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Application of 2D hydrodynamic modelling to high-magnitude outburst floods: An example from Kverkfjöll, Iceland

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Abstract

High-magnitude outburst floods cause rapid landscape change and are a hazard to life, property and infrastructure. However, high-magnitude fluvial processes and mechanisms of erosion, transport and deposition are very poorly understood, and remain largely unquantified. This poor understanding is partly because of the inherent difficulty of directly measuring high-magnitude outburst floods, but also because of the limitations and assumptions of 1D models and other palaeohydrological methods, which reconstructions of high-magnitude floods have to date relied upon. This study therefore applies a 2D hydrodynamic model; SOBEK, to reconstructing a high-magnitude outburst flood. This method offers the first calculations of high-magnitude fluvial flow characteristics within an anastomosing network of simultaneously inundated channels, including; sheet or unconfined flow, simultaneous channel and sheet flow, flow around islands, hydraulic jumps, multi-directional flow including backwater areas, hydraulic ponding and multiple points of flood initiation. 2D-modelling of outburst floods clearly has the potential to revolutionise understanding of high-magnitude spatial and temporal hydraulics and high-magnitude flow phenomena, geomorphological and sedimentological processes, and hence rapid fluvial landscape change. This potential for new understanding is because of the now wide availability of high-resolution DEM data for large and often inaccessible areas, and the availability of remotely-sensed data that can parameterise outburst flood sources, such as glacial lakes, for example. Additionally, hydraulic models and computing power are now sufficient to cope with large (5,000,000 grid cells) areas of inundation and large (100,000 $m^3 s^{-1}$) peak discharges.

Keywords: High-magnitude outburst floods; Palaeohydrology; Hydraulics; 2D hydrodynamic; Iceland

1. Introduction, background and rationale

High-magnitude outburst floods are sudden releases of water and sediment with peak discharges that are at least several orders of magnitude greater than perennial flows. In exceptional cases, high-magnitude flood discharges could be comparable to

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ocean current flow rates (Baker, 2002). High-magnitude floods can erode both unconsolidated sediments (e.g. Russell and Marren, 1999; Gomez et al., 2002) and bedrock (e.g. Baker, 1988; Tómasson, 1996; Carrivick et al., 2004a). Additionally, high-magnitude floods transport and deposit vast amounts of sediment (up to 10^8 tons; Björnsson, 2002) over extensive areas. High-magnitude floods therefore constitute a serious threat to life, property and infrastructure (see Costa and Schuster, 1988; O'Connor et al., 2002 for references).

Landscape changes due to high-magnitude floods are thus very recognisable and have been described and classified (e.g. Maizels, 1997; Tweed and Russell, 1999; Carrivick et al., 2004a). However, high-magnitude fluvial erosion, transport and depositional mechanisms are very poorly understood, and remain unquantified (Maizels, 1997; Whipple et al., 2000). This poor understanding and lack of quantification is a product of the difficulty of directly measuring high-magnitude floods. High-magnitude floods often occur with little warning and usually reach peak flow within a matter of hours (e.g. Baker, 2000; Björnsson, 2002). Furthermore, it is unlikely that data-logging equipment could withstand flow pressure and bombardment from particles within the flow.

It is crucial to understand and quantify fluvial processes and mechanisms responsible for producing erosional and depositional landforms in high-magnitude floods, because these landforms are 'records' of flood hydraulics at a specific place and in time during a flood. Fluvial landforms thus permit flow reconstructions and can offer quantification of high-magnitude fluvial processes and mechanisms.

To date, understanding of high-magnitude flood processes and mechanisms are largely scaled-up from low-magnitude flow measurements, flume experiments and extrapolated empirical relationships (e.g. Hjulström, 1935; Costa, 1983; Carling and Bevan, 1989; Rushmer et al., 2002). Palaeohydrological methods have used landforms and sediments produced by outburst floods for reconstructing properties of that flow (e.g. Baker, 1973; O'Connor, 1993; Benito, 1997; Rudoy, 2002). Equations have been formulated to relate the size of transported clast to hydraulic parameters (Costa, 1983; Williams, 1984). Cross-sectional geometry, channel slope, hydraulic roughness and the altitude of preserved wash limits or scour lines, can compute peak discharge for that cross-section (e.g. Baker, 2000). These 'slope-area' methods have been computerised to increase efficiency of large-scale studies by utilising 1D cross-sectional data from digitised topographic maps. Thus many more calculations can be performed to iteratively match model output of water surface elevations to field-observed palaeostage indicators (PSIs). Whilst these 1D 'step-backwater' models compute energy-loss between successive cross-sections and either subcritical or supercritical flow regimes, they are unable to model many features of high-magnitude floods. Indeed step-backwater models, slope area methods and other palaeohydrological methods only provide reconstructions of peak discharge. Thus these methods do not provide information on how flow conditions varied before and after peak stage, or how long peak discharge persisted. Other features of high-magnitude floods are also excluded, such as rapidly varied flow, or specifically; simultaneous inundation of multiple channels, sheet or unconfined flow, simultaneous channel and sheet flow, flow around islands, hydraulic jumps, multi-directional flow including backwater areas, hydraulic ponding and multiple points of flood initiation. Without a quantification of the hydraulics associated with these flow conditions, high-magnitude flood impacts cannot be fully understood.

