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Fresh arguments against the Shaw megaflood hypothesis. A reply to comments by David Sharpe on "Paleohydraulics of the last outburst flood from glacial Lake Agassiz and the 8200 BP cold event"

1. The megaflood hypothesis

We disagree with the premise underlying most of David Sharpe's comments, namely that the Shaw subglacial megaflood hypothesis enjoys sufficient mainstream acceptance that we were negligent in failing to cite it. Although the literature on Shavian megafloods has grown over the past decade, it is less clear that the ideas have gained ground. As a recent datum, Benn and Evans (2005) assert that "most Quaternary scientists give little or no credence to the [Shaw] megaflood interpretation, and it conflicts with an overwhelming body of modern research on past and present ice sheet beds" adding that "the idea of floods of such unimaginable dimensions is the outcome of taking flawed assumptions to their logical conclusion, a form of reductio ad absurdum in which the final absurdity is taken not as evidence of false premises but as fact." Whatever one makes of these barbed statements, they do not suggest that Quaternarists must accept the megaflood hypothesis as part of their interpretative framework. Others who have examined the hypothesis have drawn attention to the pitfalls of using form analogies (e.g., Benn and Evans, 1998, p. 447) or questioned the flood hydraulics (Walder, 1994). We suspect that these lines of criticism will not deliver a conclusive outcome. Thus, we focus on several conspicuous problems that we consider decisive. In the course of airing these concerns we shall also address the points raised by Sharpe.

2. The volume and magnitude of known megafloods

Although most of Sharpe's comments are directed at Clarke et al. (2004), his comment concerning our claim that the Lake Agassiz flood was the "largest in the last 100,000 years" refers to a different publication (Clarke et al., 2003). We agree that, in terms of peak discharge, both the Missoula floods (e.g., Clarke et al., 1984; O'Connor and Baker, 1992) and the Altay event (Baker et al., 1993) were indeed larger (roughly 17 Sv for Missoula and >18 Sv for Altay) but the released water volume was a small fraction of that released from glacial Lake Agassiz (Table 1). Furthermore, the focus of Clarke et al. (2003) was on abrupt climate change triggered by freshwater injection to the North Atlantic Ocean at 8200 BP. Freshwater volume rather than peak flood discharge is the relevant measure of flood magnitude for activation of this climate switch. Sharpe's comment that "improved knowledge of additional flood terrains is important in assessing the impact of specific outburst floods on rapid climate change" seems to miss the point that flood intensity, which controls the geomorphic imprint, is only a second-order influence on the climate impact.

Our estimate of the maximum released volume of glacial Lake Agassiz at the time of the final flood was $151,000\,km^3$, corresponding to a sea level rise of 0.42 m. If floods from a subglacial or supraglacial Lake Livingstone reservoir "contributed several metres to sea level rise", as Sharpe states, then they would dwarf the Lake Agassiz flood in terms of the released water volume. However, we strongly disagree that so much water could have been released from a huge reservoir beneath the Laurentide Ice Sheet (LIS). For those unfamiliar with the literature on the subject, Lake Livingstone is the inferred subglacial or supraglacial reservoir that held the inferred water that was released by an inferred subglacial flood over a drumlin field in the region of contemporary Livingstone Lake in northern Saskatchewan (Shaw and Kvill, 1984; Shaw et al., 1989). Unlike glacial lakes Agassiz and Ojibway, which are known to have existed and have reasonably well constrained reservoir geometries, the existence and location of glacial Lake Livingstone are unclear. There is no geological expression of the lake extent and thus the reservoir, if it existed at all, was subglacial, supraglacial or some combination of these. Consequently, its volume is only constrained by Shaw's estimate of the amount of water necessary to generate the drumlins in the manner he proposes. We shall argue

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Table 1
Estimated volume of real and conjectured freshwater reservoirs

Reservoir	Volume (10^3 km^3)	Notes
Subaerial reservoirs		
Altay flood reservoirs	1	Baker et al. (1993, Note 24)
Glacial Lake Missoula	2.184	Clarke et al. (1984)
Modern Lake Superior	12.1	www.great-lakes.net/lakes/ref/supfact.html
Glacial Lake Agassiz[1] final flood	151	Clarke et al. (2004)
Subglacial reservoirs		
Lake Vostok, East Antarctica	7	Maximum (Studinger et al., 2004)
All Antarctic subglacial lakes	12	Maximum (Dowdeswell and Siegert, 2002)
Lake Livingstone A	84	Shaw (1989)
Lake Livingstone B	724	Blanchon and Shaw (1995)
Cumulative basal melt of LIS	1860	Fig. 2

[1]As in Clarke et al. (2004) we refer to the superlake formed by the coalescence of glacial lakes Agassiz and Ojibway as "Lake Agassiz".

that Lake Livingstone either never existed or, if it did exist, it could not have been huge. Even Shaw's lower limit on reservoir volume of $84,000\,\mathrm{km}^3$ (Shaw, 1989, pp. 855; Shaw et al., 1989, p. 196) is problematic.

