dot{e}_{tot} dt$  and  $\int E_{tot} dt$ ). The feedback causing erosion rate per unit width to decrease at large transport stages (cf. Fig. 2) is such that even for the simulations where the drainage catchment width is 550 relatively small, erosion rates are comparable ( $\int \dot{e}_{tot} dt$ , Fig. 9a) despite the lower transport stages. The differences are, however, relatively large for the annually integrated total erosion (maximum

$\int E_{\text{tot}} dt$  for T\_REF is more than twice that of T\_W250, Fig. 9a) because of the effect of channel size.

The hierarchy in total integrated erosion is inverted when all R-channels within a fixed glacier width are accounted for (Fig. 9b, between km  $\sim 37$  and  $\sim 48$ ). Once the number of R-channels present is taken into account, integrated total erosion is largest for the smallest channel catchment (almost twice as large for T\_W250 than T\_REF, Fig. 9B). In this case, numerous small channels therefore produce more erosion than few large ones.

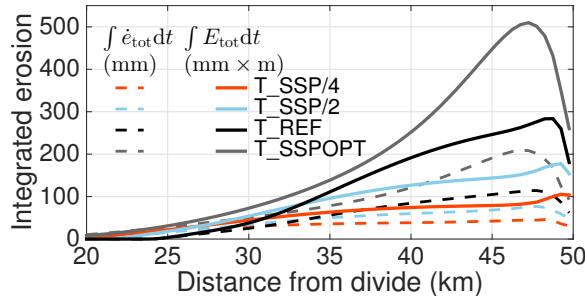
At the terminus (km 50), the simulations with a catchment width per channel smaller than that in T\_REF show values of annually integrated erosion ( $\int E_{\text{tot}} dt$ ) of about half that of the reference simulation (Fig. 9b). The relatively smaller R-channels in these simulations remain more pressurized and thus drain less water from the cavity network (Eq. (1)); the relative discharge in the cavity network near the terminus is therefore larger than in the reference simulation (T\_REF), further diminishing transport stage near the terminus (cf. Figs. 2–3).

#### 4.2.5 Sediment supply

In the present model the values and patterns of sediment supply are amongst the key unknowns. Most till is produced subglacially (e.g., Sanders et al., 2013) and the amount and size distribution of till depends on the history and patterns of production (quarrying) and comminution (abrasion). We test the sensitivity of erosion rates and patterns to different values of input sediment supply. In two simulations (T\_SSP/2 and T\_SSP/4; Table 4) the sediment supply rate per unit width is constant in space and time and is taken as a fraction of the reference supply rate in T\_REF (Table 4). The largest erosion rate in the SEM occurs when the relative sediment supply is  $q_s/q_{tc} = 0.5$ . For the TLEM and transport stages  $\tau^*/\tau_c^* < 100$ , maximum erosion rate is obtained for a relative sediment supply of  $0.5 \leq q_s/q_{tc} < 0.8$  (Lamb et al., 2008). We determine a ratio  $q_s/q_{tc}$  close to optimum and examine the resulting erosion rates and patterns (T\_SSPOPT; Table 4). This provides us an upper bound on subglacial meltwater erosion rates. The hydraulic conditions in this suite of simulations are that of T\_REF (Fig. 5).

Decreasing the sediment supply leads to a decrease in the maximum integrated erosion (Fig. 10, T\_SSP/2 and T\_SSP/4;  $\int \dot{e}_{\text{tot}} dt$  and  $\int E_{\text{tot}} dt$ ) and to the bed being eroded further up-glacier. For a relatively low sediment supply (T\_SSP/4), the peak in annually integrated erosion per unit width ( $\int \dot{e}_{\text{tot}} dt$ , Fig. 10) is hardly discernible and the peak in annually integrated total erosion ( $\int E_{\text{tot}} dt$ , Fig. 10) is controlled by channel size. In order to estimate the maximum erosion that can occur under the given hydraulic conditions and sediment size, we optimize the sediment supply rate by expressing it as a function of the transport capacity ( $q_s/q_{tc} \approx 0.6$ ). The resulting patterns of annually integrated erosion ( $\int \dot{e}_{\text{tot}} dt$  and  $\int E_{\text{tot}} dt$ , Fig. 10, T\_SSPOPT) mimic the transport stage patterns (see  $\tau^*/\tau_c^*$ , Fig. 5A), and peak at nearly twice the values of T\_REF.





