

Stott, Eilidh (2019) *Rainfall-to-reach, modelling of braided morphodynamics*. MSc(R) thesis.

https://theses.gla.ac.uk/70942/

Copyright and moral rights for this work are retained by the author

A copy can be downloaded for personal non-commercial research or study, without prior permission or charge

This work cannot be reproduced or quoted extensively from without first obtaining permission in writing from the author

The content must not be changed in any way or sold commercially in any format or medium without the formal permission of the author

When referring to this work, full bibliographic details including the author, title, awarding institution and date of the thesis must be given

Enlighten: Theses <u>https://theses.gla.ac.uk/</u> research-enlighten@glasgow.ac.uk



Rainfall-to-reach, modelling of braided river morphodynamics.

Ms. Eilidh Stott (BSc. Hons)

Submitted in fulfilment of the requirements for the Degree of MSc. By Research in Earth Sciences

School of Geographical and Earth Science College of Science and Engineering University of Glasgow

Abstract

Communities today are increasingly vulnerable to fluvial flood events due to increased development on floodplains and climate change impacts. This prompts a need to improve flood risk management strategies. In particular, the forecasting of high flow events via hydrological modelling. In piedmont regions, where sediment supply can be substantial and variable depending on seasonal hydrology and perturbations such as landslides, geomorphic change can influence channel conveyance and thus flood risk. However, the modelling of sediment transport and its impacts on geomorphic change and flood risk is rarely incorporated into flood risk management. This thesis therefore aims to assess an end-to-end modelling framework to predict discharge, sediment transport and morphological change on a braided piedmont river. First, Structure from Motion photogrammetry was used to monitor topographic change along a braided reach of the River Feshie, Scotland, to demonstrate how morphological data can be obtained to assess numerical model performance. The coupled numerical modelling framework used for this analysis included a catchment scale rainfall runoff model (CLiDE), and a reach scale geomorphic change model (CAESAR-LISFLOOD). Model results show that the parameterised rainfall runoff model could appropriately predict base and storm flow discharges. The geomorphic change model could predict the location and magnitudes of change for one to two years of model runtime. However, after multiple floods the morphodynamic model outputs substantially degraded. As lateral erosion rapidly carved out river banks and deposited sediment in the channel, resulting in a topographically smooth reach. Despite the limitations of the reach scale model to maintain braiding over multiple events, this work demonstrates how an end-to-end modelling framework could be implemented to predict geomorphic change and contribute to updating the topography in flood risk mapping and forecasting models. Results also illustrate how repeat topographic surveys can be used as input and verification data for flood risk modelling which incorporates geomorphic change.

Table of Contents	3
Acknowledgments	8
Authors Declaration:	8
1 Introduction	9
1.1 Research Context	9
1.2 Aims	10
1.3 Objectives	10
1.4 Relevance of the Site	11
2 Literature Review	14
2.1 Structure-from-Motion topographic surveys: Ground Control Point	
density and distribution	14
2.1.1 Introduction	14
2.1.2 GCP density and distribution for SfM surveys	15
2.1.3 Structure-from-Motion in the context of surveying braided river	16
2 1 4 Structure from Motion Methodology	10
2.1.4 Structure-Hom-Motion Methodology	17
considering errors associated with both survey methods	19
2.2 Bathymetric Reconstruction	21
2.2.1 Spectral Depth Approach	21
2.2.2 Bathymetric Airborne Laser Scanning	22
2.2.3 Digital Photogrammetry	22
2.2.4 Application and assessment of refraction correction to SfM deriv	ed
topography	24
2.2.5 Coupling SfM and RTK-GNSS to Improve Bathymetric Reconstruct 25	ion
2.3 Rainfall Runoff and Morphodynamic Modelling	26
2.3.1 Cellular Automata	26
2.3.1.1 CA Modelling for Braided Rivers	28
2.3.2 CAESAR	29
2.3.2.1 Determining Bend Radius of Curvature	29
2.3.2.2 Model Application	31
2.3.2.3 Sediment Transport	32
2.3.2.4 Sediment Layers	32
2.3.2.5 Slope Processes	34
2.3.3 LISFLOOD-FP	34
2.3.3.1 Saint-Venant Equations	35
2.3.3.2 2D Flow	36
2.3.3.3 Courant-Freidrichs-Lewy Condition	37

4		
2.3.4 CAE	SAR-LISFLOOD	38
2.3.4.1	Combined Model Challenges	38
2.3.4.2	Steady vs Non-Steady Flow Models	39
2.3.5 CLil	DE	40
2.3.5.1	Hydrology	42
2.3.6 Hyc	Irology of Soil Types	44
2.3.7 Lan	duse	45
2.4 Assessi	ng numerical model performance	49
2.4.1 Geo	omorphic Change Detection	50
2.4.2 Unc	certainty and Error	50
2.4.2.1	Probabilistic Thresholding	52
3 Study Area		57
3.1 Catchn	nent	57
3.1.1 Cha	iracteristics	57
3.1.2 Hist	tory as a Braided River	57
3.1.3 Lan	duse	58
3.2 Reach.		59
4 Topographi	c Reconstruction and Change Analysis	62
4.1 Structu	re from Motion - Topographic Reconstruction	62
4.1.1 Intr	oduction	62
4.1.2 Met	hodology	62
4.1.2.1	Fieldwork component	62
4.1.2.2	Post Processing	63
Overvi	ew	63
Analysi	is	65
4.1.3 Res	ults	66
4.1.3.1	Z axis error	67
4.1.3.2	Category 1 GCP scenarios	67
4.1.3.3	Category 2 GCP Scenarios	69
4.1.3.4	Category 3 GCP Scenarios	70
4.1.4 Def	ining a Best Practice Survey Protocol	72
4.2 Bathyn	netric Reconstruction	76
4.2.1 Intr	oduction	76
4.2.2 Met	hodology	76
4.2.3 Res	ults	78
4.2.3.1	A Comparison to Observed Data	78
4.2.3.2	Errors	80
4.2.3.3	2018 Bathymetric Reconstruction	81
5 Rainfall Ru	noff, Catchment Scale Modelling	86

5.1 Int	troduction	5 86
5.1 Me	ethodology	86
5.2.1	Digital Elevation Model of the Feshie Catchment	
5.2.2	Hydrology Inputs	
5.2.2	2.1 Evapotranspiration	
5.2.2	2.2 Soil type and soil hydrological properties	
5.2.2	2.3 Landuse	
5.2.3	Initial Hvdrological Conditions	
5.2.3	3.1 Precipitation	90
5.2.3	3.2 Soil moisture	92
5.2.3	3.3 Groundwater	92
5.2.3	3.4 Hydrological Boundary Conditions	93
5.2.4	Calibration	93
5.2.4	4.1 Flow Separation	93
5.2.4	4.2 Approach to Calibration	95
5.3 Res	sults	96
5.3.1	Groundwater Spin Up (1)	96
5.3.2	Surface Water Spin Up (2)	97
5.3.3	Calibration 1 (3)	97
5.3.4	Calibration 2 (4)	98
5.3.5	Calibration 3 (5)	98
5.3.6	Calibration 4 (6)	99
5.3.7	Output Run (7)	99
5.3.8	Figures	101
6 Geomo	orphic Change Reach Scale Modelling	104
6.1 Int	roduction	104
6.2 Me	ethodology	104
6.2.1	Digital Elevation Models	104
6.2.2	Bulk Sediment Analysis	106
6.2.2	2.1 Introduction	106
6.2.2	2.2 Field Data Collection	106
6.2.2	2.3 Lab Analysis	112
6.2.2	2.4 Statistical Analysis	112
6.2.2	2.5 Results	113
6.2.2	2.6 CAESAR-LISFLOOD Input	115
6.3 Res		119
6.3.1	Spin up (1)	119
6.3.2	Calibration 1 (2)	123
6.3.3	Calibration 2 (3)	124

6		
6.3.4 Ca	libration 3 (4)	125
6.3.5 Ca	libration 4 (5)	127
6.3.5.1	Persistent Low Flow Issue	128
6.3.6 Ca	libration 5 (6)	129
6.3.7 Ou	tput Run/Calibration 6 (7)	131
6.4 Geomo	orphic Change Detection	
6.4.1 Ov	erview	
6.4.2 Res	sults	
7 Discussion		
7.1 Topog	raphic reconstruction and change analysis	
7.1.1 SfA	٨	
7.1.1.1	Target distribution	
7.1.1.2	Alternatives to use of Ground Control Points for geor 143	eferencing
7.1.2 Lin	nitations of Bathymetric Correction	
7.2 Rainfa	Ill Runoff and Geomorphic Change Modelling	
7.2.1 Cat	tchment Scale	
7.2.1.1	Resolution and Representation of Reality	
7.2.1.2	Environmental and Landscape Change	150
7.2.1.3	Calibration	151
7.2.1.4	Conclusions on Hydrological Model Performance	155
7.2.2 Rea	ach Scale	156
7.2.2.1	Resolution and Representation of Reality	156
DEM		156
Hydro	logical Processes	
Flow F	Routing Directions	
Hydro	logical Boundary	159
7.2.2.2	Assumption of Bedrock Depth	160
7.2.2.3	Implications of Sediment Recirculation	161
7.2.2.4	Bulk Sampling	162
Field <i>I</i>	Method	162
Statist	tical Analysis	163
7.2.2.5	Calibration	165
Spin U	lp	165
Repres	sentation of Reality - Calibration 1	166
A Com	nparison to Ziliani et al. (2013) - Calibration 3	166
7.2.2.6	Timescales of Reach Analysis	
7.2.3 Ge	omorphic Change Detection	167
7.2.3.1	Error Surfaces	167

		7
8	Conclusion	59
9	References	72
10	Appendix	31

8 Acknowledgments

I would like to extend thanks to the following parties for making the completion of this MSc thesis possible:

- For primary supervision Richard Williams, Martin Hurst, and Brian Barrett (University of Glasgow).
- For co. supervisor and data sharing Andy Barkwith (British Geological Society).
- For expertise on the River Feshie field site and for use of gauge data Andrew Black (University of Dundee).
- For access, expertise, and assistance during field campaigns Thomas MacDonnell, Davie McGibbon and the rest of the Glen Feshie Estate.
- For help with fieldwork Kenny Roberts, Trevor Hoey, Rob Ferguson, Mirijami Lantto, Owen Bowser, Walter Gilbert, Lauren Whittaker, Marine Prieur, Helen Williams, and Chloe Williams.
- For providing fieldwork accommodation (The Neuk) Alan and Elspeth McNicol (and McArthur family).
- For continued emotional and physical support Dorothy and Tim Stott, and all additional close friends and family.

Authors Declaration:

"I declare that, except where explicit reference is made to the contribution of others, that this dissertation is the result of my own work and has not been submitted for any other degree at the University of Glasgow or any other institution."

Printed Name: Eilidh Stott

Signature:

1 Introduction

1.1 <u>Research Context</u>

Many communities are increasingly vulnerable to fluvial flood events due to factors such as, increased development on floodplains and climate change (Lewin, 2013; Hall et al., 2016). This prompts a need to improve flood risk management. A key component of this process is the forecasting of high flows. This relies on a numerical modelling framework that uses data from rain gauges to predict discharge, and in-channel and floodplain topography to route river flow. In piedmont regions, modelled systems do not typically incorporate bedload transport and morphological change. Despite this, modelling geomorphology is an extremely important component of flood risk management. As changes in river bed levels, because of geomorphic change, can have a significant impact on flood risk. This is primarily linked to the river's capacity, which is decreased as a result of sedimentation. To prevent flooding rivers therefore needs to erode and migrate to account of this increase in sediment. However, as limited work has been carried out linking geomorphic change to flood risk, many current flood management techniques exacerbate this process by manually stabilising river banks; preventing erosion from taking place.