Sharpe boldly asserts that the volume of Lake Livingstone was even greater than that of Lake Agassiz. In Shaw et al. (1989, p. 196) the volume of the water required to produce the Livingstone Lake drumlin field by their subglacial sheet flood mechanism is estimated as $84,100\,\mathrm{km}^3$ (Lake Livingstone A in Table 1). The basis of this estimate is to calculate the volume (relative to a flat topographic surface) of the drumlins that comprise the Lake Livingstone drumlin field and equate this to the volume of ice that would need to have been melted from the base of the ice sheet in order to form the cavities from which the drumlin features were assumed to evolve. The corresponding water volume is found using an energy balance calculation, assuming 33% efficiency of the melting process.

We have not succeeded in tracing the source of Sharpe's claim that the "full Livingstone Lake event contributed several metres to sea level rise". He cites Blanchon and Shaw (1995) but close reading of that paper indicates that their volume estimate was 0.23 m sea-level equivalent ($84,000\,\mathrm{km}^3$) (Blanchon and Shaw, 1995, p. 7), matching the original estimate by Shaw et al. (1989). According to Shaw (1989, pp. 855–856), if "several drumlin fields in North America and Europe were formed in this way [i.e., by sheet floods], more or less contemporaneously, the total volumes of meltwater involved could have caused several metres of eustatic sea-level rise in a few years". In a subsequent paper, Shaw (2002, p. 16) states that the $84,000\,\mathrm{km}^3$ flow was "just a thin filament of the Livingstone Lake Event" and then incorrectly cites Shaw (1989) to support the claim that "such sea level rises were probably on the order of metres". To obtain a second volume estimate for the reservoir, we shall accept that $84,000\,\mathrm{km}^3$ represents a thin filament of the lake. Conservatively, we translate "several" to equal 2 m of sea level rise to arrive at $724,000\,\mathrm{km}^3$ (Lake Livingstone B in Table 1).

3. The problem of water storage

Because there is no geomorphic or sedimentary record of large reservoirs, such as the inferred Lake Livingstone reservoir, a leading problem of the megaflood hypothesis is where such a huge volume of freshwater could be sequestered. Early papers suggested that the reservoir was subglacial (e.g., Shaw et al., 1989, p. 199) whereas more recent publications seem to prefer a supraglacial reservoir (e.g., Shaw, 1996, pp. 225–226). The problem is not whether subglacial or supraglacial freshwater reservoirs can exist: they can and do. Whether *huge* subglacial or supraglacier reservoirs did exist is an entirely different question. We can think of no direct evidence to support this conjecture and much to suggest it is baseless.

3.1. Subglacial storage

For simple geometrical reasons, ice sheets have a strong tendency to expel subglacial water rather than store it. This is consistent with the modest estimates of the volume of stored water beneath contemporary ice sheets, although not with the requirements of the megaflood hypothesis. For example (Table 1), the maximum estimated volume of Lake Vostok is $7000\,\mathrm{km}^3$ (Studinger et al., 2004) and a recent estimate of the total volume of subglacial lakes beneath the Antarctic Ice Sheet is in the range $4000–12,000\,\mathrm{km}^3$ (Dowdeswell and Siegert, 2002).

Water pressure beneath a warm-bedded ice sheet closely approximates the ice overburden pressure $p_\mathrm{I} = \rho_\mathrm{I} g(Z_\mathrm{S} - Z_\mathrm{B})$ where ρ_I is the density of ice, g is the gravity acceleration, Z_S is the ice sheet surface elevation and Z_B is the bed surface elevation. The fluid potential that drives subglacial water flow can therefore be written as $\phi = \rho_\mathrm{I} g(Z_\mathrm{S} - Z_\mathrm{B}) + \rho_\mathrm{W} g Z_\mathrm{B}$ and the downslope

