

Modelled Solute Acquisition from Suspended Sediment in a Theoretical 4-Cell Subglacial Hydrological System

GILES H. BROWN¹, BRYN HUBBARD,¹ AND ANDREW G. SEAGREN¹

ABSTRACT

Twenty-five laboratory dissolution experiments have been conducted to quantify rates of solute acquisition, measured as Ca^{2+} concentration against time, from glaciogenic sediments suspended in cold, dilute waters. Suspended sediment character was constrained by field-calibrated ranges of both concentration in meltwater (g cm^{-3}) and specific surface area by sediment mass ($\text{cm}^2 \text{g}^{-1}$). This constraint yielded, for the first time in a glacier hydrochemical study, dissolution rate data as a function of the specific sediment surface area by water volume (SSA_v ; $\text{cm}^2 \text{cm}^{-3}$). The resulting experimental data have been used to calibrate a kinetic dissolution model (Brown et al., in press), where the rate of solute acquisition is considered in terms of an initial concentration reflecting rapid ion-exchange reactions (C_0), an ultimate steady-state concentration (C_s), and a rate parameter (k). The results, which allow the dissolution equation parameters C_0 , C_s , and k to be expressed in terms of SSA_v , are used here to drive dynamic predictive model of within-channel solute acquisition in a theoretical, 4-cell subglacial channel. Using realistic values of SSA_v and residence time, this allows the magnitude of WSA to be predicted at any distance from the head of the channelised drainage system to the glacier snout.

Keywords: kinetic dissolution model, within-channel solute acquisition, subglacial hydrology, suspended sediment.

INTRODUCTION

Bulk meltwater quality reflects the nature of the glacial drainage pathways followed by that water (e.g. Collins, 1978; Sharp, 1991; Tranter and Raiswell, 1991; Brown et al., 1996; Tranter et al., 1996). This is particularly true of subglacial drainage, where relatively pure and aggressive meltwaters generated at the ice surface come into contact with large quantities of freshly eroded, finely ground (and therefore geochemically reactive) debris (e.g. Tranter and Raiswell, 1991; Sharp et al., 1995; Brown et al., 1996; Fairchild et al., 1999a; 1999b). Such subglacial drainage is commonly divided into two major components. First, a widespread distributed drainage system is comprised of elements such as film-flow (e.g. Weertman, 1972), permeable subglacial tills (e.g. Boulton and Hindmarsh, 1987; Clarke, 1987) and linked cavities (e.g. Walder, 1986; Kamb, 1987). This system transports meltwaters relatively slowly ($<0.15 \text{ m s}^{-1}$) (Nienow et al., 1996; 1998) in an environment characterized by high water-rock ratios. Second, a system of discrete channels may be incised either upwards into the glacier base (Röthlisberger, 1972; Hooke, 1989) or downwards into the underlying bedrock (Nye, 1973). This system typically transports meltwaters relatively rapidly ($>0.3\text{--}0.5 \text{ m s}^{-1}$; Nienow et al., 1996) in channels that coalesce

¹Centre for Glaciology, Institute of Geography and Earth Sciences, University of Wales, Aberystwyth, Ceredigion, SY23 3DB, Wales, UK

downglacier in an arborescent pattern, commonly emerging at the glacier portal from a small number of large, low-pressure conduits. Turbulent meltwaters can access, mobilise and transport significant quantities of suspended subglacial sediment in these major conduits (Gurnell, 1987; Willis et al., 1996; Hubbard and Nienow, 1997). Particularly high suspended sediment concentrations may be generated when new areas of the bed are tapped by growing subglacial channels, for example during the ‘Spring event’ when the subglacial drainage system is initially opened up (e.g. Nienow et al., 1998), and during summer periods of rapid channel growth as the subglacial conduit system expands headwards following the upglacier retreat of the seasonal snow cover (Richards et al., 1996; Nienow et al., 1998).