A study by Lane et al. (2007) used one- and two-dimensional flow models to simulate flood inundation in response to sediment delivery in a temperate upland gravel-bed river. They compared observed flood events, to equivalent events with the addition of a climate change parameter and found that the inundation area increased by around half for the latter case. This conclusion was especially significant in relation to piedmont regions, which can supply substantial amounts of coarse sediment to the river in relatively short timescales. This work therefore highlights the potentially severe consequences that can arise as a result of sedimentation coupled with climate change. However, the modelling technique used to quantify this process wasn't part of an end-to-end framework as it deals with discrete topographic data surveyed at fixed times. This means that limited robust conclusions can be made about how piedmont systems develop as a whole through time.

Currently, rainfall runoff models exist to predict surface and subsurface water input into river systems. Additionally, morphodynamic models exist which model sediment transport and morphological change. However, the two have yet to be 10

coupled to produce an encompassing rainfall to reach model, which models geomorphic change from a top down approach. This kind of model is likely to be most useful in piedmont regions where sediment supply to rivers is variable because of significant seasonal changes (i.e. snowmelt in the spring) coupled with an affinity to flash flood events (Raven et al., 2010).

1.2 <u>Aims</u>

This thesis aims to develop and assess an end-to-end modelling framework that will be used to predict discharge, sediment transport and morphological change on a braided piedmont river. The framework will include the following components: topographic data acquisition and modelling; rainfall-runoff modelling; and morphological change modelling

1.3 Objectives

The specific objectives are to:

(1) Generate high quality (cm precision) Digital Elevation Models (DEMs) of a c. 2 km long reach of the braided river Feshie by using a survey protocol developed through a series of field campaigns and subsequent post processing analysis.

(2) Calibrate a Catchment scale numerical model (CLiDE) to predict catchment hydrology and rainfall runoff using inputs from rain gauge data which was spatially interpreted and averaged for the River Feshie catchment.

(3) Calibrate a Reach scale numerical model (CAESAR LISFLOOD) to predict geomorphic change.

(4) Quantify morphological change of the c. 2 km braided reach using datasets obtained from 2000 to 2018.

(5) Validate and assess model performance using flow gauge data and DEMS of Difference (DoDs) for, respectively, hydrological and geomorphic change components.

1.4 <u>Relevance of the Site</u>

This research is focused on the River Feshie, which is situated in the Cairngorm National Park, Scotland. It is an ideal location to study braided river morphodynamic change, due to it being one of the most dynamic braided rivers in the UK. Performing such an analysis in this archetypal setting will mean that conclusions from this work will be representative of the global challenges associated with modelling river morphodynamics, and associated implications for flood risk management. In addition to this, the dimensions of the River Feshie, i.e. its spatial scale, are suitable for the type of high resolution topographic monitoring detailed in this thesis. Because of its characteristics, setting, and scale, the River Feshie has been the focus of multiple high-resolution topographic studies from 2000 to the present day (Brasington et al., 2003; Wheaton et al., 2010; Wheaton et al., 2013). The presence of data from these previous studies means that the River Feshie presents a unique opportunity to quantitatively assess an end-to-end morphodynamic modelling framework.

1.5 Thesis Structure

Due to the broad array of aims this thesis hopes to satisfy, this work has been broken up into seven discrete chapters. Each chapter aims to focus on a specific portion of the thesis, meanwhile following a progressive and sequential structure which represents the work flow included in an end-to-end modelling framework (as displayed in Figure 0). The thesis starts with a review of the current literature associated with the main aims and objectives stated above. This review initially covers work relating to topographic reconstruction of both dry and wet topography and focuses in on specific literature relating to the methods used in this analysis. This section relates to the first aim of this thesis, and first main block of analysis detailed in the subsequent text - the acquisition of a high quality DEM of the study site. The subsequent section of the literature review then details the current state of knowledge related to both rainfall runoff modelling, and geomorphic change modelling. Again, this text focuses on the models used in this analysis of the River Feshie while still including a broader context. A short final section of the literature review then details the assessment of model performance, providing additional information for discussion. This literature review is structured in such a way that it mirrors the analysis which is presented in the text: in a sequential manner.

The following chapter presents an overview of the study area, and comments on its appropriateness for this type of analysis. The subsequent chapter entitled Topographic Reconstruction and Change Analysis is where the first portion of original analysis is featured. This work is presented first as this details how information used in subsequent modelling was acquired. The work is broken down into two further subsections which separate how both wet and dry topography was obtained, as these were acquired via different methods. To avoid long and complex methodology and results sections later in the text these further subsections were included in this chapter as they link directly to the information being presented. The results sections include some narrative, however the main discussion of these results is included later in the thesis after the full topographic and modelling analysis and methodology is presented. This has been done so a full context is provided to the reader before each section of the work is discussed at length.

The following two chapters then detail the catchment scale modelling, and reach scale modelling respectively. Each chapter is broken into three primary subsections which include an introduction to how the model was set up, followed by a calibration section and then subsequent results and outputs. Again, this has been done to follow this step-by-step structure to build up the idea of an end-to-end framework. The catchment scale modelling results/outputs feed into the reach scale inputs. These have therefore been presented in the previous discrete chapter with an brief accompanying narrative with a larger discussion to follow after the results of the reach scale modelling are presented.

Following the reach scale geomorphic change modelling there is a final subsection which details a geomorphic change detection analysis. The outputs from the reach scale model are used in this work, however the analysis is not large enough to warrant its own chapter. Again, this section follow a similar pattern of presenting an introduction and results.

The following two chapters then represent the discussion and conclusions. These generally mirror the structure followed in the literature review, and as with the

presentation of the data/analysis. This structure provides an alternative way of presenting an end-to-end modelling framework in a sequential manner.



Figure 0. Flowchart detailing the structure of the thesis in relation to and end-to-end modelling framework.

2.1 <u>Structure-from-Motion topographic surveys: Ground Control Point density</u> <u>and distribution</u>

2.1.1 Introduction

To generate high quality topographic products for reach scale modelling, field based observation data must first be acquired. As per the first objective of this thesis, this was done via a coupling of Structure-from-Motion (SfM) photogrammetry with Real-Time Kinematic (RTK) Global Navigation Satellite Systems (GNSS) data. The methodology of how this was carried out will be detailed later in this thesis. However, this first chapter review current literature on SfM photogrammetry - a developing survey technique which is becoming increasingly popular in geomorphology. Subsequently, there is a pressing need to establish survey protocols for when this survey method is used to produce DEMs that are suitable for higher-dimensional hydro- and morphodynamic modelling (Javernick et al., 2015).

One appealing aspect of SfM is that it can have a broad range of applications across multiple scales. This is due to the fact it encompasses a wide array of imagery and imaging geometries. However, the negative impact of this is knowing what image sets to acquire across this broad array. A need therefore remains to produce a structured survey protocol that guides the user during data acquisition and post processing. Additionally, this flexibility has the potential to result in variations in data quality which can be seen when comparing SfM datasets to produce DEMs of Difference (DoDs), or when analysing data within individual surveys (James, Robson, Smith, 2017). This variation is often poorly quantified in the literature; which can be lead to ambiguity in results. It is therefore vital that the quality of SfM data, and the errors which may be associated with it, are reported throughout SfM workflows.

This section of the literature review aims to provide an overview of SfM photogrammetry, with a focus on acquiring data to monitor fluvial topography. It will also provide detail on the type and magnitude of errors associated with this survey method, and pin point where errors can arise and discuss ways to reduce these.

2.1.2 GCP density and distribution for SfM surveys

Advances in the field of photogrammetry have resulted in the development of an exceedingly automated photogrammetric technique known as Structure-from-Motion. Unlike traditional softcopy photogrammetry, SfM can solve camera pose and scene geometry automatically and simultaneously without previously knowing the 3D locations of these parameters. This method, when coupled with Multi-View Stereo (MVS), can generate dense 3D point clouds for topographic surveys that may range in longitudinal extents of 10s kilometres in scale (James and Robson, 2012; Javernick et al., 2015). Dietrich et al. (2016) demonstrated the applicability of this method over a 32-km length of the John Day River in Oregon, USA. This ability to span scales means reasonably standardised measurements of topography can be deduced over different spatial and temporal resolutions (Smith and Vericat, 2015).

This survey approach has been popularised by an increasing number of open source cloud-based processing engines. Such software is user-friendly, and the comparatively low costs involved in SfM-MVS technologies make it an appealing survey technique to low-budget researchers (Javernick et al., 2015). SfM image data is often collected via unmanned automated vehicles (UAVs). This survey mechanism allows data to be collected frequently and to a reasonably high resolution and accuracy. Advances in UAV hardware and firmware over the past decade have led to further enhancements in image collection capabilities through improved stability and flight time (Woodget et al., 2017). These factors have led to an increased quality and quantity of photographs being captured by UAVs, but there is still scope for further improvement regarding repeatability of measurements (James et al., 2017).

SfM works by using a highly redundant bundle adjustment which matches features in multiple overlapping, offset images (Westoby et al., 2012). This data, when initially collected, lacks scale and orientation parameters; as 3-D point clouds are generated in a relative 'image-space' coordinate system. These point clouds must then be transformed to an absolute coordinate system via a 3-D similarity transformation based on a small number of known ground-control points (GCPs) 16

with known object-space coordinates. Metric data can then be extracted from the newly created point clouds (Westoby et al., 2012).

2.1.3 Structure-from-Motion in the context of surveying braided river topography

The setting of this research - the braided River Feshie - poses a unique challenge regarding the accurate representation of topography in comparison to many other non-braided river landscapes (Williams et al., 2016). This is generally associated with the subtle and intricate relief typical of this type of river system which can be difficult to model accurately (Williams, et al., 2013; Javernick et al., 2014; Javernick et al., 2015). Due to a comparatively large spatial extent of braided rivers, coupled with relatively small vertical relief, there is a constant trade-off between decreasing image scale, which will increase spatial coverage but reduce theoretical precision, or the corollary which will increase precision but will also increase the number of images which need to be obtained. This trade-off means there is a minimum point precision acceptable for an effective surface representation. This value will in turn determine a study's photographic scale and thus the subsequent DEM resolution (Lane 2004; Westway et al., 2001). A second challenge that can arise during post-processing of image data is SfM's ability to deal with vegetation. Mismatching of points can result in large vertical errors relative to the typical river bed relief. These anomalous points must be identified and removed, so as to not lead to further inaccuracies within the dataset. A third challenge relating to the use of SfM-MVS in the context of braided rivers, is the ability to accurately model both exposed and submerged zones. When waters are sufficiently clear, a two-media refraction correction can be used with digital photogrammetry to predict bathymetry (Woodget et al., 2015). However, if waters are turbid or contain a high proportion of silt or mud, additional methods for measuring channel bathymetry may be required, such as an echo-sounding or RTK-GNSS survey. This issue is discussed in detail in the Bathymetric Reconstruction portion of this thesis (Section 2.2).

Despite these challenges, 3D datasets, such as those produced from SfM software, have the potential to transform our understanding of braided river systems, by enabling better predictions of morphological change and sediment transport rates through providing reasonably boundary conditions and parameters for numerical modelling of channel flow.

2.1.4 Structure-from-Motion Methodology

GCPs can be derived after a field survey has taken place, by selecting clearly visible features (e.g. road markings) within the survey extent and obtaining their coordinates by undertaking a global navigation satellite system (GNSS) survey. However, in settings without clearly identifiable features, it is common practice to deploy physical targets prior to a UAV survey. These targets should be clearly defined and highly contrasting to the surrounding landscape (Westoby et al., 2012). To georeference a point cloud that has been produced using SfM, the 3D position of three or more GCPs must be known. Once this point cloud has been transformed using these GCPs, the location of other GCPs can easily be identified through the image set. Assuming all GCPs are symmetric and identical, the image texture from one GCP can be used to refine the positions of others by using a 3D patch-based cross-correlation procedure. For this procedure to work accurately target size should be sufficiently large that they exceed the intended patch size (James et al., 2017). The number and density of GCPs deployed will ultimately depend on the specifications of any one survey. Therefore, these factors must be considered as part of the initial survey design, while at the same time considering the required internal precision and overall repeatability of the survey, as well as the required spatial resolution of associated DEMs and orthomosaic products (James et al., 2017).