Table 2
Physical constants and model parameters

Property	Value	Units	Source
Density of water, ρ_W	1028	$kg\,m^{-3}$	[1]
Density of ice, ρ_I	910	$kg\,m^{-3}$	[1]
Density of lithosphere, ρ_L	3300	$kg\,m^{-3}$	[1]
Gravity acceleration, g	9.81	$m\,s^{-2}$	[1]
Thermal conductivity of ice, K	2.1	$W\,m^{-1}\,K^{-1}$	[1]
Specific heat capacity of ice, c	2009	$J\,kg^{-1}\,K^{-1}$	[1]
Latent heat of melting of ice, L	3.35×10^5	$J\,kg^{-1}$	[1]
Pressure melting coefficient (unsaturated), C_T	7.42×10^{-8}	$K\,Pa^{-1}$	[2]
Sea level temperature, T_{SL}	−6	°C	[3]
Thermal lapse rate, β	0.01	$K\,m^{-1}$	[1]
Sea level accumulation rate, b_{SL}	0.3	$m\,yr^{-1}$	[1]
Length scale for b decay rate, Z_0	1000	yr^{-1}	[1]
Geothermal heat flux, q_G	0.045	$W\,m^{-2}$	[1]

Notes: [1] Marshall and Clarke (1997b); [2] Paterson (1994); [3] assigned so as to give reasonable predictions of LGM surface temperature of maximum ice sheet.

potential gradient is

$$\frac{d\phi}{dx} = \rho_I g\,\frac{dZ_S}{dx} + (\rho_W - \rho_I)g\,\frac{dZ_B}{dx} \tag{1}$$

(e.g., Paterson, 1994, pp. 112–113). The condition for subglacial water ponding is that $d\phi/dx = 0$, which implies that for ponding to occur an adverse bed slope (i.e., opposite to the surface slope) of magnitude $dZ_B/dx \geqslant -[\rho_I/(\rho_W - \rho_I)]\,dZ_S/dx$ must exist.

The pressure effect on the freezing temperature of water gives rise to a phenomenon known as "glaciohydraulic supercooling" (Röthlisberger and Lang, 1987; Alley et al., 1998). A detailed analysis of this effect shows that flowing subglacial water will become supercooled when an adverse bedslope exceeds

$$\frac{dZ_B}{dx} = -\frac{(1-\rho_W c_W C_T)\rho_I}{\rho_W - (1-\rho_W c_W C_T)\rho_I}\,\frac{dZ_S}{dx}, \tag{2}$$

where $c_W = 4220\,J\,kg\,K^{-1}$ is the specific heat capacity of water and $C_T = 0.0742\,K\,MPa^{-1}$ is the pressure melting coefficient.

Isostatic adjustment of the bedrock surface works to inhibit the generation of large adverse bed slopes. For isostatic equilibrium of an ice sheet resting upon a lithospheric platform of density ρ_L, the condition for buoyancy equilbrium is $\rho_L(Z_L - Z_B) = \rho_I(Z_S - Z_B)$ where Z_L is the unloaded elevation of the lithospheric surface. Taking spatial derivatives of this expression gives the result that $dZ_B/dx = -[\rho_I/(\rho_L - \rho_I)]\,dZ_S/dx$.

Taking density values for ρ_I, ρ_W and ρ_L from Table 2 gives

$$dZ_B/dx \geqslant -8\,dZ_S/dx \quad \text{geometrical ponding condition,} \tag{3a}$$

$$dZ_B/dx \geqslant -1.6\,dZ_S/dx \quad \text{supercooling condition,} \tag{3b}$$

$$dZ_B/dx = -0.4\,dZ_S/dx \quad \text{isostatic condition.} \tag{3c}$$

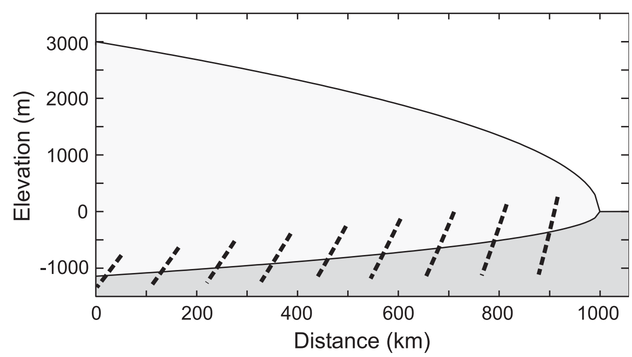

Fig. 1. Geometric relationships for isostatic balance and for subglacial water ponding. Assuming steady-state conditions, the equilibrium profiles of ice- and bed-surface geometries are plotted. The dashed lines that intersect the bed profile illustrate the adverse bed slopes required for ponding of subglacial glacial water according to the criterion of Eq. (3a).

The implications of the above expressions are as follows: (i) Relative to the surface slope of the ice sheet, very large adverse bed slopes are required for subglacial ponding. (ii) Isostatic adjustment suppresses the development of the large adverse bed slopes required to form subglacial water reservoirs. (iii) The isostatically prescribed bedslope is less than the slope for the onset of hydraulic supercooling. Because of the changing geometry of the LIS, isostatic equilibrium cannot be achieved. However, the characteristic time for isostatic adjustment, ~ 5000 yr for the LIS (e.g., Marshall et al., 2000), is much shorter than the growth time for a huge subglacial lake, as noted below; thus Eq. (3c) provides a good basis for discussing the isostatic influence on bed topography.