As such, there is a pressing need to apply depth-averaged 2D hydrodynamic models, which can simulate nearly all aspects of 3D fluvial flow, and are well-established in hydrological studies of riverine low-magnitude floods (e.g. Lane, 1998; Lane and Richards, 2000), to high-magnitude outburst floods. Only Denlinger et al. (2002) have applied a 2D hydrodynamic model to reconstructing a high-magnitude outburst flood, but this was limited to $<3500\ \mathrm{m^3\ s^{-1}}$ peak discharge within a topographically-confined, single bedrock channel. Earlier work with 2D models of high-magnitude outburst floods examined the Missoula Floods (Craig, 1987), which have been re-examined by Komatsu et al. (2000), and Miyamoto et al. (2003), again using 2D hydrodynamic modelling.

2D hydrodynamic models are capable of calculating spatial and temporal variations in hydraulic parameters at a resolution far greater than that provided by commercially-published (1:25,000) topographic maps and therefore offer considerable opportunities for researching hitherto unmeasured properties of high-magnitude floods. In particular, 2D-models route flow over a Digital Elevation Model (DEM) and interface with Geographic Information Systems (GISs). 2D hydrodynamic models are therefore particularly suitable for reconstructing high-magnitude floods in order to understand precise processes and mechanisms. With this information, 2D high-magnitude flood models within a GIS have the potential for being the basis of landscape analysis and of informed policy, mitigation and management of high-magnitude floods.

2. Aim

The aim of this paper is to demonstrate the capability of a 2D hydrodynamic model for reconstructing characteristics of high-magnitude outburst floods, within anastomosing networks of simultaneously inundated channels. These characteristics are simultaneous inundation of multiple channels, sheet (unconfined) flow, simultaneous channel and sheet flow, flow around islands, hydraulic jumps, multi-directional flow including backwater areas, hydraulic ponding and multiple points of flood initiation.

3. Study site

2D hydrodynamic reconstructions are made of the largest glacial outburst flood (jökulhlaup) from Kverkfjöll; a glaciated stratovolcano on the northern margin of Vatnajökull (Jóhannesson and Saemundsson, 1989) (Fig. 1). This study site is chosen because the proglacial area of the Kverkfjöll Volcanic System (KVS); Kverkfjallarani (Fig. 1), contains comprehensive field evidence of a high-magnitude flood (Carrivick et al., 2004a,b), and because a high-resolution (10 m horizontal grid with sub-metre vertical accuracy) DEM of Kverkfjallarani exists (Carrivick and Twigg, 2004). Kverkfjallarani is dominated by a series of parallel volcanic pillow-hyaloclastite ridges, each typically 100 m high and several kilometres long (Jóhannesson and Saemundsson, 1989). High-magnitude outburst floods with cross-sectional discharges of up to 100,000 m3 s^{-1} routed through Kverkfjallarani during the Holocene, and inundated a series of anastomosing routeways (Fig. 1) (Carrivick et al., 2004a,b). Floods through Kverkfjallarani primarily routed to the north-east along the 25 km long Hraundalur valley system, and along a series of small steep valleys spilling westwards over low cols (Fig. 1).

4. Methodology

The 2D hydrodynamic model used by this research is SOBEK, from Delft, the Netherlands. SOBEK is a graphically orientated model and therefore unlike research code-based models, is relatively easy to use by non-computer programmers. The model routes an initial user-specified hydrograph over a regular grid and is therefore ideal for DEM data input. A full description of the hydrodynamic equations used by the model is available on request from the SOBEK website (http://www.sobek.nl/Home/home.html). A model set-up therefore contains boundary data, constants, a topographic grid (a DEM), a roughness grid and specified outputs, such as flow depth (d), and flow velocity (v), per unit time. Exportation of flow depth and flow velocity grids into a GIS allowed computation of
shear stress (τ) (N m^{-2})

$$\tau = \rho g \, dS$$

stream power (w) (W m^{-2})

$$w = \tau v$$

and Froude number (F) (–)

$$F = \frac{v}{\sqrt{gd}}$$

for example, through simple grid-based analyses, where ρ is the specific density of water (kg m^{-3}), g is acceleration due to gravity (m s^{-1}) and S is the energy gradient, which can be approximated by channel slope. Computations presented in this paper are run over a DEM derived from post-flood data, and assume a fixed channel bed. It should also be noted that besides an input hydrograph, which naturally includes a source volume, peak discharge and flood duration, no other palaeohydrological data is input to