GCPs can have a varying influence on the final point cloud and subsequent DEM depending on the network's strength and quality. Photogrammetric best practice states that control measurements should be distributed across and surrounding the area of interest (Luhmann et al., 2006). Noisy or weak image networks, such as those taken by cameras with high levels of lens distortion, will require greater control input i.e. more GCPs. Past studies have therefore generally decided to err on the side of caution and deployed more GCPs than required to account for the potential of poor quality image networks. The exact number of GCPs needed will be site specific and depend on factors such as accessibility, terrain type, survey requirements, costing, and labour resources.

James et al. (2017) have posited the best way to layout GCPs is via 'working from the whole to the part' method, whereby GCP coverage is designed to extend to the edges of the survey area. This provides an overarching spatial framework which DEM data can be interpolated within, as opposed to extrapolating beyond. To further improve output quality and ensure DEM accuracy, control data should be incorporated within the image processing stage. This might involve weighting GCPs relative contribution appropriately while processing, and ensuring that any outlier points are identified and eliminated. It must also be noted that if uncertainty within the GCP measurements is not correctly accounted for, then including this data can introduce errors into the subsequent DEMs produced.

Accurate georeferencing is central to achieving data with suitable repeatability for detecting change. Considerable effort must therefore be taken when deploying GCPs. In general terms, GCP precision should exceed the required survey precision. This manual effort coupled with the expense of additional survey equipment to ensure a high level of precision has the potential to offset the costeffectiveness of UAV and SfM-MVS processing (James, Robson, and Smith, 2017). However, if the purpose of GCPs is only for scale and orientation, i.e. not to help define shape through bundle adjustment, a lower precision can be used if GCP redundancy is high enough (James et al., 2017).

If GCP data is not precise, this can subsequently influence the derived 3-D topographic point coordinate precision values. There are two types of precision associated with GCPs: (1) point locations determined by an external coordinate system which can be degraded by poor control measurement precision, or (2) relative distances between points within the survey (i.e. the 'internal' precision) which can be altered depending on the quality of the tie points. Through separating these two components of precision, one can gain an insight into the relative contributions of control measurements and tie points. Subsequent conclusions can then be drawn on the importance of control measurements in influencing the shape of a survey, as well as its overall georeferencing (James, Robson, and Smith, 2017).

Despite various guidelines on how survey data should be obtained, a gap in the knowledge remains as to what the optimum GCP density and distribution is. Limited research has been done trialling various configurations, and analysing subsequent error statistics. Generally, previous studies have given limited weight

18

to this parameter and its potential influence on the subsequent DEMs produced. Most have set out an arbitrary number of GCPs based directly on labour and monetary resources available, while the distribution parameter has generally been controlled by accessibility and terrain. This investigation therefore aims to provide a comprehensive analysis of the influence of GCP density and distribution on error statistics, with an overall objective of producing a survey protocol for topographic surveys of braided rivers using SfM, backed up with quantitative results.





2.1.5 Comparing Structure-from-Motion to Terrestrial Laser Scanning, and considering errors associated with both survey methods.

To assess the vertical error associated with different GCP layouts, errors need to be quantified. These can be calculated either from independent check points or an independent 3-D survey. Alongside this, a detailed investigation into the different causes of errors produced from SfM-MVS software is needed to better quantify and mitigate errors associated with manual identification of GCPs and the impact this has on the accuracy of the transformation matrix applied to the dataset. As SfM-MVS contains several components, it can be hard to determine where certain errors arise. Examples of sources of errors in previous studies include, errors due to: camera model; number and resolution of images; processing software; and the distribution and quality of GCPs used for georeferencing (James and Robson 2012; James and Robson 2014).

Generally, SfM-MVS software tends to have a relatively simple and easy to navigate user interface. This, however, means that a lot of the processing parameters are hidden from the user. Such a "black box" approach means there is a lack of information on parameterisation which discourages users from fully understanding underlying sources of error. Standard software output reports therefore do not generally provide the detailed quality assessment diagnostics necessary for rigorous photogrammetric analysis. James et al. (2017) have stated that through illustrating a DEM's sensitivity to the values used for processing settings, one can highlight the additional information that should be provided alongside surveys to increase confidence in results.

Considering the above, an argument remains in the literature against using SfM as a survey method, due to its magnitude of errors, in comparison to the method of Terrestrial Laser Scanning (TLS). However, a study by Westoby et al. (2012), who performed a SfM-MVS analysis, produced a dense point cloud of 11.3 x 10⁶ points, which was shown to be comparable to the survey density of terrestrial laser scan (TLS) data that contained 11.7 x 10⁶ survey point over the same area. This resolution was sufficiently precise than it revealed bedrock structures. The study further compared the two survey techniques by creating a DEM of difference (DoD) whereby they subtracted the SfM elevation model from that obtained by TLS. Results revealed that 94% of the differences were in the range of -1.0 to 1.0 metres, with 86% between -0.5 and 0.5 metres. It was subsequently hypothesised that the largest DoD values corresponded to areas of steeply sloping relief which were hard to access during the TLS survey, or areas of dense shrub and bush cover.

20

2.2 Bathymetric Reconstruction

Accurate elevation values for submerged topography are required for a wide array of scientific applications which include: geomorphic change detection (Wheaton et al., 2010; Bangen et al 2014), hydraulic modelling (Maddock, 1999), river restoration (Hicks, 2012), and sediment budgeting (Marcus et al., 2012). In any fluvial survey where a digital surface model must be produced, the researcher must make a choice on how to survey both the dry and wetted areas. These are generally obtained via different methods, which can be challenging when considering the cost and time constraints of obtaining both datasets, alongside any subsequent post processing. A key specification of many wetted channel surveys is repeatability - a consistent requirement of any investigation looking to demonstrate change over time. Survey methods that allow repeatability in a practical, and cost-efficient manner are generally preferable. In addition to this, datasets should be of an appropriate resolution for the scientific application and be spatially continuous. This is as opposed to widely distributed point or line sampling of bathymetry; methods that are not suitable for provide inputs for creating spatially continuous DEMs of both wet and dry areas (Carbonneau et al., 2012).

2.2.1 Spectral Depth Approach

A widely applied technique for modelling depth of gravel-bed rivers is the spectral depth approach (Legleiter et al., 2009). This is achieved by using an empirical correlation between depth data, obtained from accurately positioned field measurements, and the spectral properties of corresponding images. Imagery from standard digital cameras enables empirical relations to be developed between Red, Green and/or Blue (RGB) brightness values but higher-end multispectral cameras enable the development of empirical relations with narrower radiation bands. RGB values are extracted from imagery at specific point locations where depth is known and a logarithmic function is then typically applied to the data, commonly ln(G/R). One can then derive a linear regression between the logged spectral properties and the point data. Following this, the resulting equation can be applied to the spatially continuous imagery, to estimate water depth values on a spatially continuous scale (Shintani, 2016).

A study by Lejot et al. (2007) found this approach to produce mean errors in the order of 0.1 m, and outputs at a spatial resolution of 0.05 m while surveying the bathymetry of the Ain and Drome Rivers in France. Despite these results, various studies have found that this approach can result in errors and uncertainties with regard to factors such as: scene illumination; substrate; turbidity; and water surface roughness (Westaway et al., 2003; Lejot et al., 2007; Bergeron and Carbonneau 2012). In addition to this, this survey method can only produce accurate results to the maximum depth where a channel bed is illuminated which is often limited to approximately 1 metre, and the initial data collection to obtain depth measurements can be both labour intensive and time consuming (Woodget et al., 2015).

2.2.2 Bathymetric Airborne Laser Scanning

The traditional method of laser scanning with near-infrared light is far from an ideal method to try and survey river bathymetry, as this form of light is easily absorbed by water (Lane and Carbonneau 2007). To combat this issue, blue-green laser scanners were put forward as an alternative remote sensing method to measure submerged topography. A primary benefit of this survey technique is that it is influenced less by turbidity in comparison to the other remote survey techniques discussed. Additionally, since this method uses active remote sensing, it is capable of reliably surveying depths greater than 1 m; a limitation of many other passive survey methods (Kinzel et al., 2013). Despite the positive benefits of this technology, its use has been restricted in fluvial geomorphological research due to its relatively high cost, the coarse spatial resolution of data acquired from the first generation of bathymetric LiDAR systems, and relatively high reported errors where the technique has been test in shallow water gravel-bed rivers (Hicks, 2012; Kinzel et al., 2013). However, newer bathymetric LiDAR currently being tested are showing more promising results in terms of higher point densities (Lague et al., 2016) but this technology is still being developed and there is limited published work on these newer, more accurate methods.

2.2.3 Digital Photogrammetry

As discussed in the chapter above, SfM photogrammetry can reproduce dry topographic values to an acceptable accuracy. To reconstruct submerged topography a refraction correction model is required. Such reconstruction is more challenging that the reconstruction of dry topography because factors such as

22

turbidity and water depth can hinder accurate measurements (Westaway et al., 2001). When using this method, the effects of light refraction at the air-water interface must be modelled, the geometry of which is described by Snell's Law (Equation 1; Figure 2).

$$\frac{\sin r}{\sin i} = \frac{h}{h_A} = \frac{n_1}{n_2} \qquad Equation \ 1$$

Where r (deg) is the angle of the refracted light, i (deg) is the angle of the incident, h (L) is the true water depth, h_A (L) is the apparent water depth, n_1 is the refractive index of water, and n_2 is the refractive index of air - both dimensionless values. For water with minimal turbidity, n_1 has a value of c.1.34. However, this value can vary by up to 1% depending on the temperature and salinity conditions (Westaway et al., 2001).

As is demonstrated by Equation 1, there are two values for depth - true and apparent. This is due to the refraction that occurs across the air-water interface that leads to an overestimation of the actual bed elevation - displayed in Figure 2 (Westaway et al., 2001; Woodget et al., 2015). This can therefore be corrected for via Equation 2, where apparent depth can be transformed into true depth.

$$h = n_1 \times h_A$$
 Equation 2

Westaway et al. (2001) showed that by performing this correction, mean error results for depths less than 0.4 metres were comparable with that of exposed terrain. However, beyond this depth error values were significantly worse than those obtained on dry topography. In addition to this, the study found that the magnitude of errors increased with depth. Their work concluded that the correction could be of use when considering low turbidity, shallow waters but additional work would have to be performed on the refraction if this weren't the case. One method to improve the accuracy of elevation data in deeper more turbid waters, is to compare reconstructed bathymetric values with observed values obtained from a RTK-GNSS point survey - the method used for this analysis. A methodology detailing how this can be done can be found in Section 4.2.



Figure 2. Schematic diagram displaying apparent water depth verses actual water depth as per aerial imaging observations (Woodget et al., 2015).

2.2.4 Application and assessment of refraction correction to SfM derived topography

The above approach to obtain corrected depth values was used by Woodget et al. (2015) to correct a DEM that was generated using SfM photogrammetry. Woodget et al. demonstrated that errors on submerged data points were reduced via the application of a refraction correction. Initial errors in the survey ranged from 0.016 m to 0.089 m in submerged areas, and then 0.008 m to 0.053 m, respectively, when the correction was applied - reducing errors by c. 50%. In addition to this, the study shows comparable, but marginally lower levels of accuracy (0.02 m to 0.09 m) and precision (0.06 m to 0.09 m) in wet areas compared to errors on dry topography.