It may therefore be possible to infer certain glacier drainage conditions (particularly those present at the glacier bed) from meltwater quality records (Sharp, 1991; Tranter et al., 1993). Conversely, it may also be possible to predict bulk meltwater quality on the basis of known or assumed subglacial conditions. Both of these types of investigation, however, require accurate quantitative models of solute acquisition rates under realistic glacial drainage conditions (e.g. Clarke, 1996). In order to constrain solute acquisition from suspended sediment in subglacial channels (hereafter referred to as within-channel solute acquisition [WSA]) according to standard geochemical theory (e.g. Lerman, 1979), we have recently developed a generic, kinetic model of solute acquisition which incorporates the effects of the solution composition and variability in suspended sediment specific surface area by water volume (SSA_v).

The rate of dissolution, measured as a change of concentration in solution (C) of a particular dissolved species i (dC^i/dt ; $\text{g cm}^{-3} \text{sec}^{-1}$) is considered in terms of a reaction rate constant (k ; s^{-1}) and the concentration deficit of the solution relative to an ultimate (steady-state) concentration (C_s^i ; g cm^{-3}). Thus, following Lerman (1979):

$$\frac{dC^i}{dt} = k^i (C_s^i - C^i) \quad (1)$$

If boundary layer effects and particle-size changes during dissolution are ignored, and assuming k is constant, Equation 1 may be integrated to give the concentration (C) of species i in solution as function of time:

$$C^i = C_s^i + (C_0^i - C_s^i) e^{-kt} \quad (2)$$

Here, C_0 (g cm^{-3}) is the concentration of element i released by rapid, surface-exchange reactions during the early stages (<180 s) of water–rock contact. Thus, $C = C_0$ at $t \approx 0$, and $C \rightarrow C_s$ as $t \rightarrow \infty$. This incorporates two significant developments in relation to the earlier studies of Brown et al. (1994; 1996). First, the physically based influences of C_0 , C_s and k are explicitly quantified, and second, suspended sediment surface area is introduced as a measured variable, allowing its influence to be included explicitly over the reaction equation parameters.

In the current paper we apply this model to a theoretical 4-cell subglacial hydrological system in order to estimate within-channel Ca^{2+} acquisition. Solute acquisition in each cell is derived from fresh sediment mobilised in each cell, and from the dissolution of each parcel of sediment inherited from each up-glacier cell. The model output is calculated as a cumulative concentration curve, which allows Ca^{2+} acquisition to be assessed within any cell of the theoretical subglacial channelised system.

EXPERIMENTAL METHODS

A suite of laboratory dissolution experiments was undertaken to isolate and quantify the individual effects of variations in water–rock contact time and specific surface area per unit volume of solution (SSA_v) ($\text{cm}^2 \text{cm}^{-3}$) on within-channel solute acquisition (WSA). A more detailed description of the experimental and analytical details may be found in Brown et al. (in press), and is summarised as follows. Sediment samples were recovered as gravel-sized clasts from the margins of Haut Glacier d'Arolla, Switzerland, and crushed to produce freshly ground reactive surfaces. Twenty five dissolution experiments, constrained by field-based records from the glacier, were undertaken at SSCs of between $1\text{E-}4$ and $4\text{E-}3 \text{ g cm}^{-3}$ and at specific surface areas per mass of sediment (SSA_m) of between $5.00\text{E}3$ and $1.64\text{E}4 \text{ cm}^2 \text{g}^{-1}$. Equation 2 was fitted by iteration, to a relative tolerance in χ^2 of $1\text{E-}6$ (Microcal Software Inc., 1997), to each of the solute acquisition experiments. This procedure yielded individual values of C_0 , C_s and k for each SSA_v ($\text{cm}^2 \text{cm}^{-3}$), from which C_0 , C_s and k may be estimated via least-squares regression for any value of SSA_v (Table 1). However, k is not dependent on SSA_v ($r^2 = 0.00$) and is therefore considered constant.