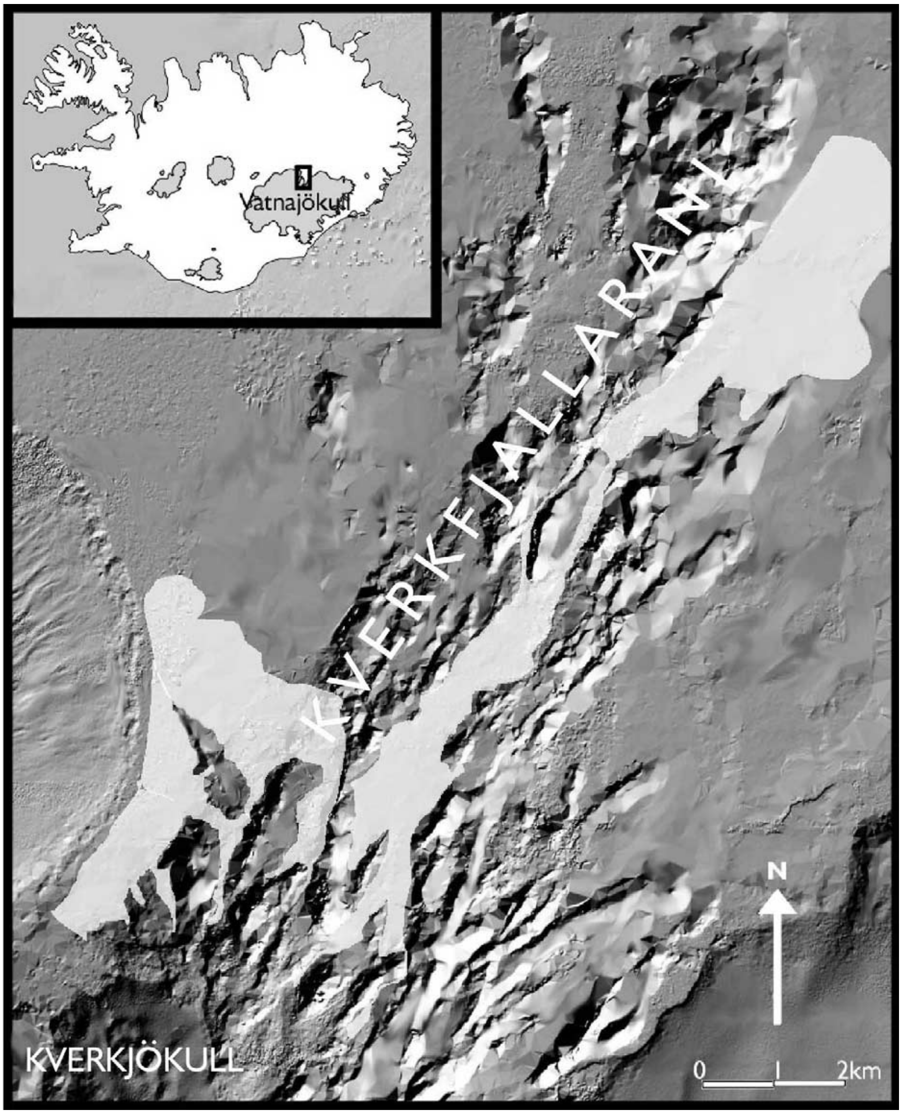

Fig. 1. Location of the proglacial area of Kverkfjöll; Kverkfjallarani, in Iceland. Surfaces inundated by high-magnitude glacial outburst floods are from Carrivick et al. (2004a). Floods primarily routed along the narrow Hraundalur valley system extending over 20 km north-eastwards, and through several small steep valleys over low cols westwards.

a 2D-model. Palaeohydrological data can therefore serve as an independent verification of 2D-model output.

A few general problems were encountered with SOBEK whilst simulating high-magnitude outburst floods through Kverkfjallarani. Excessively long model run times (up to 13 days) were experienced. These long model runs were due to a very large DEM (over 5,000,000 grid cells) and a large volume of water ($\sim$50,000,000 m^3). Additionally, to retain model robustness, i.e. to conserve mass and energy throughout the simulation, the model automatically reduces the computational time-step. Conservation of mass and energy demanded that a small DEM grid size was necessary, given the large volumes of water being modelled. Computational time steps reached 0.001 s due to the large number of active grid cells within a single model.

5. Results and discussion

This research ran a series of model iterations to best-fit the SOBEK model output with field evidence, including inundation extent and PSIs (Carrivick and Twigg, 2004; Carrivick et al., 2004a,b). The initial hydrograph that produced the best-fit model was a linear-rise hydrograph with a peak discharge of 200,000 m3 s^{-1}. It should be noted that this peak discharge was initiated at the 'input boundary' grid cell (10×10 m^2) and not for an entire cross-section. The modelled jökulhlaup inundated the northern-most extent of Kverkfjallarani (Fig. 1) in 3.5 h. The jökulhlaup therefore had a mean frontal velocity of 1.6 ms^{-1}. Jökulhlaup velocity was tremendously varied, with maxima of up to 15 ms^{-1} occurring in constricted and steep zones of flow. Lowest velocities were associated with temporary ponding, immediately upstream of valley constrictions, for example. Flow depths ranged from <1 to 4–5 m. Shear stresses ranged from <100 N m^{-2} in relatively wide reaches up to $\sim$7500 N m^{-2} within narrow and confined channels. Stream power per unit area ranged from $\sim$2000 to $>$ 50,000 W m^{-2} with an average value of 8500 W m^{-2}. Naturally, the greatest stream powers occurred within narrow channels of steeper gradient. Relatively high stream powers were exerted across wide channels and for much of the length of Hraundalur.

However, high-magnitude outburst floods have tremendously varied hydraulics, spatially and temporally. Therefore, summary statements such as those in the paragraph above and generalised landscape scale maps, are really quite insufficient for quantifying high-magnitude flood hydraulics. Rather, by way of specific examples, the following section of this paper will examine the capability of a 2D hydrodynamic model for reconstructing the details of high-magnitude outburst floods, which other models and methods cannot accommodate. For conciseness, data are not exhaustive, but rather designed to give an example of the range of 2D-model output and the possibilities of using this, through a GIS, for hydraulic, geomorphic and sedimentological analyses.