Woodget et al's results showed that refraction correction tended to over predict elevation values, and that this trend became more pronounced as the depth of water increased. Thus, the error in modelled DEM values was depth dependant. This overestimation of depth values is thought to arise through a combination of the effects of refraction, coupled with the photogrammetric process fixing matches at points with the water column, as opposed to on the channel bed (Westaway et al., 2000, 2001; Feurer et al., 2008; Woodget et al., 2015). Their study focused on two case study locations - the Coledale Beck and the River Arrow. The results of the refraction correction differed depending on site. For the latter, mean error values were significantly improved as a result of the correction. However, this was not the case for the former. The Coledale Beck site showed lower initial error values prior to the correction, with a mean error of 0.017 m - a value comparable to that of the dry topography. Westway et al. (2001) noted that this method of inferring a true water depth, is unlikely to be valid at water depths

this method of inferring a true water depth, is unlikely to be valid at water depths less than 0.2 m, because at this depth the effect of refraction is deemed negligible and therefore there is no need to correct for it. The Coledale Beck site therefore reinforces this statement as it is a shallow stream were 83% of depth values are less than or equal to 0.2 m (Woodget et al., 2015). This reiterates the limitations of this procedure and highlights its dependence on channel variables such as depth.

Woodget's method detailed above only considers how refraction influences depth points (i.e. the z dimension). However, the process of refraction is known also have a significant influence in both horizon dimensions. Meaning the refraction observed at the water surface would change all three apparent coordinates of the returned point. Each dimension will be distorted differently, so there is no one values which can be used as a generic correction for both the horizontal and vertical planes. The correction one should perform is also dependant on the type of data collected. A novel technique developed by Dietrich (2017) detailed a correction method which could be used in conjunction with a SfM analysis, and produced mean errors of c. 0.02%. This method used a multi camera based refraction correction method which incorporated an iterative approach to solve a series of refraction correction equations for every point/camera combination in a SfM produced point cloud. Despite the positive results displayed in this work, this method is still in its infancy and was only tested in a very small study area. Further work is needed to increase the confidence of these results before this method is to be used on a wider scale.

2.2.5 Coupling SfM and RTK-GNSS to Improve Bathymetric Reconstruction

A significant source of error in calculating channel bed levels using either a spectral depth or refraction correction approach is in the mapping of water surface elevation from remotely sensed topographic point cloud data. To reduce

the uncertainty in water surface elevations, when completing the field based component of a study, a channel's edge can be surveyed with an RTK-GNSS point survey. Additionally, RTK-GNSS surveys contain known error values related to individual points. This method can be preferential in terms of accuracy and precision in comparison to WSE data extracted from, for example, an SfM derived DEM. Despite the fact overall error values of a SfM reconstruction and a RTK-GNSS survey might be comparable. When considering the precise elevation values of channel banks, an observed RTK-GNSS survey is thought to provide a higher level of confidence in results. This is as SfM has been observed to provide higher error values associated with areas where there is a step increase in elevation over a small horizontal distance (Clapuyt et al., 2016). The method by which a RTK-GNSS water edge survey is incorporated in a bathymetric reconstruction is detailed later in the text.

2.3 Rainfall Runoff and Morphodynamic Modelling

As per the objectives of this thesis, topographic datasets were obtained to be used as inputs into both catchment and reach scale morphodynamic models. This chapter discusses the development of the two different models used for this research, and how they link to one and other. As well as outlining how such hydromorphic models fit within the broader field of computational modelling.

2.3.1 Cellular Automata

Both models used in the analysis of the River Feshie are termed Cellular Automata (CA) models. The initial development of CA was to investigate self-replication, and provide an alternative method to solving the governing equations of fluid dynamics - the continuity, momentum and energy equations. These denote mathematical statements, which describe the three fundamental physical principles upon which all fluid dynamics is based (Anderson, 1992). CA was developed with the hope that it would allow a fast, exact numerical simulation of a physical system (Von Neumann 1951). This was done to combat the fact that, to simulate a distributed dynamic system one often has to solve complex differential equations which require large computational resources (Toffoli, 1984). Since their use began, CA have been applied to a large array of physical research areas

26

centred around natural systems and fluid dynamics (eg., D'ambrosio et al., 2001; Coulthard et al., 2002; Barkwith et al., 2015).

A cellular automaton consists of a regular grid of cells, each in one of a finite number of states, such as on and off. For each cell, a set of adjacent cells are called its neighbourhood, and are defined relative to the specified cell. An initial state, where time (t) equals zero, is selected by assigning a state for each cell by the user. For example, in relation to a channel system, a cell could be labelled either wet or dry. A new generation is then subsequently created (t + 1), in accordance to some fixed rule - generally, a mathematical function that determines the new state of each cell in terms of the current state of the cell and the states of the cells in its neighbourhood area (Toffoli and Margolus, 1987). In the example of water flow, this could represent the flow of water in and out of a cell from and to its neighbours. Each cell can pass and gather information regarding neighbouring cells and subsequently modify its contents based on a transfer function. Information that can be exchanged between cells might be variables such as water depth, surface gradient, and water surface gradient.

Interaction with neighbouring cells for two-dimensional cases generally consists of one of two methods - displayed in Figure 3. The first being a Moore-type method, where all surrounding cells interact with the central node (Moore, 1962). The second is the von Neumann (Manhattan) method where interaction is solely with adjacent cells (Von Neumann and Burks, 1996). The rule for updating the state of cells is, in a majority of cases, the same for every cell and does not change with time. It is generally applied to the whole grid simultaneously, however some exceptions do exist such as the stochastic cellular automaton and asynchronous cellular automaton (Schiff, 2011).



Figure 3. Depiction of CA neighbourhoods; the von Neumann neighbourhood (left) considers cells in the cardinal directions, and the Moore neighbourhood (right) allows interactions with all neighbouring cells (Barkwith et al., 2015).

2.3.1.1 CA Modelling for Braided Rivers

CA models have shown great promise in their ability to simulate braided river dynamics (e.g., Thomas and Nicholas, 2002) and landscape evolution (e.g., Coulthard and Macklin, 2003). Such models simplify the complex equations used to calculate flow, to reduce computation time and allow sediment transport and morphological change to be modelled in a simple and concise manner. However, in earlier CA models, the terms for momentum, and for describing secondary circulation were not accounted for in this simplified approach. An additional issue with earlier studies was that cellular discretization, i.e. the concept that each cell has only knowledge of itself and its immediate neighbours, makes it difficult to determine where the cell is in exact relation to the channel's course and, in turn, the momentum and direction of water entering the cell. Data regarding the exact location of a cell in respect to channel banks is therefore necessary to replicate meandering processes accurately (Coulthard et al., 2002).

The lack of these concepts in earlier CA models prevented meandering from being incorporated - as it is determined by encompassing channel planform and circulation patterns. Earlier models only contained knowledge of the local flow parameters in point values of depth and velocity, meaning that they failed to replicate multiple braided river channels. River meandering had been successfully modelled using vector based methods (Howard, 1992; 1996), but these could not simulate multiple or braided channels. Conversely, cellular braided river models

fail to replicate meandering. This consequentially hindered the development of an encompassing landscape evolution model which can replicate a bread of fluvial systems until recently.

2.3.2 CAESAR

2.3.2.1 Determining Bend Radius of Curvature

To solve this modelling gap, Coulthard and Van de Wiel (2006) therefore created the Cellular Automaton Evolutionary Slope and River model also known as CAESAR. This model incorporates a novel technique for determining bend radius of curvature. This is done via a four-stage methodology that operates over a grid where water depths are known, and cells are labelled either wet or dry.

The first of these stages is to determine which cells should be denoted 'edge', by observing if a cell has any neighbouring wet cells. Following this, a nine-cell filter is placed on the grid, and if there is an 'edge' cell at the centre of the filter, the wet and dry cells contained within it are summed. A ratio between wet and dry cells is then obtained by subtracting the wet from the dry to give a local expression of radius curvature. The sign of this ratio indicates whether it is outside (positive) or inside (negative) the channel's banks. This method provides an approximate measure of curvature, however there can be errors in this method because of the discretization of cells into a grid. A smoothing filter is therefore applied to the grid to reduce any errors associated with this rough gridded format. The filter averages curvature between adjacent cells, to create moving average values of the balance between wet and dry for edge cells and can be denoted R_{ca}. This methodology is displayed in Figure 4 below.



Figure 4. An example of a bend in a river channel, where blue cells represent wetted areas and orange represents edge cells. Sections A, B and C display zoom ins of the inside, outside and straight sections of the channel respectively (Coulthard and Van de Wiel, 2006).

This local curvature is then used to drive lateral erosion and meandering. The technique by which to modelling former can be easily implemented via the method by Lancaster and Bras (2002) which uses the concept of topographic steering to drive lateral erosion. Here, erosion is determined by the angle at which water impacts a bank, and the slope of the channel. However, meandering, via channel migration, requires the system to model the process of deposition as well - a harder parameter to quantify.

Coulthard and Van de Wiel (2006) tested two approaches to replicate deposition, and the subsequent redistribution of sediment, however neither managed to produce optimum results. The first method lacked the inherent processes that operate to produce deposition of sediment, as well as producing unrealistic values for sediment mass balance. The second coped better at replicating the hydrodynamics processes of deposition, and was shown to lead to the development of appropriate geomorphic features such as point bars. This deposition

30

subsequently allows the internal hydraulics of the model the ability to change the position of the thalweg, and in turn induce lateral erosion on the opposite bank. If this component did not exist, lateral erosion could only exist when the channel contained initial curvature.

One must note that the model developed still contains some limitations such as: it cannot accurately represent meander cut offs; sediment deposition processes along the inside bend of meanders lack robustness; the model has had no form of sensitivity testing (from internal and external factors); and the model still requires further calibration and validation of migration rates and planform sinuosity. Despite these limitations, the model created has greatly surpassed its predecessors in its ability to model a breadth of fluvial systems.

2.3.2.2 Model Application

CAESAR is a CA landscape evolution model that simulates geomorphic change due to the movement of water and hillslope diffusion across a regular grid cell network. When water moves across a cell it can change the surface elevation by entraining or depositing sediment, from fluvial and slope processes.

CAESAR requires elevation data in a gridded format for either a catchment or reach of a chosen river. Rainfall and river discharge values can then be input over this gridded surface to stimulate changes in the landscape. CAESAR includes an inherent hydrological system which is coded as per TOPMODEL - a physically based distributed watershed model that allows hydrological fluxes of ground and surface water within a defined area (Beven and Kirkby, 1979). This is done via defining the movement of the water table, and therefore where the ground is sufficiently saturated to produce overland flow. Once overland flow is present, the model then routes surface water via a multiple flow routing algorithm which creates values of flow depth over a cellular grid (Van de Wiel et al., 2007). A value for shear stress is then incorporated into the model when both slope and flow depth are considered together in a cell. This shear stress can then be related to erosion and sediment transport values. In CAESAR bedload transport is calculated using either the Einstein (1950) method, or the Wilcock and Crowe (2003) method; both of which relate to sediment transport equations. (Coulthard et al., 2002; Coulthard 2005; Van De Wiel et al., 2007; Pasculli et al., 2015).

32 2.3.2.3 Sediment Transport

A commonly used bedload transport formula is the method developed by Wilcock and Crowe (2003). This method incorporates variation in grainsize, which the user supplies to the model. Grainsize information can be denoted as either bedload or suspended load, and the subsequent geomorphic processes modelled will reflect as such. For example, when considering deposition, sediment treated as bedload will be moved directly from cell to cell, and only neighbouring cells with a lower bed elevation are considered - Figure 5 (a). Whereas sediment which is classed as suspended load will be deposited dependent on fall velocities and sediment concentration in the water column. Therefore, meaning that it can be deposited in any neighbouring downstream cell where the bed elevation is less than the water elevation in the cell being considered - Figure 5 (b).