The model suggests that the concentration deficit, represented by the difference between the current concentration (C) and a steady-state concentration (C_s), of the meltwater solution exerts a strong control over dissolution rate. This effect would be significant in the subglacial drainage environment, where the concentration deficit may either be enhanced (for example, by the addition of relatively dilute surface meltwaters through subglacial channel confluence or through the delivery of fresh surface meltwaters via crevasses or moulins), or reduced (for example, by the delivery of relatively sediment-rich meltwaters by channel confluence or as migrating channels tap fresh sediment reserves). In such situations, the rate of WSA will reflect the influences of both (i) the current evolution of the solute concentration of any given water parcel towards C_s , and (ii) the episodic and sudden displacement of that dissolution system as meltwaters of a different concentration or suspended sediment are added, in effect displacing both C and C_s . In order to simulate the influence of such effects experimentally, we mimicked the inflow of suspended-sediment-poor surface waters by diluting the experimental system with chilled, air-equilibrated deionised water after 10800 s, effectively halving SSC and C . Figure 1 illustrates both measured and modelled Ca^{2+} concentrations during these dilution experiments, where C_0 , C_s and k were derived from experimental data for both pre- and post-dilution systems.

Table 1. Best-fit least-squares linear (LSL) regression equations for C_0 and C_s expressed as a function of specific surface area by volume (SSA_v), for 25 dissolution experiments conducted by Brown et al. (in press). k is represented by a constant, defined as the intercept of SSA_v , versus k .

LSL regression	r^2
$C_0 = 4.65\text{E-}8 \text{ SSA}_v$	0.88
$C_s = 1.46\text{E-}7 \text{ SSA}_v$	0.77
$k = 3.71\text{E-}4$	0.00

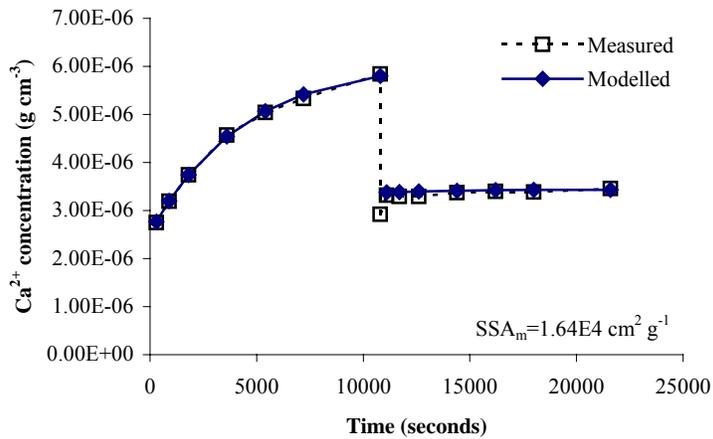
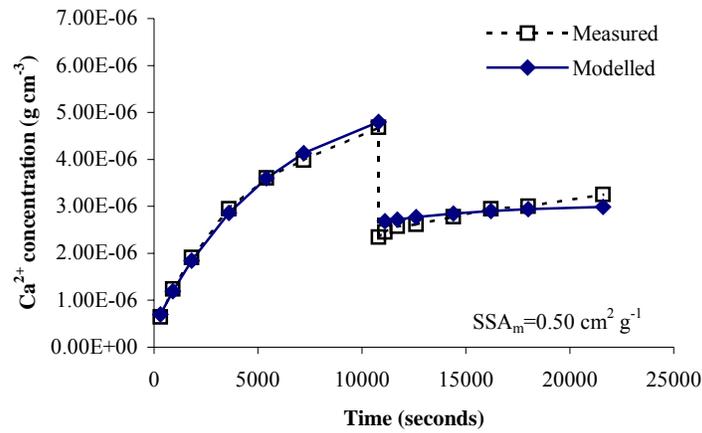


Figure 1. Measured and modelled Ca²⁺ concentrations expressed as a function of time in the dissolution experiments carried out using suspended sediment with a specific surface area by mass (SSA_m) of 0.50 and 1.64E4 cm² g⁻¹. Best-fit model parameters and χ^2 statistics are presented in Table 1.