**Figure 10.** Time integrated erosion per unit width ( $\int \dot{e}_{\text{tot}} dt$ ) and time integrated total erosion ( $\int E_{\text{tot}} dt$ ) as a function of sediment supply rate ( $q_s$ ) (Table 4). The first 20 km of the profile are not shown because the bed is alluviated and erosion rates are negligible.

## 5 Discussion

### 5.1 Significance of model simplifications

We have detailed the simplifications and underlying assumptions of the model while describing the model and the strategy, we therefore focus on the potential implications of the most important simplifications. At the onset of the melt season, sliding is expected to accelerate as a response to increased water supply to a distributed drainage system (e.g., Iken, 1981; Hooke et al., 1989; Mair et al., 2003; Anderson et al., 2004; Sole et al., 2011; Meierbachtol et al., 2013; Hewitt, 2013; Hoffman and Price, 2014) which would promote cavity enlargement and water flow through the distributed rather than the incipient channelized drainage system. This sliding feedback alone could produce a small decrease in water pressure (e.g., Hoffman and Price, 2014), hence a decrease in transport stage.

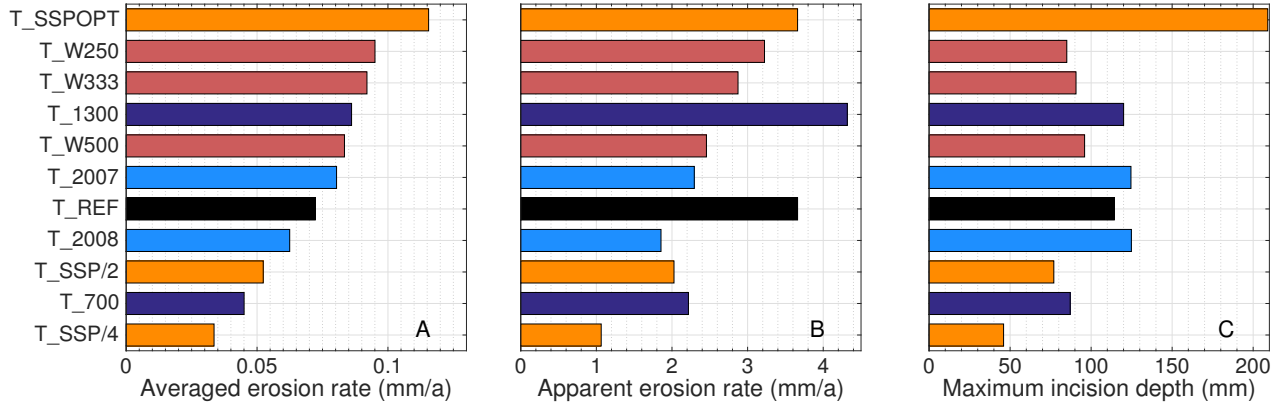
In this study, we treat only the case of bedrock erosion by abrasion and we neglect the effect of quarrying. Although the latter can lead to erosion rates up to an order of magnitude larger than abrasion, it requires that the bedrock be highly jointed (Whipple et al., 2000, 2013). Quarrying is a two-step process: (1) loosening of blocks around pre-existing cracks (or possibly opening of new cracks) and (2) mobilization and transport of loose blocks (Whipple et al., 2000, 2013; Chatanantavet and Parker, 2009; Dubinski and Wohl, 2013; Lamb et al., 2015). The depth of loose cracks could be related to sediment availability (Chatanantavet and Parker, 2009) and mobilization and transport of quarried blocks scale with the transport stage (Dubinski and Wohl, 2013; Lamb et al., 2015). Therefore we expect that the patterns of quarrying would be similar to the transport stage, yet limited by the thickness of the loosened layer.