Fig. 1 illustrates the contrast between the bed slope required for isostatic equilibrium and that required for ponding according to Eq. (3a). Ponding requires a large adverse bedslope whereas buoyancy equilibrium requires a small adverse bedslope. Thermal processes, including glaciohydraulic supercooling, might help to confine water subglacially by forming a "cold ring" around the water body. However, these processes are unlikely to be conducive to the formation of a huge subglacial reservoir because any breach of the ring, for example an ice stream emanating from within the ring boundary, would limit the lake growth. The existence of subglacial Lake Vostok, East Antarctica, does not weaken the claim that large ice sheets have a strong tendency to expel water. It simply demonstrates that Lake Vostok exists because of special circumstances: the lake is situated in a tectonic depression (Studinger et al., 2003) and by this means escapes the water-expelling tendency implied by Eq. (3c). The formation of a huge subglacial lake beneath the LIS would require the existence of a huge, steep-sided, pre-glacial physiographic basin.

3.2. Supraglacial storage

The formation of huge supraglacial lakes is equally challenging to explain. First, their formation is restricted to times when extensive surface melting occurs. Thus the proposal that Heinrich events are triggered by the release of water from a huge supraglacial reservoir (Shaw, 1996, p. 228) is not credible because Heinrich events are known to be initiated during periods of extreme cold (Bond et al., 1993) when surface melting would be at a minimum. However, we agree with Shaw's appraisal of the situation: "It is important that the supraglacial lake and the subglacial meltwater system remained disconnected as the lake grew. If this had not been the case, catastrophic drainage of the required magnitude could not have happened" (Shaw, 1996, p. 226). It is an inescapable truth that a huge lake will have a large filling time, a large area and substantial depth. None of these features favours the formation of huge supraglacial lakes. Large supraglacial lakes are unlikely to remain stable for the time required to allow them to grow huge.

The annual drainage of small supraglacial lakes in west central Greenland (e.g., Zwally et al., 2002) reinforces the point that large lakes cannot remain stable for a long time, even if underlain by cold ice. The greater the lake area the greater the likelihood that the lake will encounter a downward drainage connection or overtop its margins, emptying the reservoir. Reducing the area of the lake only forces it to be deeper, thereby reducing the distance between the ice surface and the bed—again increasing the likelihood of premature drainage. If Lake Livingstone A ($84,000\,\mathrm{km}^3$) was supraglacial and formed from a drainage catchment having an area of $10^6\,\mathrm{km}^2$ (e.g., Shoemaker, 1992, p. 1250) with a melt rate of $1\,\mathrm{m\,yr}^{-1}$, it would take 84 yr to fill the lake and the resulting lake would have an average depth of 84 m. By a similar calculation, if Lake Livingstone B ($724,000\,\mathrm{km}^3$) was supraglacial with the same catchment area and melt rate, it would take 724 yr to fill and have an average depth of 724 m. If the area of Lake Livingstone B equalled the entire area of the LGM LIS, the lake would still be 45 m deep. (According to Rains et al. (1993, p. 326) the flood occurred around the time of the LGM; according to Marshall et al. (2000, Fig. 5) the LGM area of the LIS, excluding the Cordilleran Ice Sheet, was $15 \times 10^6\,\mathrm{km}^2$.) Thus huge supraglacial lakes must be simultaneously areally-extensive and deep, characteristics that are incompatible with the existence of a stable reservoir to accumulate and ultimately release surface meltwater.

For the sake of argument, let us accept the melting zone of the west-central Greenland Ice Sheet as a modern analogue to the LIS during deglaciation. At Swiss Camp (69.57°N, 49.31°W), near the present equilibrium line, the ice thickness is around 1200 m and the 10-m ice temperature is around $-12.5\,^\circ\mathrm{C}$ (Wang et al., 2002). Citing Thomsen et al. (1998), Zwally et al. (2002) claim that the aerial density of moulins (vertical water conduits) in the vicinity of Swiss Camp is $0.2\,\mathrm{km}^{-2}$. This is an area where surface melting is active and numerous supraglacial lakes form. Despite the large supply of meltwater, these lakes do not manifest a steady increase in volume but tend to fill and drain on an annual basis. Taking the areal density of moulins from Greenland as a guide, a $10^6\,\mathrm{km}^2$ lake would encounter 2×10^5 moulins, any one of which could drain the supraglacial reservoir. Decreasing the area to escape this difficulty simply introduces a new one: reducing the distance between the surface reservoir and the subglacial bed increases the likelihood of establishing a connection between water ponded on the ice sheet surface and the subglacial bed. Although it might be argued that present conditions in west-central Greenland are warmer and wetter than those prevailing at the surface of the LIS, this defence simply substitutes one problem for another. If the surface temperature of the LIS was substantially colder, surface melting would become negligible and supraglacial lakes could not form.