APPLICATION

In order to illustrate subglacial WSA, a theoretical 4-cell model of the channelised hydrological system is considered. It is assumed that as meltwaters travel from the channel head (Cell 1) to the snout (Cell 4), subglacial discharge increases within each cell, fed by surficial ice melt via moulins and crevasses. In response, fresh sediments are incorporated into the suspended load and these liberate solute as an exponential function (Equation 2). Further, within Cells 2–4 suspended sediment is inherited from up-glacier cells, which continue to release solute. Results indicate that the rate of WSA from these sediments decreases with time from the point of mobilisation, controlled by the competing effects of (a) meltwater dilution reducing SSA_v (and therefore C₀ and C_s) and (b) increasing time.

Therefore, in order to estimate total WSA in a cell, it is necessary to partition total SSC into fresh sediment mobilised within the cell and older sediments inherited from up-glacier cells. This

involves calculating the flux (g s^{-1}) of fresh sediment acquired by a cell and the flux of older sediment from each up-glacier cell. These data are then converted into suspended sediment concentrations within the cell using the subglacial discharge predicted for that cell (Equation 3 and 4).

$$SSC_{(fresh)} = \frac{SSF_{(c_x)} - SSF_{(c_{x-1})}}{Q_{(c_x)}} \quad (3)$$

$$SSC_{(c_{x-y})} = \frac{SSF_{(c_{x-y})}}{Q_{(c_x)}} \quad (4)$$

where c_x = current cell for which total SSC is being estimated, $c_{(x-1)}$ = up-glacier cell adjacent to cell x , $c_{(x-y)}$ = any cell up-glacier of cell x , $SSC_{(fresh)}$ = concentration of fresh sediment mobilised in cell x , $SSF_{(c_x)}$ = flux of sediment through cell x , $SSF_{(c_{x-1})}$ = flux of sediment through the up-glacier cell adjacent to cell x , $SSC_{(c_{x-y})}$ = concentration of pre-weathered sediment from any up-glacier cell present in cell x , $SSF_{(c_{x-y})}$ = flux of sediment through cell $x-y$, and $Q_{(c_x)}$ = discharge of cell x estimated from a glacier melt model (e.g. Richards et al., 1996).

Thus, for a cell fed by three up-glacier cells, it is necessary to calculate the concentration of fresh sediment mobilised by that cell and a concentration of pre-weathered sediment present in the cell mobilised by each of the three up-glacier contributing cells, *ie.* four separate SSCs are calculated. SSA_v may then be derived for each parcel of sediment from the strong linear relationship between SSC and SSA_v ($r^2 = 0.89$ during the 1996 ablation season at Haut Glacier d'Arolla).

In order to calculate WSA within a cell, the model parameters C_0 and C_s need to be estimated for each parcel of sediment present in that cell. This is because each will follow a different dissolution pathway, reflecting changes in SSA_v (and hence C_0 and C_s) with successive dilutions and increasing water-rock contact time. To calculate solute acquisition within a cell from freshly mobilised sediments, the concentrations of fresh sediments ($SSC_{(fresh)}$), and the meltwater residence time (t) for that cell are fed into Equation 3. The same procedure is adopted for calculating solute acquisition from each of the parcels of sediments inherited from each up-glacier contributing cell. SSA_v may then be derived from the new concentration of that sediment parcel, with t equal to the cumulative water-rock contact time for a particular sediment parcel. However, sediments mobilised within up-glacier cells are already pre-weathered. The total WSA therefore reflects within-cell solute acquisition, plus solute inherited from up-glacier cells, divided by the new discharge.