We compute erosion with only a single particle size that is assumed to be the median of size of the sediment mixture (e.g. Sklar and Dietrich, 2004, 2006, 2008; Turowski et al., 2007; Lamb et al., 2008; Chatanantavet and Parker, 2009; Nelson and Seminara, 2011). The SEM (Sklar and Dietrich, 2004) was generalized for a grain size distribution by Egholm et al. (2013), a study in which they

however omit a discussion of the implications of the generalization of the SEM. As for the TLEM, Lamb et al. (2008) suggest that a generalization to grain size distribution would require re-evaluation of some of the equations to account for the interactions between particles of different sizes within the bedload layer. A decrease in median sediment size would probably result in an erosion profile  
615 more spread out along the bed and an increase in median sediment size would result in a localization of erosion (see Supplement). The changes in erosion would be quantitatively similar to a decrease in sediment supply ( $q_s$ ) and thus a decrease in relative sediment supply ( $q_s/q_{tc}$ ), which strongly controls erosion patterns.

We make the assumption of a supply-limited glacier bed, hence neglect the effect of sediment  
620 transport and the interactions between sediment thickness and water flow. The mobilization and particularly deposition of sediment affect the flow regime by enlarging or reducing the cross-section of flow (Creys et al., 2013). On a timescale of days, when sediment is mobilized, the cross-section of flow is enlarged and could result in a drop in channel water pressure and a corresponding loss of flow strength. The opposite effect, leading to flow strengthening, could occur when sediment is  
625 deposited. We do not treat the case of transport-limited conditions, where the channelized drainage system would more closely resemble canals (Walder and Fowler, 1994; Ng, 2000; Kyrke-Smith and Fowler, 2014). To implement sediment transport adequately, it is necessary to improve existing models for subglacial water flow through canals (Walder and Fowler, 1994; Ng, 2000; Kyrke-Smith and Fowler, 2014) with time-evolving effective pressure. Alley et al. (1997), however, argue that in  
630 the case of subglacial water flow through canals, the water pressure remains relatively high and very little water would be drained from the distributed system, limiting the capacity of canals to transport sediment. In contrast, the steady-state water pressure in an R-channel, decreases with increasing discharge, favouring water flow in the channelized system and enhancing transport and erosion.

Accounting for the production of sediment and the evolution of particle diameter at the glacier  
635 bed would also largely influence sediment supply patterns. For example, we can speculate that if the sediment sources are localized in areas of more easily eroded bedrock (e.g., Dühnforth et al., 2010), tools would only be present downstream from these areas. If instead of fixing the sediment supply per unit width, we fix the total sediment supply (simulation not shown), tools are less available at peak flows, reducing erosion, whereas the cover effect is enhanced for a relatively small channel. We also  
640 tested a simple power-law downstream fining function (e.g., Sklar and Dietrich, 2006; simulation not shown). The results were very similar to those obtained with a decrease in relative sediment supply, because particles were smaller than the reference size of 60 mm in the region of the bed where channels form. Another means of obtaining insight into sediment supply rates and patterns would be through the use of a comprehensive model of glacial erosion, i.e. a model encompassing transient  
645 subglacial hydrology (between distributed and channelized systems), ice dynamics, glacial abrasion and quarrying. Such a model is however yet to be developed as patterns of glacial erosion remain poorly understood (cf. Beaud et al., 2014). Finally, subglacial water flow evacuates a significant



**Figure 11.** Synthesis of transient simulations (Table 4) through comparison of the following quantities calculated for one model year: (A) Erosion rate averaged over the whole glacier bed ( $\int \int E_{tot} dt dx$ ); (B) Apparent erosion rate calculated as the volume of sediment that is transported across the last grid node, i.e. terminus, ( $\int \min(q_s W_{ch}, Q_{tc}) dt$ ) averaged over the glacier area. This quantity corresponds to what one would measure as the sediment flux in a proglacial stream. (C) Maximum incision depth ( $\max(\int \dot{e}_{tot} dt)$ ). Simulations are ranked by averaged erosion rate and the colours represent different simulation suites: black for “reference”, blue for “water input”, purple for “ice geometry”, red for “drainage width” and orange for “sediment supply” (Table. 4).