4. The problem of meltwater supply

A separate challenge for proponents of the megaflood hypothesis is to identify the source of the meltwater released during the postulated floods. In this section we consider the question of meltwater supply for subglacial, supraglacial and proglacial water.

4.1. Subglacial supply

To gain a measure of the volume of subglacial meltwater available to form subglacial lakes beneath

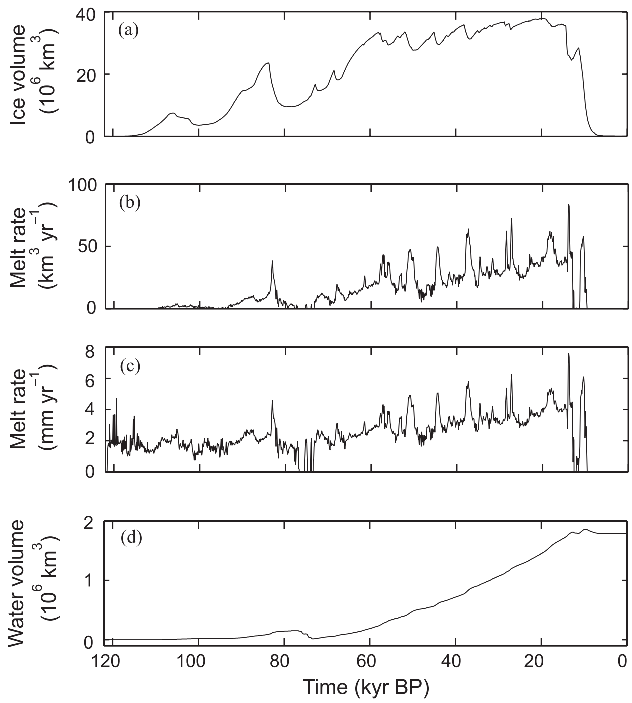

Fig. 2. Simulated melt history of the Laurentide Ice Sheet. (a) Ice volume (10^6 km^3). (b) Basal melt rate for entire ice sheet (km^3 yr^{-1}). (c) Area-averaged basal melt rate (mm yr^{-1}). (d) Cumulative basal meltwater production (10^6 km^3). Times at which the cumulative melt decreases are associated with net basal freeze-on.

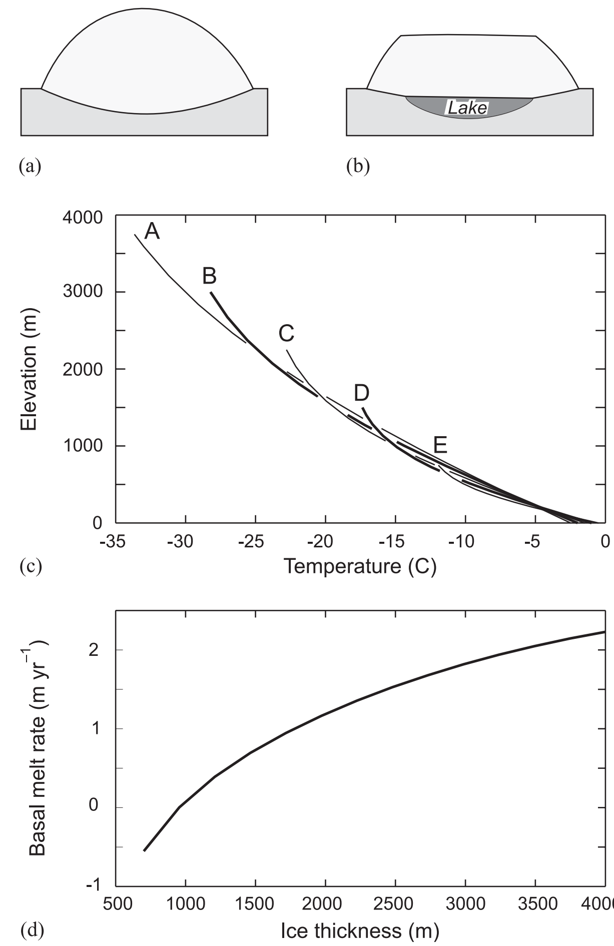

Fig. 3. Calculation of the strength of the negative feedback between ice thickness and basal melt rate. (a) Ice sheet resting on bedrock. (b) Ice sheet underlain by a large subglacial lake. Decoupling of the lower boundary from the glacier bed leads to thinning of the ice sheet and flattening of the upper surface. (c) Calculated ice temperature profiles for a range of ice sheet thicknesses. (d) Basal melt rate as function of ice sheet thickness.