To illustrate this procedure graphically, Figure 2 shows Ca^{2+} acquisition calculated for a theoretical 4-cell channelised hydrological system. Discharge increases from $0.1 \text{ m}^3 \text{ s}^{-1}$ in Cell 1 to $0.37 \text{ m}^3 \text{ s}^{-1}$ in Cell 4. Total SSC is $3.03\text{E-}5 \text{ g cm}^{-3}$ in Cell 4, and is estimated for each up-glacier cell according to the relationships in Equations 3 and 4. Meltwater residence time is displayed cumulatively on the x-axis. Figure 2 indicates that the majority of solute acquired within each cell is derived from the dissolution of sediments mobilised within that cell. Conversely, within each cell, relatively little solute is acquired from older sediments. For example, solute acquired within Cell 4 from sediments inherited from Cell 1 constitutes only 4 % of the total WSA in the cell. This is because this parcel of sediment is reacting at a low concentration (and hence low SSA_v) within Cell 4 and has already undergone more rapid dissolution within Cells 1–3.

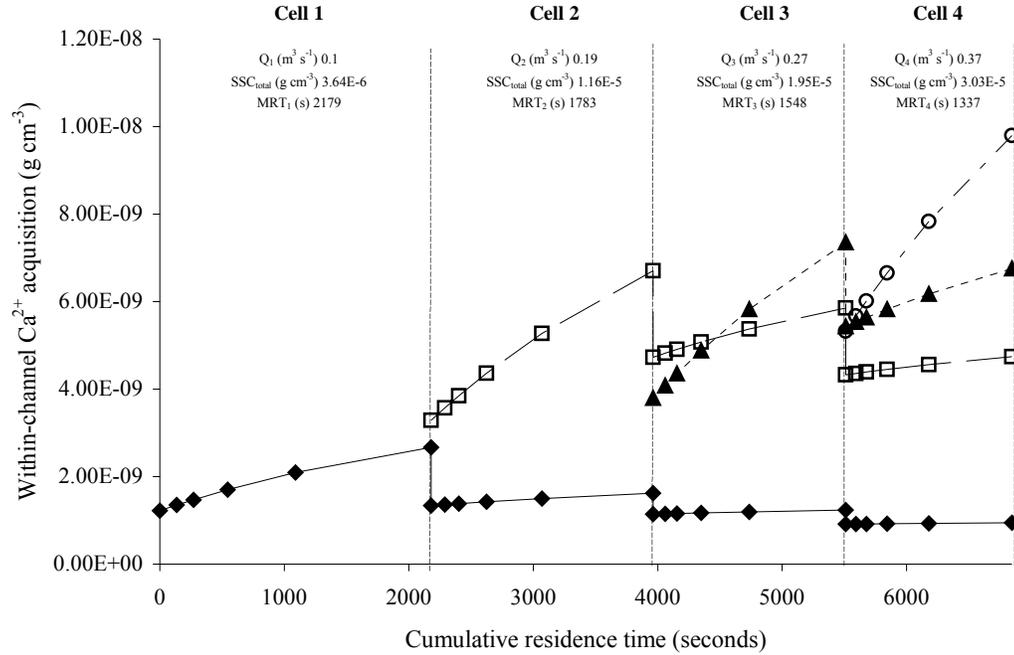


Figure 2. Modelled Ca^{2+} acquisition from a theoretical 4-cell subglacial channelised drainage system. Discharge (Q), total suspended sediment concentration ($\text{SSC}_{\text{total}}$) and meltwater residence time (MRT) are displayed for each cell. Changes in SSC, SSA_v and the dissolution parameters C_o , C_s and k are detailed in Table 2. Closed diamonds (\blacklozenge) indicate sediment mobilised in Cell 1, open squares (\square) sediment mobilized in Cell 2, closed triangles (\blacktriangle) sediment mobilized in Cell 3 and open circles (\circ) sediment mobilized in Cell 4.