650 volume of sediment despite the small area over which R-channels operate and the tendency of these channels to remain stably positioned in association with moulins (e.g., Gulley et al., 2012). The mechanism by which large volume of sediment are delivered to the channels remains elusive. More work is therefore required to quantify subglacial sediment production patterns.

## 5.2 What are the major controls on subglacial meltwater erosion?

We rank the transient simulations by glacier-area-averaged erosion rate in Fig. 11a. Because we prescribe water input rates sufficient to form a channelized drainage system, it stands out from the model formulation that sediment supply is the most important parameter. A lack or overabundance of tools 655 inhibits erosion. In our results this is shown by the fact that T\_SSPOPT (sediment supply optimized for erosion; Table 4) produces the most erosion and T\_SSP/4 (smallest sediment supply; Table 4) the least. Changing the ice geometry also leads to a relatively large range of averaged erosion rates as T\_1300 (thick ice; Table 4) yields twice as much erosion as T\_700 (thin ice; Table 4; Fig. 11a). 660 Larger hydraulic potential gradients in T\_1300 cause the shear stress be large enough over larger portions of the bed to create erosion (Fig. 9). Subglacial drainage catchment width, within the range tested, plays a lesser role than sediment supply or ice geometry although the averaged erosion rate in T\_W250 (four channels; Table 4) is  $\sim 30\%$  more than that of T\_REF. The fact that T\_2007 (realistic

melt season from 2007 record; Table 4) produces more averaged erosion than T\_REF suggests that  
665 an increase in the multi-day variability of the water input enhances erosion.

The relations are different for apparent erosion rate (Fig. 11b), here defined as the equivalent  
thickness of bed material evacuated by the integrated sediment flux ( $\int \min(q_s W_{ch}, Q_{tc}) dt$ ) at the  
terminus (km 50). The apparent erosion rate corresponds to the quantity estimated by studies of  
sediment yield in proglacial channels, lake or fjords. Relatively large hydraulic potential gradients  
670 and relatively thick ice close to the terminus, both of which inhibit R-channels growth, compete  
against the loss of transport capacity. Therefore the largest apparent erosion occurs for the thickest  
ice (T\_1300, Fig. 3). Interestingly, the lowest drainage density (T\_REF) yields more apparent erosion  
than the highest (T\_W250). Discharge through the cavity network close to the terminus increases  
with R-channel density; the smaller the channel, the larger the water pressure and the lower the  
675 pressure gradient between the two systems. This feedback, in addition to the discharge in the R-  
channel being smaller due to the R-channel drainage catchment size, reduces the transport stage  
close to the terminus.

The results in Fig. 11b suggest that, despite the increase in apparent erosion that accompanies an  
increase in meltwater input (e.g., the total melt in T\_2007 is about 1.25 times that of T\_2008), the  
680 thinning associated with the retreat of an ice mass would have a competing effect by decreasing the  
hydraulic potential gradient (see Fig. 3). The flushing power of subglacial water flow is conducive  
to the removal of subglacial sediment enabling glacial abrasion and quarrying to be efficient. Our  
results hint that the subglacial drainage conditions most favourable for glacial erosion occur where  
significant surface melt and relatively steep surface slopes occur simultaneously, i.e. during an ice  
685 sheet maximum advance or during early phases of retreat. This corroborates the hypothesis of Jør-  
gensen and Sandersen (2006) that some Danish tunnel valleys were excavated during the stagnation  
of the Scandinavian ice sheet. However, these findings challenge the hypothesis that glaciers de-  
liver more sediment to proglacial areas during retreat than during advance (e.g., Hallet et al., 1996;  
Koppes and Hallet, 2002, 2006; Koppes and Montgomery, 2009), yet more work is required to ex-  
690 plore this hypothesis. The lack of flow strength in the upper reaches of the glacier (upstream from km  
20 for most simulations) suggests that subglacial sediment in the accumulation area is transported  
almost solely by entrainment due to sliding at the ice-bed interface.

In the steady state simulations we find that significant erosion can occur in a network of cavities  
(see Supplement). In the transient simulations, however, the coupling with R-channels prevents large  
695 shear stresses from developing in the distributed drainage system, and the threshold for sediment  
motion is not even reached for particles of 1 mm diameter. We thus argue that bedrock erosion  
in the distributed drainage system is limited unless specific conditions are satisfied, for example a  
subglacial flood or a surge.