Figure 5. Sediment routing directions for both bed load (a), and suspended sediment load (b) (Van der Weil et al., 2007).

2.3.2.4 Sediment Layers

CAESAR allows for spatial variability of the sediment size distribution, where grainsize data can be variable in both the vertical and horizontal axes. This is conveyed with an 'active layer' of sediment, i.e. the exposed part of the regolith,

that can be user defined dependant on field data collection. Figure 6 displays an example of the sediment layers which can be present within CAESAR, where strata represent multiple buried layers of sediment. Up to twenty strata layers can be included in a model. The strata layers are underlain by a base layer that represents the lower part of the buried regolith, and a bedrock layer that cannot be eroded as per the model parameters.



Figure 6. Example of sediment layers which can be incorporated into CAESAR (Van de Wiel et al., 2007).

Erosional processes occurring on the surface can remove sediment and cause the thickness of the active layer to decrease. The model contains a threshold whereby if the thickness is less than 25% of the thickness of the strata, then the upper stratum is incorporated in the active layer to roughly maintain its thickness. The opposite is true for deposition, where if the active layer becomes 150% of the strata, a new stratum is created to avoid the layer becoming too thick (Van der Weil et al., 2007) - this process is displayed in Figure 7.



Figure 7. Schematic representation of (a) erosion and (b) deposition, of the active layer. Where, n denotes the number of burried layers at stage 1, and n* and n" denote the adjusted number of layers after a process as occurred (Van der Weil et al., 2007).

2.3.2.5 Slope Processes

Slope processes are also included in the model's base code, and allow material to be transported from the slopes into the main river channel. These processes occur on a variety of scales - from large slope failures, to small bank collapses. Equations relate failure to the critical slope threshold and allow for processes such as soil creep and land sliding to occur. A soil erosion rate is also incorporated into the model, which is a parameter of slope length and angle.

2.3.3 LISFLOOD-FP

As with CAESAR, LISFLOOD-FP was designed to fill the gap in CA modelling that existed, whereby previous CA models used overly simplified equations that could not accurately represent the hydrodynamic processes that simulate meander migration. LISFLOOD-FP is a two-dimensional hydrodynamic model designed to simulate floodplain inundation in a computationally efficient manner over complex topography. The model predicts water depths in a grid cell format, repeatable for a given time step, and hence can simulate the dynamic propagation of flood waves over fluvial, coastal and estuarine floodplains. In the case of fluvial flooding, as well as producing outputs for depth, water surface elevation, and velocity, it also outputs predicted stage and discharge hydrographs at the outlet of the reach and other specified locations (Bates et al., 2013).

2.3.3.1 Saint-Venant Equations

The model solves a reduced form of the shallow water equations using a simple numerical scheme. This has allowed for an increase in computational efficiency in comparison to previous hydrodynamic models. These equations, also known as the Saint-Venant equations, model fast inundation in 2D. Previous equations describing these processes lacked the terms for inertia, which meant that modelled system had smaller less stable time steps; resulting in slower runs (Hunter et al., 2008). The inclusion of the effects of inertia aimed to solve this problem, as well as introducing a potentially important component of flow physics in particular environmental settings (Bates et al., 2010). To derive such shallow water equations one must first start with their one dimensional, quasi-linear form:

$$\frac{\partial Q}{\partial t} + \frac{\partial}{\partial x} \begin{pmatrix} Q^2 \\ A \end{pmatrix} + \frac{gA\partial(h+z)}{\partial x} + \frac{gn^2Q^2}{R^{4/3}A} = 0 \qquad \text{Equation 3}$$

Where *Q* is discharge ($L^3 T^{-1}$), A is the cross sectional area of flow (L^2), *Z* is the elevation of the bed (L), *R* is the hydraulic radius (L), *g* is gravity ($L T^{-2}$), *h* is the water free surface height (L), and *n* is the Manning's coefficient of friction ($L^{-1/3}$ T). Each term within Equation 3 represents a specific component of flow and they can be denoted as, acceleration, advection, water slope and friction slope respectively.

The following assumptions can then be made in conjunction with the 1D equation: (i) for a majority of floodplain flows advection is considered unimportant (Hunter et al., 2007), one can therefore remove this term and assume a rectangular channel divided by a constant flow width (w [L]), to define a flow per unit width - q (L² T⁻¹); and (ii) for shallow and wide channels, one can approximate the hydraulic radius (R) to be the flow depth (h).

The shallow water equation can then be rewritten to describe q, at a given time (t + Δ t):

$$q_{t+\Delta t} = q_t - gh_t \Delta t \left[\frac{\partial (h_t + z)}{\partial x} + \frac{n^2 q_t^2}{h_t^{10/3}} \right] \qquad Equation \ 4$$
This equation now includes the parameters of acceleration and mass, and is less likely to lead to modelled instabilities than previous iterations of equations representing these processes. The equation also can represent shallow water wave propagation, rather than a purely diffusive behaviour that was typical of previous storage cell models.

To further reduce instabilities that might occur at shallow depths when friction becomes large, one can replace q_t in the friction term with $q_{t+\Delta}$. This produces an equation which is linear in the term $q_{t+\Delta}$, and has improved convergence properties denoted by an implicit time stepping scheme. The resultant equation contains a friction term where, as the denominator increases, flow is forced towards zero - a situation that would be expected at shallow depths (Bates et al., 2010).

2.3.3.2 2D Flow

The above equations can then be transformed to represent a two-dimensional dynamic flow field, using a storage cell concept applied over a raster grid (Bates et al., 2013). These equations include the assumption that, flow between two cells is a function of the water surface height difference between said cells (Estrela and Quintas, 1994):

$$\frac{\Delta h^{i\cdot j}}{\Delta t} = \frac{Q_x^{i-1\cdot j} Q_x^{i\cdot j} Q_y^{i\cdot j-1} Q_y^{i\cdot j}}{\Delta x^2} \qquad Equation 5$$
$$Q_x^{i\cdot j} = \frac{h_{flow}^{5/3}}{n} \left(\frac{h^{i-1\cdot j} - h^{i\cdot j}}{\Delta x}\right)^{1/2} \qquad Equation 6$$

Where h (L) represents the water depth, and i and j represent cell co-ordinates. Therefore meaning $h^{i,j}$ is the water free surface height at node (i,j). x^2 represents a cube with a dimension x (L) - displayed in Figure 8. Subsequently, n is the Manning's coefficient of friction, and Q_x and Q_y represents the volumetric flow rates (L/T) between adjacent cells. Flow depth (h_{flow}) is defined as the depth (L) through which water can flow between adjacent cells - explicitly the difference between the highest water free surface and the highest bed elevation in the cells considered. Finally, Δh (L) is defined as water depth, which is a function of how much water is flowing in from upstream, minus how much is flowing downstream (Horritt and Bates, 2002) - displayed in Figure 8.



Figure 8. Bucket Model showing discharges (Q) in and out for a cube with dimensions x^2 at coordinates (i,j).

2.3.3.3 Courant-Freidrichs-Lewy Condition

The final part of the LISFLOOD-FP formulation is the time step (t) that is controlled by the shallow water Courant-Freidrichs-Lewy (CFL) condition - Equation 7. Whereby, for the solution of a hyperbolic system, such as that of the shallow water equations, CFL conditions are required to control the system so that the wave does not propagate across more than one cell per time step.

$$\Delta t_{max} = \alpha \cdot \frac{\Delta x}{\sqrt{gh}} \qquad Equation 7$$

Where *a* is a coefficient typically defined between 0.3 and 0.7 (Bates et al., 2010). This works to improve the robustness of a model, and to reduce numerical instabilities. However, it is not sufficient to produce full stability in a nonlinear system. One can also note from Equation 7, that stability is strongly influenced by grid cell size and water depth - as the larger the cell size the longer the model time-step.

The above text explains the core hydrological equations contained within LISFLOOD-FP. Based on these fundamental equations, the model then contains three options for the calculation of water flow between cells, which vary in their physical complexity. The choice of numerical scheme will depend on the characteristics of the system to be modelled, requirements on time of execution,

and the type of data available. In the simplest case, the model assumes that flood spreading over low-lying topography is a function of gravity and topography, whilst the most complex case uses the full shallow water equation. Channels can also be represented as features within the 2D grid structure using a sub-grid version of the model. This calculates the combined flow of water within each cell, contained both within any section of channel, located in that cell, and across the adjacent floodplain, using an approximation to the one-dimensional St. Vernant equation without advection (Horritt and Bates, 2002).

2.3.4 CAESAR-LISFLOOD

Recent increases in computational power, coupled with the development of new flow algorithms such as LISFLOOD-FP, has meant it is now possible to model longerterm geomorphic change in a more realistic and physically based way than previously. This development led to Coulthard et al. (2013) to produce CAESAR-LISFLOOD (CL) - the model used for the reach scale analysis of this thesis. This model takes the hydrodynamic flow model (LISFLOOD-FP) and incorporates it within the larger structure of the landscape evolution model (CAESAR), to route flow with a higher level of mathematical reasoning than was accounted for in the previous adaptation of the model. The combined model hopes to retain an appropriate run time speed through computational efficiency, while having a stronger physical basis than the previous CAESAR model. Coulthard et al. (2013)'s work represents the first example of complex hydrodynamic effects being represented in a landscape evolution model. This development has addressed a process limitation which existed in previous landscape evolution models, and is a step change leading to the development of a 'second generation' of hydrodynamic models that could lead to significantly different conclusions about the physical basis of landscape evolution.

2.3.4.1 Combined Model Challenges

When integrating the two models, detailed above, there were some disparities which caused minor complications. In CAESAR, flow is routed from a cell in eight directions, whereas in LISFLOOD-FP it is routed in four. When considering larger areas and coarser DEMs, four cardinal directions (i.e. the von Neumann method displayed in Figure 3) are not sufficient as it will not allow a single thread channel to develop diagonally along the inter-cardinal directions. This is compared to the Moore method which allows interactions with all neighbouring cells. Therefore, a

finer grid resolution may be necessary when considering such narrow diagonal channels (Coulthard et al., 2013).

This difference in the number of flow routing directions should also be considered in relation to sediment transport. In the combined model, bedload transport is calculated for four flow directions. However, this produced a positive feedback which led to the progressive narrowing of channels. A function which describes in channel lateral erosion was subsequently added to the model to control this channel narrowing feedback, as per the method described in Coulthard and Van de Wiel (2006) and Van de Wiel et al. (2007). Additionally, the model continued to maintain the multiple grainsizes, active layers and sediment transport equations used in the original CAESAR model - described above.

2.3.4.2 Steady vs Non-Steady Flow Models

Initial work by Coulthard et al. (2013) demonstrated that CL produced significantly different sediment outputs in comparison to steady flow models, namely the flow sweeping model, which was the primary modelling method used to model fluvial flow prior to the time of this study. Steady flow models, such as the initial adaptation of CAESAR, use 'flow-sweeping' algorithms to calculate steady state uniform flow approximations for a field. From this, a discharge is then distributed to every cell within a two to five cell range downstream. This is done whilst considering the water surface elevation between the contributing and receiving cells. Value for flow depth and velocity are then calculated from these discharges using Manning's Equation (8). Subsequent values for shear stress and fluvial erosion and deposition are then also produced (Coulthard et al., 2013).

$$V = \frac{k}{n} R_h^{2/3} S^{1/2} \qquad Equation 8$$

Equation 8 describes velocity (V (L/T)), averaged by cross sectional area. Where n is Manning's coefficient of friction, R_h is the hydraulic radius (L), S is the channel bed slope (L/L) when water depth is constant, and k is the conversion factor which allows for both the incorporation of SI and English units (Manning et al., 1890).