Table 2. Variations in suspended sediment concentration (SSC), suspended sediment surface area per unit volume (SSA_v), meltwater residence time (MRT) and the dissolution parameters C_o , C_s and k for individual sediment parcels traveling through the theoretical 4-cell model depicted in Figure 2.

	Cell 1	Cell 2	Cell 3	Cell 4
Q ($\text{m}^3 \text{s}^{-1}$)	0.10	0.19	0.27	0.37
MRT (s)	2179	1783	1548	1337
SSC_1 (g cm^{-3})	3.64E-6	1.82E-6	1.29E-6	9.51E-7
SSC_2 (g cm^{-3})	/	9.82E-6	6.93E-6	5.12E-6
SSC_3 (g cm^{-3})	/	/	1.13E-5	8.37E-6
SSC_4 (g cm^{-3})	/	/	/	1.59E-5
SSA_{v1} ($\text{cm}^2 \text{cm}^{-3}$)	2.63E-2	1.31E-2	9.27E-3	6.85E-3
SSA_{v2} ($\text{cm}^2 \text{cm}^{-3}$)	/	7.07E-2	4.99E-2	3.69E-2
SSA_{v3} ($\text{cm}^2 \text{cm}^{-3}$)	/	/	8.16E-2	6.03E-2
SSC_{v4} ($\text{cm}^2 \text{cm}^{-3}$)	/	/	/	1.15E-1
C_{o1} (g cm^{-3})	1.22E-9	6.10E-10	4.31E-10	3.18E-10
C_{o2} (g cm^{-3})	/	3.29E-9	2.32E-9	1.72E-9
C_{o3} (g cm^{-3})	/	/	3.79E-9	2.80E-9
C_{o4} (g cm^{-3})	/	/	/	5.33E-9
C_{s1} (g cm^{-3})	3.84E-9	1.92E-9	1.35E-9	1.00E-9
C_{s2} (g cm^{-3})	/	1.03E-8	7.30E-9	5.40E-9
C_{s3} (g cm^{-3})	/	/	1.19E-8	8.82E-9
C_{s4} (g cm^{-3})	/	/	/	1.68E-8
k	3.71E-4	3.71E-4	3.71E-4	3.71E-4

The total within-channel Ca^{2+} acquired in each cell is determined by summing the concentrations of within-channel Ca^{2+} derived from each parcel of suspended sediment at the downglacier end of the cell (Table 3). The final model output is a concentration of Ca^{2+} acquired by WSA processes.

Table 3. Predicted within-channel Ca^{2+} acquisition from individual sediment parcels during transit through the theoretical 4-cell hydrological system.

	Cell 1	Cell 2	Cell 3	Cell 4
WSA^{Ca}₁ (g cm⁻³)	2.67E-9	1.62E-9	1.24E-9	9.48E-10
WSA^{Ca}₂ (g cm⁻³)	/	6.70E-9	5.85E-9	4.75E-9
WSA^{Ca}₃ (g cm⁻³)	/	/	7.35E-9	6.76E-9
WSA^{Ca}₄ (g cm⁻³)	/	/	/	9.79E-9
Total WSA^{Ca}		8.32E-9	1.44E-9	2.22E-8

CONCLUSIONS

Glacial meltwaters may acquire a significant proportion of their solute load from within-channel solute acquisition, resulting from the interaction of dilute meltwaters and geochemically reactive suspended sediment. A generic, kinetic model of solute acquisition, which incorporates the effects of the solution concentration deficit and variability in suspended sediment specific surface area by water volume (SSA_v), is applied to a theoretical 4-cell subglacial channelised system. This model, configured to mimic a known hydroglacial configuration, may be used to drive a predictive model of solute acquisition in subglacial channels. Constrained by realistic values of SSA_v and residence time, this allows the magnitude of WSA to be predicted at any distance from the head of the channelised drainage system to the glacier snout.

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