### 5.3 How important is subglacial meltwater erosion compared to overall glacial erosion?

700 In most literature on modelling landscape evolution by glacial erosion it is assumed that subglacial meltwater efficiently removes sediment from the glacier bed, while its effect on bedrock erosion is neglected (e.g., MacGregor et al., 2000, 2009; Brocklehurst and Whipple, 2002; Tomkin and Braun, 2002; Anderson et al., 2006; Herman and Braun, 2008; Egholm et al., 2009, 2011a, b; Herman et al., 2011). On the other hand, in formerly glaciated landscapes, erosional features like tunnel valleys (e.g., Glasser and Bennett, 2004; Denton and Sugden, 2005; Dürst Stucki et al., 2010, 2012; Kehew et al., 2012; Jansen et al., 2014) indicate that subglacial water can produce significant bedrock erosion. The results we obtain with our simple ice geometries and water input forcings indicate that the areally averaged bedrock erosion produced by subglacial water flow is on the order of  $10^{-1}$  –  $10^{-2}$   $\text{mm a}^{-1}$  (Fig. 11) while glacial erosion rates are most often on the order of  $1$  –  $10$   $\text{mm a}^{-1}$  (e.g., 710 Gurnell et al., 1996; Hallet et al., 1996; Koppes and Montgomery, 2009; Riihimäki et al., 2005). Bedrock erosion by abrasion from sediment-bearing subglacial water appears negligible compared to reported erosion rates in proglacial areas. Our results corroborate the assumption that subglacial meltwater efficiently removes sediment from the bed and we postulate that this flushing action is necessary for glacial abrasion and quarrying to access an exposed bed and remain efficient.

### 715 5.4 Can ordinary seasonal melt processes lead to subglacial bedrock channel incision?

We find maximum modelled vertical bedrock incision ranging from  $\sim 50$  to  $\sim 200$   $\text{mm a}^{-1}$  (Fig. 11c). Assuming that over a period of 20 years climate is relatively steady and the bedrock does not change significantly, the location of moulins would remain relatively fixed laterally and so would the channel paths (Gulley et al., 2012). Using the lowest incision rate (T\_SSP/4), a Nye channel almost a 720 meter deep and a few meters wide could be carved near an ice sheet margin in 20 years. A similar N-channel would be carved in only five years assuming the largest incision rate (T\_SSPOPT).

Landforms created by former continental ice sheets indicate that subglacial water ways can occupy persistent paths throughout a deglaciation. Eskers deposited by the retreating Laurentide ice sheet can be traced for up to several hundred kilometres and show a dendritic pattern almost as far 725 upstream as the former divide (e.g., Storrar et al., 2014). Some tunnel valleys show several cut-and-fill structures suggesting different excavation events; moreover, tunnel valleys carved during different glaciations tend to follow the same paths (e.g., Jørgensen and Sandersen, 2006). Eskers also commonly lie inside tunnel valleys (e.g., Jørgensen and Sandersen, 2006; Kehew et al., 2012). Assuming an incision rate of  $100$   $\text{mm a}^{-1}$  (e.g., Fig. 7), a simple volume calculation suggests that it 730 would take about 15000 years to carve a 30 m deep and 100 m wide V-shaped tunnel valley, similar to the dimensions of tunnel valleys observed in Ireland (Knight, 2003).