the LIS, we have used a state-of-the-art thermomechanical ice dynamics model (Marshall and Clarke, 1997a; Marshall et al., 2000) to estimate the meltwater production beneath the LIS during the last glacial cycle (Fig. 2). The assumed model parameters and climate forcings are described in Marshall et al. (2000). Note in Fig. 2 that the LIS is near its maximum volume from around 60 to 17 kyr BP (Fig. 2a) but the subglacial melt rate has an increasing trend over this time interval (Fig. 2b), an indication that much of the base of the ice sheet was initially cold. The area-averaged basal melt rate (Fig. 2c) follows a similar trend but with different scaling. Finally, the cumulative melt volume (Fig. 2d) remains low until around 75 kyr BP and then rises monotonically to reach a maximum around 10 kyr BP. At this maximum, the cumulative melt volume is 1.86×10^6 km^3, somewhat more than double the volume of Lake Livingstone B. The implication is that if the entire basal meltwater production of the LIS could somehow be stored without leakage losses, which is highly improbable given the ice marginal recession that occurred before 10 kyr BP, there would be sufficient water to fill two reservoirs the size of Lake Livingstone B. We deem it highly unlikely that such efficient storage could occur and consider 10% of the cumulative total to be an upper limit on the stored total volume. With these assumptions there is sufficient basal meltwater to fill only 25% of Lake Livingstone B by 10 kyr BP. This volume may be sufficient to form a single drumlin field, if discharge could be focused along a single path, but it could hardly account for the continental array of such features.

Other difficulties arise. As noted by Shoemaker (1990, 1999), the formation of a large subglacial lake will flatten the ice surface above the lake and, in doing so, thin the ice cover. The thermal consequences of this ice sheet response have not previously been examined but we will show that the formation of a large subglacial lake near the centre of an ice sheet leads to a reduction in the basal melt rate—a negative feedback that reduces the rate of growth of lake volume. For simplicity we represent the ice as a horizontal slab (Fig. 3b) and assume steady-state conditions so that energy transport and bottom melting are governed by the following equations:

$$K\,\frac{\mathrm{d}^2T}{\mathrm{d}z^2}=\rho_{\mathrm{I}}c\,v\,\frac{\mathrm{d}T}{\mathrm{d}z}\quad \text{heat transfer,} \tag{4a}$$

$$v(z)=-m-(b-m)\,\frac{z}{H}\quad \text{downward ice velocity,} \tag{4b}$$

$$m=\frac{q_{\mathrm{G}}-q_{\mathrm{I}}}{\rho_{\mathrm{I}}L}\quad \text{basal melting rate,} \tag{4c}$$

$$q_{\mathrm{I}}=-K\,\frac{\mathrm{d}T(0)}{\mathrm{d}z}\quad \text{heat flux into basal ice,} \tag{4d}$$

where K is the thermal conductivity of ice, T is ice temperature, c is the specific heat capacity of ice, v is the vertical component of ice velocity, z is a positive-upward distance coordinate, H is the slab thickness, m is the basal melt rate ($\mathrm{m\,s^{-1}}$), b the ice-equivalent surface accumulation rate ($\mathrm{m\,s^{-1}}$), q_{G} is the geothermal flux, q_{I} the heat flux into the base of the ice column and L the latent heat of melting. The lower surface of the slab is at $z=0$ where z is the elevation above the bed rather than elevation above sea level. In a geographical coordinate system we take $Z=0$ to correspond to sea level, with Z_{B} and Z_{S} corresponding to the elevation of the lower and upper boundaries of the ice sheet so that $H=Z_{\mathrm{S}}-Z_{\mathrm{B}}$ corresponds to the slab thickness. The boundary conditions on the ice slab are taken as

$$T_{\mathrm{B}}=-C_{\mathrm{T}}\rho_{\mathrm{I}}gH, \tag{5a}$$

$$T_{\mathrm{S}}=T_{\mathrm{SL}}-\beta Z_{\mathrm{S}}, \tag{5b}$$

where C_{T} is the pressure melting coefficient, β the atmospheric lapse rate and, following Marshall and Clarke (1997b), the "elevation desert" effect on surface accumulation is described by the function

$$b=b_{\mathrm{SL}}\exp(-Z_{\mathrm{S}}/Z_0) \tag{6}$$

and the assumption of isostatic equilibrium gives the elevation of the lower and upper ice boundaries as

$$Z_{\mathrm{B}}=Z_{\mathrm{L}}-\frac{\rho_{\mathrm{I}}H}{\rho_{\mathrm{L}}}, \tag{7a}$$

$$Z_{\mathrm{S}}=Z_{\mathrm{B}}+H. \tag{7b}$$

The assignment of physical constants in Table 2 is consistent with that of Table tabplace1 in Marshall and Clarke (1997b) but the qualitative results of our calculation are insensitive to these assignments.