5.1. Multiple points of flood initiation

2D-models have the capacity to reconstruct multiple points of flood initiation. This is particularly useful if a flood source is distant from the area of interest. In the Kverkfjöll example, floods most likely initiated from a volcanic caldera (Carrivick et al., 2004a). Modelling of this outburst flood from the source, would require routing an outburst flood under/through/over the Kverkjökull glacier, a much larger DEM, tunnel/pipe flow hydraulic equations and thermal melting and tunnel widening to be considered, for example. Rather, in accordance with the area of interest, 2D-modelling allows a single outburst flood to be initiated from several simultaneous exit points at the glacier margin, and thence over the proglacial DEM of Kverkfjallarani (Fig. 2). Geomorphological cross-cutting relationships and sedimentary stratigraphy provide evidence that these routeways were inundated by the same outburst flood (Carrivick et al., 2004a,b). Fig. 3 details the velocity fields associated with flows spilling over shallow cols, onto an area of unconfined outwash, as depicted in Fig. 2. Flow velocities of over 10 ms^{-1} are recorded directly at the base of steep slopes and cliffs but quickly dissipate to less than 2 ms^{-1} on the outwash plain (Fig. 2). Flow depths are low over cliffs, as water is drawn down over a lip. By exporting flow depth and velocity grids into a GIS, and performing simple grid-based analyses, 2D-modelling has the capability to describe distributed shear stress and stream power, for example, associated with spillways, cataracts and plunge pools. Hence erosional and depositional thresholds associated with specific hydraulic mechanisms can be described. Shear stresses and stream powers associated with spillways are typically 200–500 N m^{-2} and 310–830 W m^{-2}, respectively.

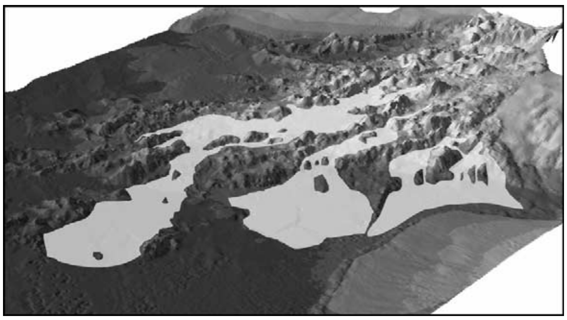

Fig. 2. 3D visualisation of high-magnitude outburst floods from Kverkfjöll, as modelled by routing over a DEM. Note simultaneous inundation of topographically-controlled channels and spills westwards over low cols.

Fig. 3. Example of flow initiated in two channels simultaneously, in the same area as depicted in Fig. 2. Variation in flow velocity demonstrates higher velocities in narrow and steep upper reaches, accelerated velocities over spillways, and braided sheet flow on unconfined outwash area

5.2. Simultaneous inundation of multiple channels

High-magnitude outburst floods cause avulsions, which frequently result in simultaneous inundation of an anastomosing network of channels. The locations of these channels are determined by upstanding topography and discharge (Fig. 4). High-magnitude avulsions have not been modelled previously because 1D models and other palaeohydrological techniques all require the user to specify pre-defined channel characteristics such as channel margins and thalwegs, and hence to pre-define routeways. Multiple channels cannot be modelled simultaneously by 1D models or other palaeohydrological methods because they rely on the transfer of cross-sectionally averaged hydraulics through successive cross-sections, or on individual inferences of hydraulics at a point, drawn from observations of landforms and sediments. Fig. 4 details hydraulics through an initially unconfined reach of relatively shallow (<3 m flow depth), braided flow, which subsequently become channelled through several parallel valleys. Valleys confine flow, raising water depth, typically to 8–10 m (Fig. 4). Some of these valleys are so narrow they cause hydraulic ponding, which will be discussed in Section 5.6. The inundation and depth are in accordance with observations of high-magnitude flood landforms and PSIs (Carrivick et al., 2004a,b).

5.3. Simultaneous channel and sheet flow

High-magnitude floods commonly produce overbank or sheet flow (Fig. 5). Sheet flow can also occur in areas of valley expansion, where valley walls further upstream have channelled flow (Fig. 5). Sheet flow is not uniform, but rather a near-chaotic region of relatively unconfined shallow (<2 m deep) flow, comprising numerous ‘braided’ rivulets without a dominant thalweg (Fig. 5). 1D models cannot reconstruct sheet flow because of the lack of a discernible thalweg and flow margins, and thus because of a lack of a cross-sectional area. Palaeohydrological techniques cannot reconstruct spatial and temporal variations in sheet flow, without recourse to an exhaustive field investigation, such as by Mayer and Pearthree (2002); a 3D sedimentary characterisation of a sedimentary fan or sheet deposit, for example. Regions of sheet flow are often such relatively low-energy, less than 100 W m^{-2} (Fig. 5), that aside from large-scale braided-river type deposits, distinguishable landforms are unlikely to have a high preservation potential.