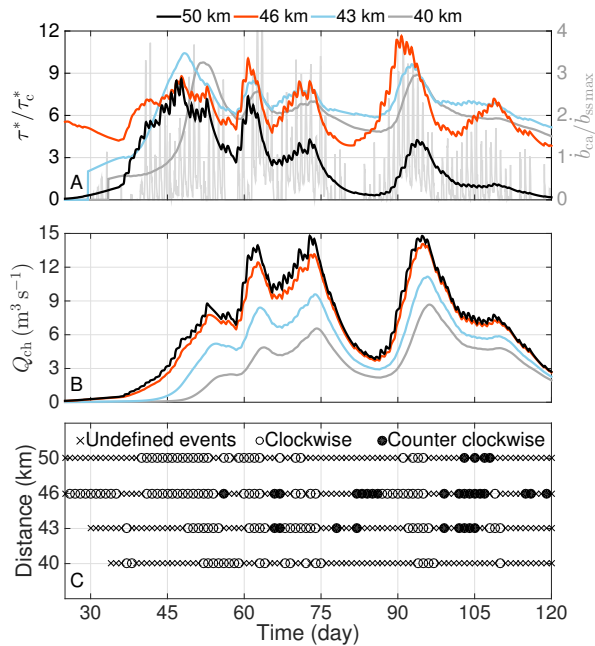
In the context of an alpine glacier, valley geometry tends to focus subglacial water flow paths toward the thalweg. Assuming that a glacier occupies topography strongly imprinted by fluvial pro-

cesses, erosion by subglacial water flow may tend to preserve if not enhance the pre-existing fluvial features along the valley centreline. For an alpine glacier eroding its bed at a pace of  $2 \text{ mm a}^{-1}$  (e.g., Hallet et al., 1996; Riihimaki et al., 2005; Koppes and Montgomery, 2009), in the case of simulation T\_2008 (Table 3, apparent erosion of  $\sim 2 \text{ mm/a}$ ; ice geometry comparable to that of a large valley glacier), the maximum incision depth in one year is  $\sim 125 \text{ mm}$  (Fig. 11c). The relief of a canyon with the maximum width of the N-channel ( $\sim 4.5 \text{ m}$ , for T\_2008) would increase by more than  $\sim 120 \text{ mm a}^{-1}$  (rate of vertical bedrock incision minus rate of surrounding glacial erosion). If the canyon were five times as wide ( $\sim 22.5 \text{ m}$ ), the maximum rate of relief increase would still be  $\sim 24 \text{ mm/a}$ , about twice the measured incision rates in a metres-wide gorge in the French Western Alps (e.g., Valla et al., 2010), highlighting the erosional power of localized subglacial meltwater action.

## 5.5 What are the implications of the water flow regime in channels for hysteresis and sediment transport?

We calculate the direction of daily hysteresis between modelled transport stage (Fig. 12a) and water discharge (Fig. 12b) at four locations within the last 10 km of the glacier profile (Fig. 12c) for simulation T\_2008 (Table 4; Fig 6). Overall, hysteresis is dominated by clockwise events, with counter clockwise events only occurring during the second half of the melt season. Clockwise hysteresis correlates well with the rising limb of multi-day water discharge and transport stage peaks, while counter clockwise hysteresis correlates with the falling limb, particularly at km 46 (Fig. 12). During the rising limb of a multi-day melt event, changes in channel size are dominated by enlargement; the pressure in the channel therefore peaks before the discharge, as do the averaged water flow velocity and transport stage. During the falling limb of the melt event, if the channelized drainage is relatively well established, changes in channel size are dominated by closure, and the pressure peak can occur after the peak in discharge. In the case of a proglacial stream carrying a sediment load smaller than its transport capacity, peaks in transport stage would act as mobilizing events propagating sediment pulses downstream. We therefore surmise that the direction of hysteresis in sediment transport and discharge is not necessarily linked to changes in sediment supply conditions or the tapping of new sediment sources, but may be the result of changes in subglacial sediment mobilization in the vicinity of the glacier terminus.

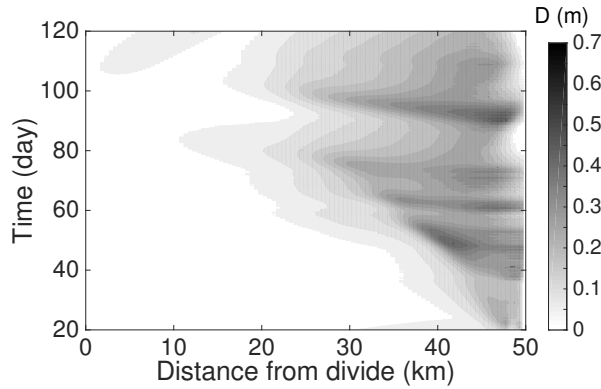
In coarse-bedded streams, grain hiding has a significant effect on sediment transport (e.g., Yager et al., 2012; Scheingross et al., 2013), as mobile grains can be trapped behind larger immobile particles. We calculate the maximum particle diameter for which movement would be initiated in simulation T\_2008 (Fig. 13) and find that boulders of up to 70 cm in diameter can be transported within the last 10 km of the profile and would correspond to flood-like conditions in rivers. The analogy to river systems might have influenced the interpretation of glacial deposits such as eskers,