Fig. 3 summarizes the results of the simulations. Ice thinning associated with the formation of a large subglacial lake is sketched in Fig. 3a,b. Temperature profiles for a range of ice thicknesses are presented in Fig. 3c. Thicknesses range from 3750 m (curve A) to 1000 m (curve E). Because the surface temperature is elevation-dependent a thin ice slab has a warmer surface than a thick one. The surface accumulation rate b is also elevation-dependent, as indicated in Eq. (6), and this leads to an increasing influence of vertical advection as the slab thickness decreases. The temperature at the base of the slab varies with thickness owing to the pressure influence on melting temperature. Fig. 3d summarizes the main result of this modelling study: the melt rate beneath a thin ice slab is less than that beneath a thick one. Thus the thinning accompanying the formation of a large subglacial lake introduces a negative feedback that opposes the growth of the lake and would further exacerbate the problem of basal meltwater supply.

4.2. Supraglacial supply

During deglaciation, surface melt rates can be large and an abundance of meltwater is available to form lakes. Proglacial lakes are especially favoured because the receding ice front leaves a depressed land surface in its wake (e.g., Marshall and Clarke, 1999, Fig. 5) and regional drainage patterns are dammed. For reasons discussed in Section 3.2, supraglacial lakes are unlikely to be long-lived or large.

4.3. Proglacial supply

Shoemaker was clearly troubled by the water supply problem and proposed several ingenious approaches to ameliorating it. One such idea was that of "reverse floods" (Shoemaker, 1992, pp. 1257–1258), which involved floods of water *from* proglacial lakes delivering water *to* a subglacial lake which he positioned in Hudson Bay. Such "inburst floods" are not impossible but occur under restrictive conditions: a favourable hydraulic gradient must exist between the source and the sink and subglacial water pressure must approach or exceed the ice overburden pressure along the water flow path. Our depiction of the Lake Agassiz outburst (Clarke et al., 2004) might be regarded as a reverse flood but for the fact that the water was delivered to the Tyrrell Sea rather than to a subglacial reservoir. We have no argument with paleogeographic inferences suggesting that, under special circumstances, water flowed from proglacial lakes along the southern margin of the LIS to an outlet in the vicinity of Hudson Bay. This is indeed what we proposed as did Dredge (1983) and Klassen (1983, 1986).

However this is not the point of Sharpe's comment or of Shoemaker's proposal. For Shoemaker the essential role of reverse floods was to deliver water to a subglacial reservoir where it could be stably stored and subsequently released as a drumlin-forming subglacial megaflood. For a reverse flood to be possible, Hudson Bay must be either open or ice-covered with such low topographic surface (hence very thin ice) that it would seem impossible to maintain the required hydraulic isolation from the Labrador Sea. Furthermore the sequence of inburst and outburst flooding would need to proceed very rapidly because the LIS was itself undergoing rapid disintegration. In our judgement, reverse floods cannot make a significant contribution to subglacial water storage.

An alternative process for converting proglacial water to subglacial water might be to capture it during the expansion of an ice sheet. Shoemaker (1991, pp. 1978–1979) proposed that as an ice sheet advanced over pre-existing water bodies (e.g. Hudson Bay or the Great Lakes) ice shelves would have spread across these lakes. Continued advance of the ice sheet would entrap these water bodies beneath the ice. More recently Alley et al. (in press) have developed a similar argument in an attempt to explain how Heinrich events occur during cold climate phases. The process can probably operate and might well have been involved in trapping water from Hudson Bay or from the Great Lakes, but the water volumes resulting from this water-trapping mechanism are unlikely to result in huge volumes of trapped subglacial water. Table 1 shows that the volume of modern Lake Superior was only 14% of that of Lake Livingstone A and 1.7% that of Lake Livingstone B.

5. Sheet-like outbursts from contemporary reservoirs

Sharpe would like us to have considered alternative geometries for the subglacial drainage conduit and points to the publication by Björnsson (2002) as an example. He seems to imply that contemporary sheet-like floods, such as that associated with the 1996 outburst from subglacial lake Grímsvötn in Vatnajökull ice cap, are modern analogues to the sheet-like floods invoked in the megaflood hypothesis. Although we ignored the possibility of a sheet-like flood from Lake Agassiz we do not, in general, rule out this possibility (e.g., the recent paper by Flowers et al. (2004), of which Björnsson and Clarke are co-authors, on the hydraulics of the 1996 Grímsvötn flood). Our motivation for not entertaining the possibility of a similar sheet-like flood from glacial Lake Agassiz is that we consider it highly unlikely that the flood proceeded in this manner. It is more likely that the thinning ice dam over Hudson Bay, resting on an irregular bed and variously attacked by iceberg calving on both aquatic margins, reached a critical thickness at some point that allowed channelized outflow from Lake Agassiz to develop than that this critical thickness was reached simultaneously across the width of Hudson Bay, thus allowing Sharpe's drumlin forming sheet flow.