5.4. Meandering thalweg and flow around islands

Flow along any river channel is disrupted by diversion of flow around obstructions. Thalwegs that change position within a valley are particularly important for assessing geomorphic changes due to outburst floods. As long as the orientation of a thalweg is parallel to valley walls and the valley width does not alter very much, there is relatively little opportunity for development of hydraulic discontinuities and hence of high-magnitude fluvial erosion

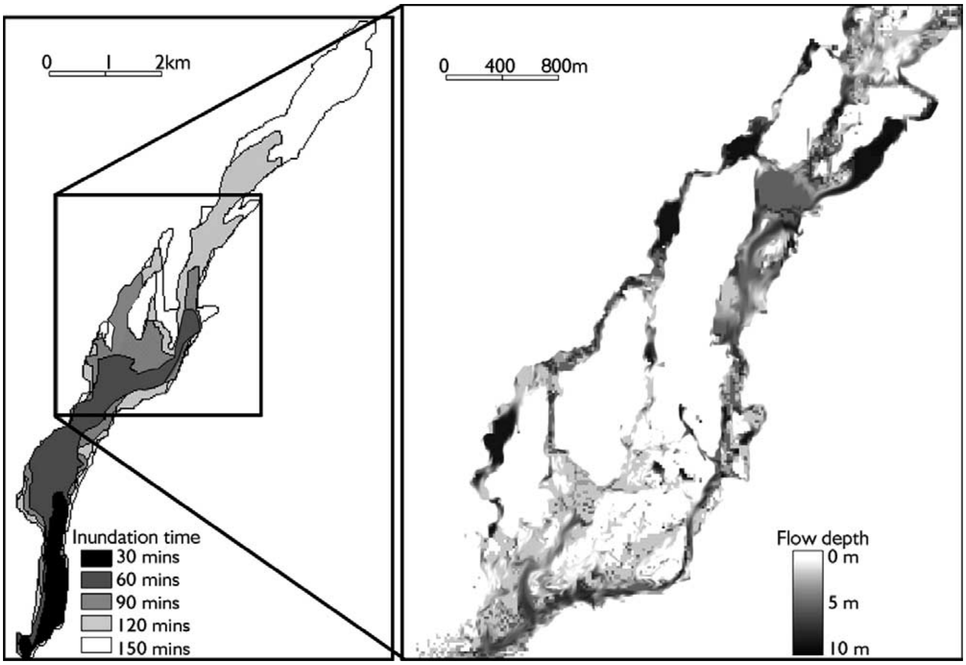

Fig. 4. Variation of flow depth within simultaneously inundated anastomosing channels, along Hraundalur. Inundation is controlled by flow stage and topography. Wider channels feature a meandering thalweg, whilst some narrow channels cause hydraulic ponding.

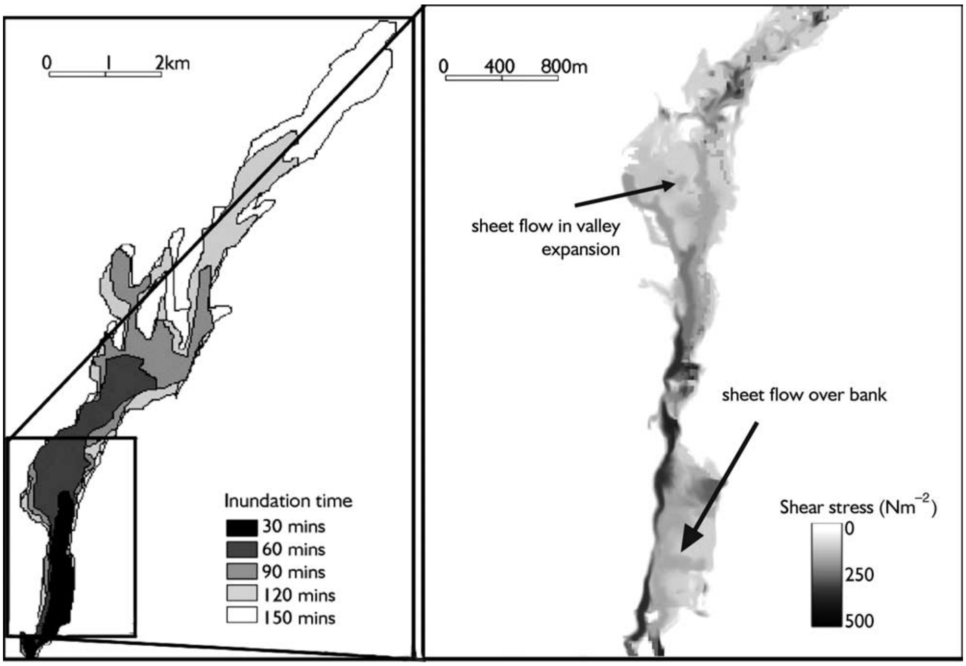

Fig. 5. Variation of shear stress in reaches of simultaneous channel and sheet flow, along Hraundalur. Sheet flow can occur as over-bank flow, or in the absence of a dominant thalweg in zones of valley expansion.