**Figure 12.** Hysteresis between water discharge and transport stage for simulation T\_2008 (see Table 4 and Fig. 6). Timeseries at four locations (km 40, 43, 46 and 50) distributed over the last 10 km of the glacier profile of: (A) Transport stage ( $\tau^*/\tau_c$ ) and normalized water input ( $\dot{b}_{ca}/\dot{b}_{ss\ max}$ ) at the terminus (light gray); (B) Discharge in the channel,  $Q_{ch}$ ; (C) Calculated direction of the daily hysteresis when transport stage is plotted against water discharge. The hysteresis is clockwise when transport stage peaks before discharge over a daily cycle. Undefined events represent days where the fluctuations in transport stage or discharge are either simultaneous or not strong enough to produce hysteresis.

where the presence of boulders or lack of fines is often used to infer emplacement during flood events  
 770 (e.g., Brennand, 1994; Burke et al., 2012).

For simulation T\_2008, we find that transport stage exhibits a sharp decrease close to the terminus  
 (Figs. 5A, 6A and 6C) which leads to a correspondingly sharp decrease in the size of particles trans-  
 ported (Fig. 13) and could lead to a bottleneck in sediment transport. This bottleneck effect could  
 lead to the deposition of sediment, filling the channel toward the end of the melt season. A similar  
 775 process, but operational over a longer time scale, would be consistent with time-transgressive depo-  
 sition of eskers near the mouths of R-channels beneath retreating ice margins (e.g., Brennand, 1994;  
 Burke et al., 2012).



**Figure 13.** Maximum particle diameter for which movement would be initiated in simulation T\_2008 assuming  $\tau_c^* = 0.03$  (see Table 4 and Fig. 6).

## 6 Conclusions

This study is the first attempt to quantify bedrock erosion rates by transient subglacial water flow with a numerical model. We implement a 1-D model of subglacial drainage in which a network of cavities and R-channels interact. We compute the shear stress exerted on the bed and the resulting bedrock erosion by abrasion (saltation erosion, after Sklar and Dietrich (2004), and total load erosion after Lamb et al. (2008)). Because of the large calculated transport stage we argue that in the case of subglacial meltwater erosion it is probably more appropriate to use the TLEM than the SEM.

Assuming that a significant amount of meltwater is produced and reaches the bed, the main drivers of subglacial water erosion that we isolate are: the rate of sediment supply, particularly the relative sediment supply, and ice geometry.

From this exercise, we conclude that:

1. Bedrock erosion and transport stage in the subglacial drainage system do not scale directly with water discharge. Instead, transport stage and erosion are related to the hydraulic potential gradient and hence a combination of water discharge, ice-surface slope and channel (or cavity) cross-sectional area. In our simulations, this combination of discharge, slope and channel cross-sectional area leads to a drop in transport stage close to the terminus as water pressure approaches atmospheric.
2. Erosion rates due to the action of subglacial water flow averaged over the whole glacier bed are negligible compared to the rates of glacial erosion necessary to produce the sediment supply rates we impose.
3. In our transient simulations, a bedrock channel a few to several decimetres in depth could be carved over a single melt season as erosion is concentrated at the base of R-channels.



800 4. The vertical incision rates we calculate are a few to several times larger than published rates  
of fluvial incision in gorges. Therefore, this mechanism may explain the gradual excavation of  
tunnel valleys in bedrock and the preservation or even initiation of inner gorges.

Though we have demonstrated the potential for subglacial water flow to incise bedrock on seasonal  
timescales, site-specific and quantitative assessments of its importance will require more realistic 2-  
805 D hydrology models (e.g., Hewitt, 2013; Werder et al., 2013) and simulations over timescales of  
glacial advance and retreat.

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