In our opinion the unusual hydrograph and sheet-like conduit geometry associated with the 1996 Grímsvötn flood result from special conditions that are unlikely to be reproduced within the Lake Agassiz reservoir. Specifically, these are the presence of an actively erupting volcano within the reservoir and the fact that

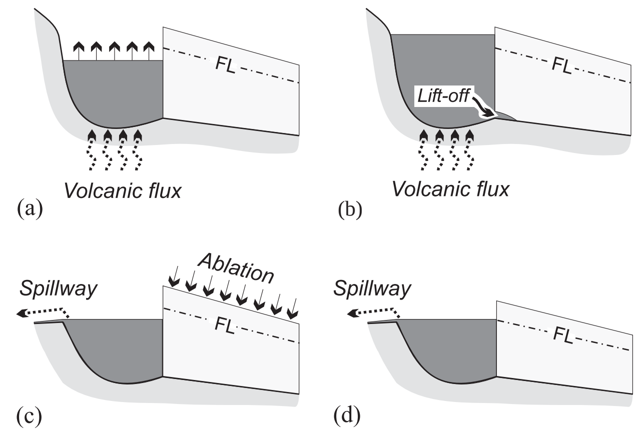

Fig. 4. Contrasting flood release mechanisms. (a,b) Flood release by raising lake level above the ice flotation line (FL). If this overfilling occurs rapidly a sheet flood might result. This is the likely release mechanism for the 1996 outburst from Grímsvötn in Vatnajökull ice cap. (c,d) Flood release by lowering the ice dam. In this situation the lake level cannot rise to the flotation line (FL) because of the presence of a spillway controlling water level. Flood release occurs when the dam thins.

the reservoir could be rapidly overfilled because the lake level was not subject to a topographic control. At the time of the final flood from Lake Agassiz, the lake level is thought to have been controlled by a spillway that routed water through the valley of the Kinojévis River to the valley of the Ottawa River (Vincent and Hardy, 1979; Veillette, 1994; Leverington et al., 2002). We estimated the height of this spillway as 230 m above sea level at that time. The presumed trigger for the 1996 Grímsvötn flood was rapid flotation of the ice dam caused by the fast rise in lake level (Fig. 4a,b). Since rapid overfilling is not possible for Lake Agassiz, we presume the flood was triggered by slow downwasting of the ice dam (Fig. 4c,d).

6. Red bed

Our intent in calling attention to the red clay marker bed in Hudson Bay, which is considered to have been deposited contemporaneously with the Agassiz flushing event (see Barber et al., 1999), was simply to indicate that this bed, if it has a source in red Dubawnt till of the western Hudson Bay region, favours a western rather than eastern Hudson Bay routing for the flushing event. The exact process of deposition has little bearing on the problem we were addressing. We do not hold a strong opinion on the matter, not having examined the bed.

7. Concluding remarks

We conclude that a fatal problem of the meltwater hypothesis for drumlin formation is the requirement that a huge volume water be subglacially or supraglacially stored, then suddenly released. Efforts to salvage the core idea, that drumlins have a subglacial fluvial origin, should focus on ways to reduce the water volume requirements.

Meanwhile, other lines of investigation hold promise. Contemporary West Antarctic ice streams have width scales (Raymond et al., 2001, Fig. 3) that are comparable to that of the Livingstone Lake drumlin field, and submarine deglaciated surfaces beyond the grounding lines of currently active ice streams show drumlin-like features (Shipp et al., 1999). Acoustic imaging of the sea floor off the Antarctic Peninsula also reveals spectacular drumlin-like bedforms and megalineations that are attributed to a paleo-ice stream (Ó Cofaigh et al., 2002). Although these features cannot be linked to a currently active ice stream, the geotechnical properties (Dowdeswell et al., 2004) of the sediments are strikingly similar to those encountered beneath active West Antarctic ice streams. Finally, the subglacial environment is remarkably complex and involves the interplay of thermal, mechanical, hydrological and soil-mechanical processes (e.g., Clarke, 2005). It is conceivable that the problem of drumlin genesis cannot be solved until we gain a better understanding of subglacial process interactions.

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