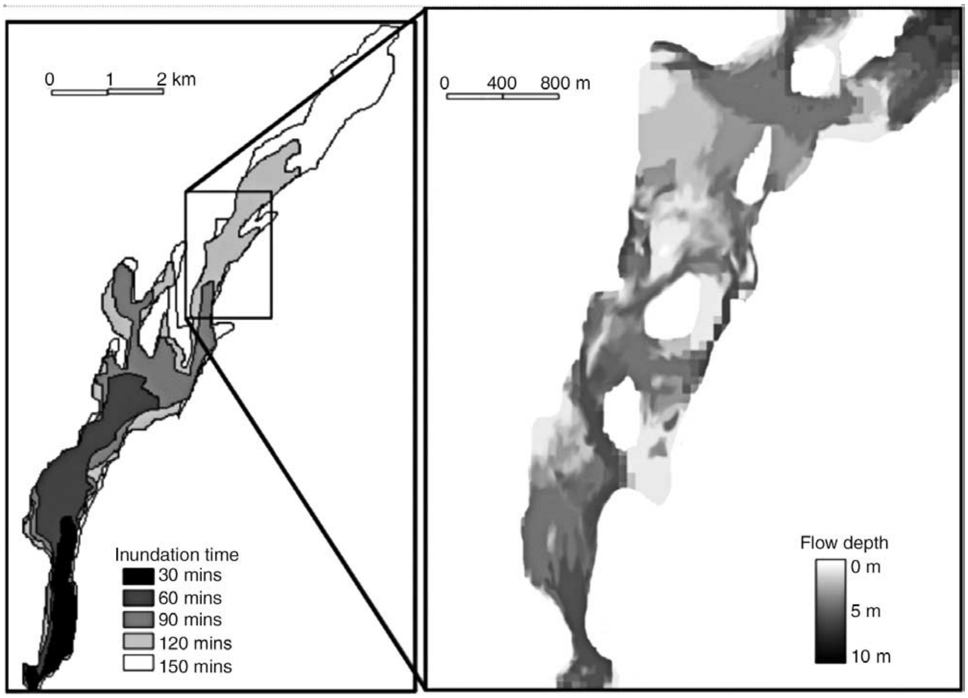

Fig. 6. Variation of flow depth in a zone of flow around mid-channel islands, and with a meandering thalweg, along Hraundalur.

and deposition (Miller, 1995). A meandering thalweg is a product of valley width changes (Fig. 4), and upstanding topographic islands (Fig. 6), but is also stage-dependant. This means that 1D models or other methods of palaeo-discharge reconstruction that average flow velocities per cross-section, could be in error, due to an inability to accommodate thalweg variations, or variations of discharge with time. Velocities within a thalweg can be up to 10 times greater than marginal flow, less than 500 m away (Fig. 6). Variations in the position of a thalweg relative to channel width are deterministic of lateral patterns of erosion and deposition such as scour marks, terrace deposits and bars (e.g. Miller, 1995). Thalweg variations also produce two other characteristics of high-magnitude outburst floods; the super-elevation of water surfaces on outside bends and on upslope sides of topographic islands (Fig. 6), and a surging flow. Super-elevated water surfaces are caused by a surging flow and have been suggested to be responsible for boulder run-ups (Carrivick et al., 2004a).

Meandering thalwegs are discharge-dependant and hence cannot be examined by 1D models or other palaeohydrological methods. Super-elevation of water surfaces is also discharge dependant, but could be recognised from PSI data, as noted by Carling et al. (2002); Carrivick et al. (2004a).

5.5. Hydraulic jumps

Subcritical and supercritical flow regimes commonly co-exist in high-magnitude floods. The transition between the two regimes is a zone of turbulence, high-pressure and potential for high--magnitude fluvial phenomena such as cavitation and plucking (Whipple et al., 2000). These zones have been attributed to the production of cataracts, gorges and bedrock steps for example (Carrivick et al., 2004a), but precise hydraulics remain unquantified. Supercritical flow, where $F > 1$, predominates in narrow, steep channels and subcritical flow, $F < 1$, predominates in wide, shallow reaches (Fig. 7). However, supercritical flow also persists at some channel margins and around islands (Fig. 7). Supercritical flow zones at channel margins in Kverkfjallarani correspond to a gap between valley floor lava

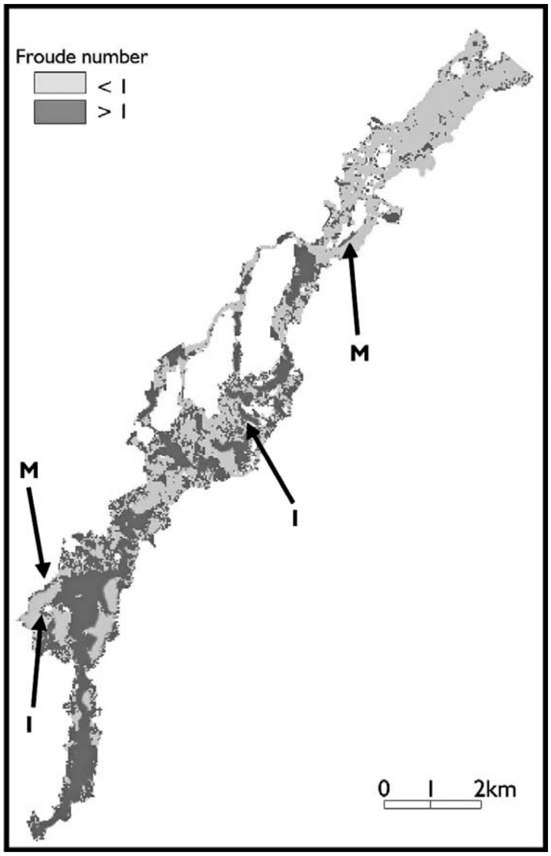

Fig. 7. Variation of Froude number (F), distinguishing subcritical flow ($F<1$), supercritical flow ($F>1$), and thus hydraulic jumps, over the entire 25 km length of along Hraundalur. M denotes valley-marginal supercritical flow, due to a gap between lava flows abutting against valley walls. I denotes supercritical flow around upslope island sides.