Considering this, differences in the outputs from different types of flow models could have a substantial influence on the rates and patterns of deposition and erosion, and the subsequent geomorphology of the wider river landscape. When

considering the river at a catchment scale, Coulthard et al. (2013) found that steady flow models introduced a bias of an increased sediment yield - an influence which was more pronounced for larger flow events. The corollary of this is that non-steady flow simulations were shown to have consistently lower sediment yields, and less pronounced peak outputs. Leading one to the assumption that nonsteady flow might act as an important diffusive process in a landscape in relation to sediment discharge.

One uncertainty noted by Coulthard et al. (2013) in relation to CL is its robustness over a variety of timescales and environments. In general, previous landscape evolution models were tested at a long-term equilibrium or steady state, where time evolving processes could be reasonably ignored (Hancock et al., 2002). However, CL contrasts this approach and includes transient hydrodynamic processes. The outputs from Coulthard's initial paper suggest that these transient processes could contribute significantly to the evolution of a given modelled landscape. Despite this, there is limited knowledge on the influence of these processes at varying temporal and spatial scales. With a key remaining question being, is the scale of a study relevant to these processes? This question is more prevalent now that models like this are increasingly being used in both theoretical and applied studies.

The model used for this analysis on the River Feshie is the output from Coulthard's (2013) paper - as this is the version incorporated into CLiDE (Barkwith et al., 2015). However, one should note that an updated version of CL now exists which is detailed in Coulthard and Van der Wiel (2017). This model has had a few minor adaptations in comparison to its predecessor. The most notable of these being that the updated platform is now able to model spatially and temporally variable hydrology. The code was modified so that the hydrological parameters and precipitation rates could be input via spatially fixed pre-defined areas. More detail on the minor updates are detailed in Coulthard and Van der Wiel (2017).

2.3.5 CLiDE

For the analysis of the River Feshie's catchment scale hydrological system the computer model CAESAR-LISFLOOD-DESC (CLiDE) was used - a CA model which simulates earth system interactions to produce an encompassing landscape evolution model that includes geomorphic change and a comprehensive

hydrological system (Barkwith et al., 2015). This model was created as part of the Dynamic Environmental Sensitivity to Change (DESC) project at the British Geological Survey and aims to simulate distributed surface and subsurface hydrology from sub-annual to centennial time scales.

The cohesion of the models, detailed above, was chosen because of CAESAR's ability to model sediment transport, as well as erosional and depositional processes under different climatic and land use environments on a reach and catchment scale. Combined with LISFLOOD-FP, which has been proven successful in modelling non-steady surface water flows (Coulthard et al., 2013). There was however one missing process which was not contained within CL - groundwater flow, and groundwater discharge to rivers. To fill this gap in the landscape model, the non-LISFLOOD-FP controlled surface hydrological processes within CLiDE was replaced with a distributed hydrological model that included a model for groundwater. This groundwater component is coupled to the surface model through the exchange of water via groundwater recharge, and baseflow return to rivers. A flow chart outlining the processes acting within CLiDE, and their interactions, is shown in Figure 9.





A breakdown of the specific surface and subsurface processes which are occurring within CLiDE is displayed in Figure 10. To partition rainfall between evapotranspiration, surface water runoff, and groundwater recharge, CLiDE contains the Soil and Landuse based rainfall-runoff Model (SLiM) which is based on a single soil layer groundwater recharge model (Rushton et al., 2006; Wang et al., 2012). It is a simplified method that can represent runoff and potential groundwater recharge based on temporal and distributed soil moisture conditions, which are obtained from catchment characteristics such as rainfall, potential evapotranspiration, and soil moisture content. The model has been proven to be successful via a case study by Wang et al. (2012) in the Eden Valley UK, and work has begun to apply the model to other catchments in temperate and semi-arid climates. The model is also designed so that it is easily integrated into other

environmental process based models due to its simplicity. After passing through the SLiM model, surface water is then routed using the LISFLOOD-FP model (Bates et al., 2010) - detailed above - which includes parameters such as topographical gradient and terrain frictional properties.

CLiDE's subsurface hydrology, displayed in Figure 10, can be seen to be a twolayered finite difference model, which is discussed by Ravazzani et al. (2011) to be appropriate considering such a CA approach. The upper of the two layers represents water flow along the bedrock-soil interface (saturated soil), whereas the lower represents water within the bedrock (groundwater). The model incorporates lateral heterogeneity when considering the hydraulic conductivity and specific yield parameters to allow versatility in the range of hydrological environments the model can represent. This coupled surface-subsurface component of the model allows a constrained environment to be driven by climatic data such as rainfall (Barkwith et al., 2015) - as is the case with the River Feshie.



Figure 10. Schematic representation of CLiDE model, displaying the interactions between the four main hydrological components (Barkwith et al., 2015).

44 2.3.6 Hydrology of Soil Types

CLiDE is designed to include information from various external sources, which give the model additional information regarding location specific parameters. One such input is the hydrology-based classification of the soils of the United Kingdom, developed by Boorman et al. (1995). This classification was based on existing UK data sets that described the soil type, its distribution, and the hydrological response of catchments. The classification was based on conceptual models of the processes that occur in the soil and, where appropriate, the substrate. The resulting scheme defined 29 different Hydrology of Soil Type (HOST) classes. Soils are assigned to classes based on their physical properties, and with reference to the hydrogeology of the substrate. Although information exists that describes soil types and their distributions, most of this information needs considerable interpretation prior to being used in, for example, a hydrological model. The HOST dataset aimed to facilitate easier use of this type of data by producing a classification of soils for the UK that could be applied via existing national maps to aid hydrological studies and analyses (Boorman et al., 1995).

Models produced by the HOST classification system can be subdivided into three categories: a soil on a permeable substrate in which there is a deep aquifer or groundwater (i.e., > 2m depth); a soil on permeable substrate in which there is normally a shallow water table (i.e., < 2m depth); and a soil which contains an impermeable of semi permeable layer within 1m of the surface. Within each of these three situations are variations that allow for different soil properties, and wetness regimes, that generate 11 different models. These are further sub-divided into 29 HOST classes, based on the additional properties and geology of the substrate.

The classification was developed using databases of physical soil properties alongside feedbacks from catchment scale hydrological variables - most significantly baseflow index and standard percentage runoff Boorman et al. (1995). Baseflow is defined as the portion of stream flow that is not runoff and results from seepage of water from the ground into a channel slowly over time. This is generally considered the primary source of running water in a stream during dry weather (Arnold et al., 1995). Baseflow index (BFI) is calculated from daily data and is a dimensionless variable that expresses baseflow as a fraction of total

flow volume. It is therefore possible to calculate BFI for any catchment across the UK for which flow data is stored. Additionally, Standard Percentage Runoff (SPR) is the percentage of rainfall that contributes to the increase in surface runoff (Young, 2006). In the HOST dataset, SPR comprises of 174 catchment averaged values across the UK (Boorman et al., 1995).

Another key parameter which is included within the HOST dataset is field capacity. This represents the amount of soil moisture held in the soil after excess water has drained away and the rate of downward movement has decreased. This generally occurs two to three days after rain, in pervious soils of uniform structure and texture. The physical definition of field capacity is the bulk water content retained in soil at -33 J/kg of suction pressure (Israelson and West, 1922; Veihmeyer and Hendrickson, 1931). A final parameter which is incorporated into HOST model, and is relevant for this study, is the wilting coefficient or wilting point. This is defined as the water is held so tightly by the soil matrix that the roots cannot absorb this water any further. The value for the wilting coefficient is determined physically by the amount of water per unit weight in the soil expressed as a percentage (Briggs and Shantz, 1912; Kirkham, 2005).

2.3.7 Landuse

As well as the HOST dataset, CLiDE also includes a link to the UK Land Cover Map 2007 (LCM2007). This was the first of its kind to produce a UK wide dataset with a spatial framework derived from national cartography via a generalisation process of collection. This method has allowed for dramatically improved spatial accuracy, and means the model can better represent real world objects. 99.5% of the dataset was classified using automated procedures. However, the remaining 0.5% had to be classified via visual interpretation to be deemed accurate. Following the initial classification, knowledge-based enhancements were subsequently applied to increase the refinement and accuracy of the classification using soil, altitude, and urban extent datasets (Morton et al., 2011).

The framework for LCM2007 is based on Ordnance Survey Master Map topography layer. The spatial framework was also further refined by supplementing the generalised national cartography with agricultural census data boundaries and image segments. LCM2007 is subsequently the first land cover map to provide continuous vector coverage of UK Broad Habitats derived from satellite data. The

dataset was created from roughly seventy satellite images, which were combined to create 34 multi-date summer-winter images; each representing a single year. This method was chosen as the multi-date images allow for a contrast between land cover types and therefore increase the accuracy of the classification. 91% of the UK's land area was mapped through this method, compared to 84% in the previous LCM survey carried out in 2000 (LCM2000) (Morton et al., 2011).

The classification scheme used in LCM2007 was adapted from that of the UK Biodiversity Group's Broad Habitats (BH) classification scheme - for a detailed description of these habitats see Jackson (2000). LCM2007 classifies the land cover of the UK using classes based on the Broad Habitats, with some minor differences. Table 1 summaries the classes used in LCM2007 in comparison to those of the BH scheme and previous LCM survey, additionally it describes any additions made to the system.

Broad Habitat	LCM2007 class	Notes on LCM2007 class		
"Broadleaved, Mixed and Yew Woodland"	Broadleaved woodland	Same as LCM2000 class. Differs from BH due to exclusion of Yew, which is not extensive enough for LCM to map.		
"Coniferous Woodland"	"Coniferous Woodland"	As BH and LCM2000		
"Arable and Horticulture"	"Arable and Horticulture"	As BH and LCM2000		
"Improved Grassland"	"Improved Grassland"	As BH and LCM2000		
	Rough grassland	Mix of areas of managed, low productivity grassland, plus some areas of semi-natural grassland, which could not be assigned Neutral, Calcareous or Acid Grassland with confidence by the knowledge-based enhancements.		
"Neutral Grassland"	"Neutral Grassland"	As BH and LCM2000		
"Calcareous Grassland"	"Calcareous Grassland"	As BH and LCM2000		
"Acid Grassland"	Acid grassland	Acid grassland incorporates Bracken. Bracken is a Broad Habitat under certain circumstances (see Appendix 1). Bracken can be mapped using LCM2007 methods, but it depends on image timing, so for consistency it is assigned to <i>"Acid Grassland</i> ".		
"Fen, Marsh and Swamp"	"Fen, Marsh and Swamp"	As BH and LCM2000		
	Heather	As LCM2000; spectral differences between		
"Dwarf Shrub Heath"	Heather grassland	dense heather and heather grassland enable separation spectrally.		
"Bog"	"Bog"	As BH and LCM2000		
"Montane Habitats"	"Montane Habitats"	As BH; Altitude cut-off differs from LCM2000		
"Inland Rock"	"Inland Rock"	As BH and LCM2000		
	Saltwater	As LCM2000		
"Standing Water and Canals" "Rivers and Streams'		Merged 2 freshwater BHs, as they cannot be separated from each other using the method and data used for LCM2007. In many cases small and/or narrow water bodies fall below the MMU.		
"Supra-littoral Rock"	"Supra-littoral Rock"	As BH and LCM2000		
"Supra-littoral Sediment"	"Supra-littoral Sediment"	As BH and LCM2000		
"Littoral Rock"	"Littoral Rock"	As BH and LCM2000		
	Littoral sediment	As LCM2000		
"Littoral Sediment"	Saltmarsh	Priority Habitat and of sufficient extent and spectral distinction to be mapped consistently		
"Built-up Areas and	Suburban	As LCM2000; spectral differences between		
Gardens"	Urban	urban and suburban enable separation spectrally.		