flows and valley walls (Carrivick et al., 2004a). Supercritical zones around islands relate to flow acceleration. The fact that peak discharge through Kverkfjallarani attenuated very rapidly, due to very high channel roughness (Carrivick et al., 2004a,b), is evidenced by supercritical flows in the most southerly, or proximal reach, and subcritical flows in the most northerly or distal reach (Fig. 7). Peak discharge attenuated due to high longitudinal energy losses, predominantly caused by high hydraulic roughness; Mannings $n=0.1$ (Carrivick et al., 2004b), but also due to loss of flow through spillways and recirculation of water in backwater areas. Form roughness quantifies the height and density of immobile protrusions into a flow, and is thus stage-dependant. Therefore, flows through Kverkfjallarani attenuated more rapidly in wider, shallower, distal valleys, than in narrower, deeper, proximal reaches (Fig. 7).

1D models or other palaeohydrological techniques cannot model hydraulic jumps or variable regime flows. Step-backwater models can reconstruct either supercritical flows, or subcritical flows, but cannot accommodate both simultaneously. If supercritical flows are modelled by step-backwater technique, cross-sectional discharges are calculated in an upstream direction. Hence downstream attenuation of peak discharge, such as that observed at Kverkfjöll (Carrivick et al., 2004a,b), is unlikely to be realised.

5.6. Hydraulic ponding

As discharge rises, channel or valley constrictions can cause upstream ponding, particularly where channel widths are reduced by the presence of mid-valley islands (Figs. 4 and 6, respectively). Ponding depends not just upon the magnitude of discharge, but also upon the rate of discharge rise. Ponding can be recognised in 2D-model output, either by regions of relatively deep flow depth or by regions of relatively low flow velocity (Figs. 4 and 6). Ponding in Kverkfjallarani could have been responsible for the deposition of valley-fill sediments (Carrivick et al., 2004a,b). If ponding leaves palaeostage indicators (PSIs), flood reconstructions using PSIs could erroneously calculate flood hydraulics, by assuming that those PSIs relate to channel flow. 1D models or other palaeohydrological methods do not consider hydraulic ponding because they cannot consider temporal variability.

5.7. Multi-directional flow and backwater areas

High-magnitude outburst floods that inundate valley-wide, back up tributary valleys, creating back water areas of ‘slack’ or low-energy flow (e.g. Baker et al., 1983; Carrivick et al., 2004b). Deposition in these areas occurs at peak stage and thus preserved slackwater sediments are useful PSIs (e.g. O’Connor, 1993; Smith, 1993; Benito, 1997). Whilst the phenomena of backwaters and slackwater deposition

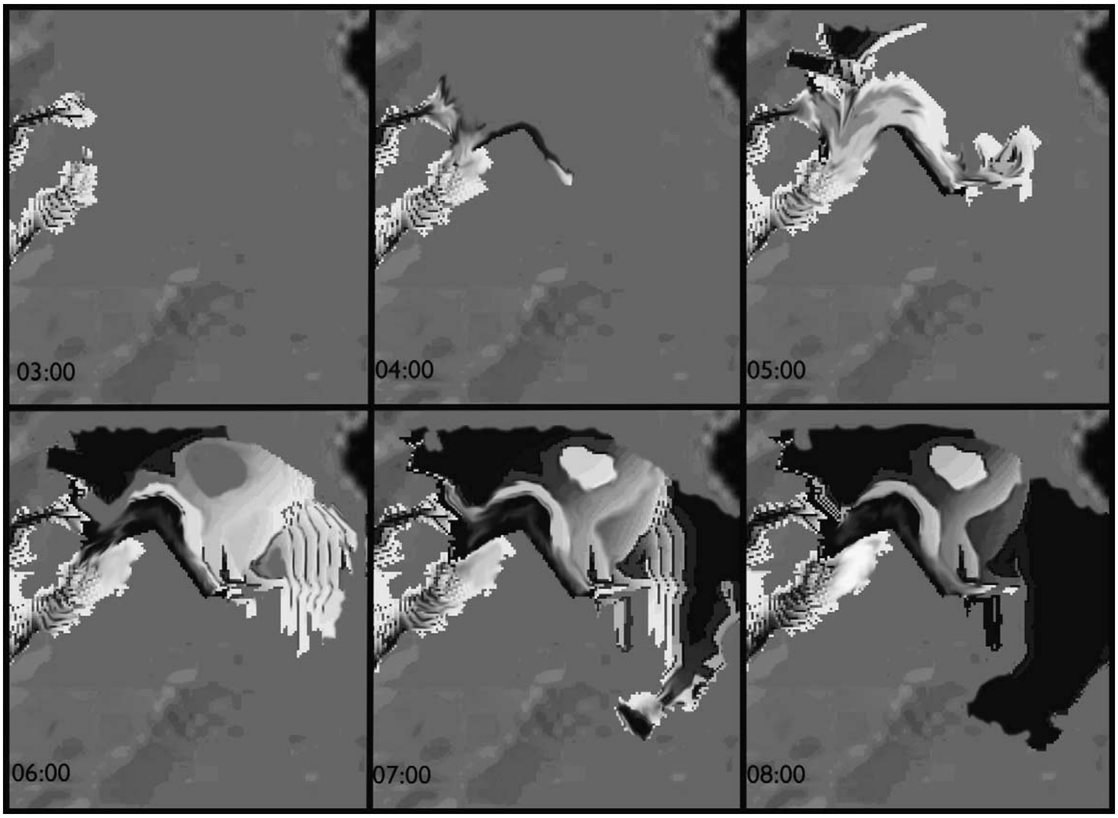

Fig. 8. Inundation of an outwash fan and a major backwater, at the 1.6 km wide mouth of Hraundalur, as a function of time since flood initiation. Note capability to produce flow in all directions, pulsing flow and braided sheet flow. Gray scale is flow depth; lightest $\sim<1$ m, darkest ~ 10 m.

due to high-magnitude outburst floods has been modelled in a flume (Kochel and Baker, 1988), hydraulic processes and mechanisms of deposition in slackwater areas, such as suspension drop-out and low-energy traction, remain unquantified.

Inundation of tributary valleys and other backwater sites depends on flow depth. Time series output from a 2D-model shows progressive inundation and deepening flows within backwater areas, whilst main channel flow hydraulics remain relatively stable (Fig. 8). Main channel hydraulics remain relatively stable, despite rising stage, because backwaters accommodate water mass and thus help dissipate flow energy and to attenuate main channel discharge. Backwaters also develop a surging flow (Fig. 8), which seems to be partly associated with flow around islands (Fig. 6), circulatory flow in semi-enclosed ponds (Figs. 4 and 6) and flow within backwaters (Fig. 8). An exclusion of backwaters from palaeohydrological reconstructions is therefore likely to produce inaccurate calculations of flow hydraulics. A 3D visualisation of the same area as in Fig. 8, is given in Fig. 9. Fig. 9 details flow into tributary

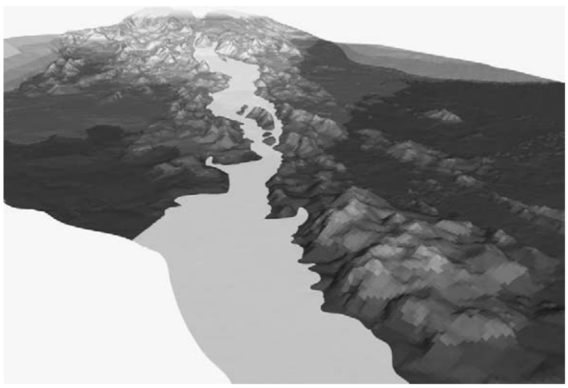

Fig. 9. Visualisation of backwater and outwash fan inundation. This is the mouth of Hraundalur and thus the same area as in Fig. 8.

valleys, flow expansions, lateral flow and back-flooding.

6. Conclusions, wider implications and future work

This paper has demonstrated the capability of a 2D hydrodynamic model for reconstructing characteristics of a high-magnitude outburst flood. These characteristics cannot be reconstructed by 1D models or other palaeohydrological techniques, due to the reliance on cross-sectional parameterisation and thus on PSIs, or on preserved suites of other geomorphological and sedimentological palaeoflood evidence. High-magnitude outburst flood characteristics include multiple points of flood initiation, simultaneous inundation of multiple channels, sheet (unconfined) flow, simultaneous channel and sheet flow, flow around islands, hydraulic jumps, multi-directional flow including backwater areas and hydraulic ponding.

2D-modelling of outburst floods clearly has the potential to revolutionise understanding of high-magnitude spatial and temporal hydraulics and high-magnitude flow phenomena, geomorphological and sedimentological processes, and hence rapid fluvial landscape change. This potential for new understanding is because of the now wide availability of high-resolution DEM data for large and often inaccessible areas and the availability of remotely-sensed data that can parameterise outburst flood sources, such as glacial lakes, for example. Additionally, hydraulic models and computing power are now sufficient to cope with large (5,000,000 grid cells) areas of inundation and large (100,000 $m^3 s^{-1}$) peak discharges.

It will not have escaped attention that computations presented in this paper assume a fixed channel bed, partly because modelling of sediment transport and hence of bed elevation change is premature (Cao and Carling, 2002) due to uncertainties in the physics of turbulence, sediment transport, and the interaction between them. Furthermore, model validation is much harder where sediment transport is reconstructed, for little short of a pre-flood DEM input and a post-flood DEM to validate model output. Nonetheless, further work should endeavour to address the issue of sediment transport in high-magnitude outburst floods, through 2D-modelling, because other palaeohydrological methods are unable to explicitly examine water-sediment interactions.

Iterative variations of peak discharge can constrain channel roughness, by independently checking 2D-model output against PSIs, for example (Denlinger et al., 2002). Future work should also make iterative variations of hydrograph input to 2D-models of high-magnitude floods, in order to examine the controls of hydrograph shape, peak flow duration, flood duration, flood volume and peak discharge, on spatial and temporal high-magnitude hydraulics.

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