Table 1. LCM2007 classes and associated BH, as well as a description of how the two classes differ (Morton et al., 2011).

The LCM2007 dataset provides three parameters which are used in the modelling of surface water flow via the SLiM. These are the: rooting depth (m), depletion factor (-), and crop coefficient (Morton et al., 2011). Rooting depth is averaged for an area of specific landuse, and represents the mean depth to which roots

grow. Root development is highly site specific and is largely related to the physical and chemical characteristics of the soil. Maximum rooting depth can vary from 0.30 m in shallow soils, to greater than 2 m in deep sands and loam soils (Steduto, 2012). The depletion factor within a soil is the amount of water that the soil can be depleted by before crops begin to wilt, or when the soil moisture drops below an optimum level. This factor varies but it generally around 50% of the total water stored within the soil, and can also be known as soil water availability (Allen et al., 1998). This effect is primarily dependent on: the hydro-physical characteristics of the soil (i.e. hydraulic conductivity, field capacity and wilting point); the crop growth stage; the demand of water vapour by the atmosphere; and the characteristics of the root zone (Allen et al., 1998; Lyra et al., 2016). Finally, the crop coefficient (Kc) can be defined in the following equation:

$$Kc = \frac{ET_x}{ET_0}$$
 Equation 9

Where, ET_0 is the reference crop evapotranspiration, and ET_x is the maximum crop evapotranspiration. *Kc* therefore varies depending on the type of vegetation cover, and its internal properties (Allen et al., 1998; Steduto, 2012).

48

2.4 Assessing numerical model performance

As per the objectives of this thesis, the quantification of how good a given morphodynamic model is at reproducing reality is an important question in order to assess model performance. There are a number of ways to quantify this performance, such as via channel bank migration (Motta et al., 2012), a comparison of aerial imagery (Ziliani et al., 2013), or sediment budgets (Wheaton et al., 2010). These different metrics span spatial and temporal scales and resolutions. An overview of the different metrics which can be used to compare observed and predicted braided river morphodynamics is displayed in Figure 11. This breadth of metrics poses a challenge from a morphodynamic modelling perspective, whereby a user must decide which metric is best used to assess performance and observe change.



Figure 11. Composite image displaying the different metrics which can be used in all three dimensions to measure fluvial morphodynamic change (Williams, 2014).

2.4.1 Geomorphic Change Detection

A promising approach in recent fluvial studies has been the application of geomorphic change detection (Wheaton et al., 2010). This is a broadly encompassing approach which combines various river metrics and produces quantifiable statistics which tell a user about how a modelled system compares to reality - or to itself in relation to time.

Advances in remote sensing technologies have allowed for rapid acquisition of high quality topographic data (e.g., Milan et al., 2007). This in turn has advanced the monitoring of geomorphic change, as estimating sediment budgets is allowed via repeat topographic surveys (Church and Ashmore, 1998) as opposed to measuring sediment transport directly (e.g., Lane 1998). This morphological method of monitoring has developed with time and since the early 1990s has generally included the production of DEMs and DEMs of Difference (DoDs) to observe morphological change (Lane, 1994; Wheaton et al., 2010).

2.4.2 Uncertainty and Error

DoDs have proven successful in monitoring rates of geomorphic change, however there have been many questions posed regarding the uncertainties associated with DoD analysis (Brasington et al., 2000; Lane et al., 2003). The primary question associated with this type of analysis being, 'is it possible to differentiate actual geomorphic change from noise within DEM datasets?'

There are a breadth of different accuracies associated with the specific survey techniques that are used to generate DEMs, used as inputs for DoD analyses. Previous studies have shown errors which range from +/- 0.02 m to +/- 1 metre - depending on the survey technique (Carter et al., 2007). This influence of uncertainty is significant when considering a fluvial environment, as topographic changes can be of a similar magnitude to that of dataset noise. Therefore, how well these uncertainties are accounted for in an analysis can have substantial impacts for how meaningful the interpretation of a survey's results are (Lane et al., 2003, Brasington et al., 2010). One must also note that uncertainties associated with bathymetric data can be considerably higher than that of dry

50

topographic data (Sear and Milne, 2000). These kinds of uncertainties will be discussed in detail in the Bathymetric Analysis portion of this thesis. It can therefore be stated that the fundamental control on the quality of a DoD analysis will be the initial DEM quality which is primarily based on the quality of the field survey data. The errors associated with field survey data are a function of: sampling technique; survey point quality; topographic characteristics and complexity; and interpolation methods used (Wechsler and Kroll, 2006). The quality of survey data might therefore not be uniform across a given field area. Despite this, many previous studies looking at geomorphic change have assumed spatially uniform uncertainties (Brasington et al., 2000), or varied spatially only based on wet vs dry topography (Lane et al., 2003). This has consequentially resulted in over and under representation of uncertainties in key areas of interest such as banks where is there a significant step in elevation change which is not consistent throughout the landscape.

Considering this gap in the field, Wheaton et al. (2010) chose to develop a new technique that allows for the robust estimation of DEM quality and its influence on sediment budgets derived from DoDs. This has been done via a set of tools that considers the spatial variability of surface representation uncertainty, which can be calibrated and applied to any given set of topographic survey data. The broad components included in this technique are displayed in Figure 12. The GCD software has been developed to implement the probabilistic thresholding framework for uncertainty analysis in DEM Differencing developed by Brasington et al. (2000), and Lane et al. (2003), within a GIS environment. It also implements a minimum level of detection threshold approach to DoD differencing, as described by Fuller et al. (2003)., as well as estimating the magnitude of DEM uncertainty in a spatially variable way using fuzzy set theory (Wheaton et al., 2010), and discriminating DoD uncertainty on the basis of the spatial coherence of erosion and deposition using Bayes Theorem (Milan et al., 2011).

Error Analysis



Figure 12. Workflow of DEM differencing with error assessment (Williams et al., 2012).

2.4.2.1 Probabilistic Thresholding

The type of uncertainty analysis considered in this study was via the use of a probabilistic threshold - the final step displayed in Figure 12. This is done via a user defined confidence interval to which a probabilistic threshold is applied to the total error produced on a DoD (Lane et al., 2003; Williams, 2012). To apply this type of threshold one must first consider the equation for the combined error in a DoD (δU_{DOD}):

$$\delta U_{DOD} = \sqrt{\delta z_1^2 + \delta z_1^2}$$
 Equation 10

Where δ_{z1} and δ_{z2} are errors related to Z_1 (the initial DEM) and Z_2 (the subsequent DEM), respectively. This equation can then be used to apply a Minimum Level of Detection in a geomorphic change detection analysis. Whereby δU_{DOD} is applied as a constant threshold across the DoD. This method is deemed as a conservative approach to determining errors associated with DoDs, as nothing below the minimum level is then deemed reliable. Whereas in reality this is likely not the

case for the entirety of the DEM, and in fact volumetric and areal estimates of geomorphic change are sensitive to such a minimum threshold. This means that real morphological change is liable to be lost below such a threshold (Brasington et al., 2003; Wheaton et al., 2010). One might consequently look for a more appropriate way to represent DoD errors. If it is assumed that estimates of δz are approximated by the standard deviation error (σ), and have a normal distribution Equation 10 becomes:

$$U_{crit} = t \sqrt{SDE_1^2 + SDE_1^2}$$
 Equation 11

Where U_{crit} represents the critical threshold error, and the SDE parameters are the standard deviations of both Z variables. The variable t is the critical t-value for a two-tailed Students t-distribution for a chosen confidence interval:

$$t = \frac{|z_1 - z_2|}{\delta u} \qquad Equation \ 12$$

Where $|z_1 - z_2|$ is the absolute value of the DoD. One can then relate it to the standard deviation error and desired confidence interval of an analysis. Where, for a test at a 68% confidence interval (1 σ) $t \ge 1$, and for a test at a 95% confidence interval (2 σ) $t \ge 1.96$.

This method of analysis offers a technique to remove any systematic bias in a dataset, through filtering elevation changes based on confidence that detected change is real. The additional benefit of this is that a user can define a confidence interval suitable for their analysis. However, the standard deviation error for both a 68%, and 95% confidence interval are not always uniform across an area. Previous studies which have performed a probabilistic threshold DoD analysis, have defined σ 1 and σ 2 areas based on whether the topography is wet or dry, with the assumption that dry areas are likely to be more accurately represented (Brasington et al., 2003; Lane et al., 2003; Milan et al., 2007; Bird et al., 2010). However, this is still a broad and spatially coarse view of errors across a landscape. To produce a higher level of accuracy on results it might therefore be necessary to use information obtained from specific GCPs (Brasington et al., 2000). An example of a spatially varied approach to this method was carried out Williams et al. (2011),

who used TLS survey data to vary standard deviation error values across dry topography.

This method of analysis is therefore more likely to produce reliable estimates of morphological change in comparisons to a minimum level of detection approach. However, probabilistic thresholding is still not deemed fully accurate and reliable, as smaller changes in elevation might not be picked up by this form of analysis. This becomes most significant when considering scenarios such as sedimentation on a floodplain, which can have a significant areal extent but due to its minimal elevation change is often misclassified in error analysis as noise, as opposed to morphological change (Williams, 2012).

2.5 Broader Context

2.5.1 Accurately Modelling Geomorphic Change

Many numerical models now exist which replicate the processes of river hydro and morph dynamics (Guan et al., 2016). However, a considerable amount of uncertainty still exists when trying to replicate the processes of meandering and braided river dynamics due to the presence of helical, or secondary, flows which are more complex to accurately model. (Blanckaert, 2015). There is therefore a gap in the field for models which can represent these dynamics, to produce reasonable results with regard to geomorphic change.

A study by Guan et al., (2016) used lab and field experiments to demonstrate the applicability of using a 2-D depth averaged model to model morphological change at river bend, with an aim of improving the abilities of 2-D numerical modelling in for sinuous flow paths. Their results showed that the model was able to reconstruct both hydrodynamics and morphological features to a reasonable degree. However, the model was not completely robust with further work needing to be carried out to better represent secondary flow, and the parameterisation of grain size. Results did however show that factors such as sediment transport formula and roughness height have less significance than previously thought on the evolution of bed morphology at a river bend.

54

This type of modelling is still in its infancy, with models needing further work to produce robust conclusions. Guan's analysis only considered a single river bend over a one-year period, meaning limited conclusions can be based on the results of this analysis. Therefore, for the analysis performed in this thesis a more robust and accepted model. However, model's such as this are vital in highlighting the intricacies involved in modelling river flow and subsequent bed change. More research is therefore needed into developing such models, and solving the underpinning physical processes which occur in meandering or braided river channels.

2.5.2 Geomorphic Change's influence on Flood Risk

Due to a warming climate, flood events in rivers are increasing in occurrence and magnitude (Hirabayashi et al., 2003). When considering how flood risk in a river system will evolve with time in a warming climate, it is vital to account for the impact of geomorphic change. Work by Lane et al., (2007) demonstrated this, where they used one- and two-dimensional flow models to simulate flood inundation in response to sediment delivery in a temperate upland gravel-bed river. They compared observed flood events, to equivalent events with the addition of a climate change parameter and found that the inundation area increased by around half when climate change was added. This conclusion was especially significant in relation to piedmont regions, which can supply substantial amounts of coarse sediment to the river in relatively short timescales. This work highlights the potentially severe consequences that can arise as a result of sedimentation coupled with climate change. Additionally, it highlights how study areas such as the River Feshie, which have significant sediment supply and mobility, should be areas of focus in order to draw broader conclusions which have globally reaching implications.

One caution associated with the results of Lane's study was that the modelling technique used to quantify this process wasn't part of an end-to-end framework as it dealt with discrete topographic data surveyed at fixed times. This means that limited robust conclusions can be made about how piedmont systems develop through time. Highlighting a key knowledge gap which still exists when considering geomorphic modelling which incorporates the influence of climate change. The modelling framework provided in this thesis hopes to guide other similar studies

on how to apply a comprehensive and continuous geomorphic change model in a upland temperate river setting.

3.1 Catchment

3.1.1 Characteristics

The River Feshie is situated in the western Cairngorms, an area in the south east of the Scottish Highlands. It drains a catchment of 231 km² (Figure 13) and flows in a general northward direction, downstream into the Spey Valley (Rumsby et al., 2008). The catchment has a total relief of 1030m (Wheaton et al., 2010), with a maximum altitude of 1260m (Kasprak, 2015). This relatively high altitude is due to the catchment being underlain by resistant igneous and metamorphic rocks. Moine schist comprises most of the catchment's bedrock, along with a small portion of granite which underlies higher land in the north-east (Werritty and McEwen, 1993). These two rock types dominate the coarse proportion of the river's sedimentary load, with typical d50 values for schistose material being between 50 and 110 mm (Brasington et al., 2012).

In the catchment's more recent geological history, it can be noted from landforms that Glen Feshie has been subject to substantial glacial influence. The area was last glaciated during the Late Devensian and went through an episode of deglaciation ~13 000 BP (Gilvear et al., 2000; Wheaton et al., 2010; Wheaton et al., 2013). The remnants of this glaciation can now be observed as large fluvio-glacial outwash terraces which comprise the flanks of a wide U-shaped valley - representing a significant sediment source for the river.

3.1.2 History as a Braided River

The River Feshie is a gravel-bed river, which has been classed as braided in sections since the mid 1800's (Ferguson and Werritty, 1983). Braiding has occurred in Glen Feshie due to the combination of a substantial amount of coarse sediment availability coupled with steep channel gradients, high runoff, and a subsequent flashy flow regime (Rumsby et al., 2008). This dynamic regime has allowed for periods of avulsion and channel switching (Rumsby et al., 2001) and has led to the development of extensive and variable gravel bar deposits. The river's discharge is highly influenced by snowmelt with its annual hydrograph reaching a maximum value between the months of January and May (Ferguson, 1984). However, the

catchment receives mean annual precipitation of ~1400mm, highlighting that there is the potential for flooding via rainfall all year round (Rumsby et al., 2008).

The river is characterised by this distinct flashy flow regime because of its landscape characteristics - a mountainous catchment with steep valley sides. This has resulted in numerous flash flood events over the river's history which have had a sizeable impact on the river's morphology. The average discharge recorded at the SEPA Feshiebridge gauging station, the reference location which will be used for comparison in this study, is 91.63 (m³ s⁻¹). However, this has been recorded to reach 317.67 (m³ s⁻¹) (December 2015), as a result of a flash flood (SEPA, 2018). In spate, the river has a high erosive power and sediment transport capacity which has been observed to dramatically reshape its fluvial landscape (Wheaton et al., 2013).

3.1.3 Landuse

In the nineteenth and twentieth centuries, an increased number of grazing livestock and timber felling in Glen Feshie lead to a decrease in the extent of woodland. This situation was worsened by a boom in stag shooting, which meant a higher number of deer were maintained within the estate than would be naturally be the case. Large herds of deer dramatically contribute to the decline in woodland, as seedlings which took root were browsed out and did not have a chance to prosper, meaning the new generation of trees could not replace the old that were falling. Coupled with extensive felling during the second world war, over time this lead to a dramatic decrease in the number of Caledonian Pine trees in the Glen Feshie estate. This reduction in tree cover is likely to have had impacts on the broader landscape - influencing factors such as slope instability and hydrological processes (Macdonald, 2016).

This situation remained until the start of the twenty-first century, at which point a team of government agency scientists were appointed to carry out a research project in the catchment with the aim of developing management strategies to protect the area's overall ecology. This study concluded that the deer population should be cut, in conjunction with a replanting effort to rebuild the Caledonian Pine's previous stronghold. However, the controversy of a deer cull prohibited the implementation of this strategy for multiple years following the report. Despite this, in recent years the catchment has experienced a significant evolution from its state in the late twentieth century, to the present day. The estate owners are now implementing longer term development plans, led by the rise of similar ecological restoration initiatives (Macdonald, 2016).

After an estate-wide replanting effort, and a significant deer cull, from 2010 onwards there has been a marked reduction in deer numbers in the Glen Feshie estate, which has coincided with a strong pulse of natural regeneration of various tree species and other shrubs. Additionally, it has been noted that the ecology in the Glen is more diverse, and that woodland's herb layer is much more well-established, than previously. As well as vegetation, there has also been a marked increase in the number of species, and their populations, present in the catchment. Increasing number of black grouse and salmon have led to a further strengthened ecology and displayed the success of the conservation of the landscape (Macdonald, 2016). The ecological evolution of the River Feshie's catchment may therefore be a factor that should be taken into consideration when evaluating catchment scale models over decadal to centennial timescales.

3.2 <u>Reach</u>

There are three significant braided sections of the River Feshie: at the confluence with the River Spey, at the Allt Chomraig confluence and at Feshie Lodge (Werrity and McEwen, 1993). The last of these, a 3km long section, is the most active and dynamic of the three reaches (Wheaton et al., 2013). A 2-km length within this reach will be the focus area for this study (Figure 13). This 2-km long section has been the subject of multiple scientific studies over the past decade (Brasington et al., 2007; Hodge et al., 2009; Wheaton et al., 2010, 2013), meaning there is an archive of datasets which allow observation of channel change over varying timescales. Such datasets, which have been updated to the present day as part of this study, contain: RTK-GPS topographic data from 2000 through to 2018 (see Table 2), hydrograph datasets for a period of almost 60 years, aerial photography records which date back ~60 years and Ordnance Survey channel planform maps from 1869 onwards (Kasprak, 2015).



Figure 13. Location of both Reach and Catchment scale models in relation to Scotland.

Date	Method	Author
2000	Photogrammetry	Brasington et al., 2003
2002-07	RTK-GNSS	Wheaton et al., 2010
2008	RTK-GNSS &TLS	Kasprak et al., 2018
2013	RTK-GNSS & TLS	Kasprak et al., 2018
2017	RTK-GNSS & TLS &	Williams et al., 2017
	SfM	*Stott, Masters Thesis
2018	RTK-GNSS & SfM	*Stott, Masters Thesis

Table 2. Reach scale DEMs obtained from topographic surveys - date, method, and author.

4.1 Structure from Motion - Topographic Reconstruction

4.1.1 Introduction

This chapter of the thesis focuses on reconstructing the topography of the River Feshie field site, at a reach scale. The chapter has two sections; reconstruction of dry topography, and reconstruction of bathymetry. The former was be done through the use of a SfM analysis which included a rigorous investigation of GCP deployment and placement. The latter was done through the use of a refraction correction technique. In more detail, the first section (4.1) of this chapter aims to define a survey protocol for deploying Ground Control Point (GCP) targets across the River Feshie field site. Data from a field campaign on the River Feshie carried out in June 2017 was used for the first component of this analysis. The results and subsequent conclusions from this work were then used to define a survey protocol for subsequent field campaigns carried out on the River Feshie site, throughout the course of 2018. This analysis will use the breadth of data available from the 2017 campaign, to define efficient protocols to be carried out on successive field surveys. This is with the hope that it will shorten time spent in the field, reduce the labour needed, and minimise uncertainties regarding survey method.

4.1.2 Methodology

4.1.2.1 Fieldwork component

A field campaign was carried out on a ~3 km reach of the braided river. Data collection commenced on the 01/06/17, and over the period of two weeks, a team collected all the required aerial imagery and topographic survey data for this analysis. The fieldwork included three steps. First, a GNSS Base Station was set up over a new benchmark. The Base Station was initially left to record GNSS observations for a period of approximately seven hours. These observations were then post-processed with RINEX data from a nearby permanent GNSS station so as to calculate the horizontal position of the benchmark in the OSGGB36 coordinate system and the orthometric height of the benchmark using the National Geoid Model OSGM15. Second, 86 plastic targets were set out across the study area. Each target was 0.5 m by 0.5 m and had a yellow cross constructed from 50 mm wide marking tape. The pattern of targets tried to maintain a fairly consistent spacing

across the breadth and length of the study area. RTK-GNSS was then used to survey the centre point of each target. Positions and associated errors in the x, y, and z directions were stored on the Leica Viva RTK-GNSS receiver. Third, a DJI Phantom 4 Professional UAV was then flown at a height of approximately 50 metres above the river bed, along the full length and width of the braidplain. Six separate flights were completed to obtain the total area desired, as the UAV was limited to approximately a 25-minute flight time due to its battery life. Flight paths were chosen so that adjacent images would overlap in order to produce a representative overall coverage of the study area. The aerial image side and end overlap produced was 70%, which is deemed to be within the appropriate bounds when considering the flight altitude and the type of terrain surveyed (Carrivick et al., 2016; James et al., 2017). More details on the UAV data obtained can be viewed in Table 3 below.

Setting	Braided River Survey, River Feshie (01/06/17)
Camera Model Name	FC330_3.6_4000x3000 (RGB)
Number of Images	1489
Sensor Size	1/2.3" CMOS/ Effective pixels:12.4 M
Focal Length	2285.720 [pixel] / 3.610 [mm]
Aperture	20 mm [35 mm format equivalent]
Median Camera Speed	5.115 [m/s]
Median Rolling Shutter	53.4498 [ms]
Readout Time	
Flight Height	50 m above ground level
GSD (Average)	2.74 cm
Area Covered	0.5703 km ²
Perspective of Images	Vertically
Acquired	-

Table 3. Background information of UAV imagery acquired based on Standards by O'Connor et al.,(2017).

4.1.2.2 Post Processing

Overview

After completion of the June 2017 field campaign, two primary data sets had been obtained: a compiled list of RTK-GNSS target measurements, and a consistent set of aerial images. First, the RTK-GNSS point data were converted in to the same coordinate system as the UAV data - in this case the OSGB36 British National Grid (using OSTN15/OSGM15 transformation models (Ordnance Survey, 2018). The aerial images were then imported into the SfM-MVS software Pix4d. The aim of

applying this software is to, after three stages of processing, reconstruct the topography of a chosen study area by producing a Digital Surface Model (DSM).

Prior to the first stage of processing, the correct coordinate system for the UAV images must be chosen and the 'Image Properties' must be altered to account for the specific type of UAV used. Once these initial parameters had been input, processing step one could commence. This involves an automated process which: allows the set-up of a targeted number of keypoints, calibration of cameras, rematching, pre-processing, and then exportation. Following this, the file containing the target information was imported into Pix4d via the 'GCP Manager'. The information from which was subsequently manipulated using the 'rayCloud' Editor' which displays the position of all the images collected, as well as an initial point cloud. Using this tool, the position of three targets were calibrated manually. To do so, the user locates the centre point of the target on all UAV images it is displayed in. After three targets are fully defined, the software then uses an automatic marking system to allocate the location of the remaining targets. Manual checks were then done to make sure the automatic marking process was working accurately, and to move and recalibrate the center points on any targets where this was not the case.

Pix4d initially sets all the targets as 3D GCPs, which are used for scaling and georeferencing. However, for this analysis it was necessary to have a combination of both GCPs and Checkpoints (CKPs) because the 3D error of the DEM needed to be assessed. CKPs were necessary so that the errors associated with georeferencing - based solely on the GCP data - could be quantified based on a chosen point location that is not used during the SfM-MVS processing; a checkpoint. The user must therefore specify on each individual target if they want it to be a GCP or CKP. Once the user is happy with their configuration they can run a 'Rematch and Optimize' process which recalibrates the new configuration of GCPs and CKPs. Once this processing step has been run, an option can then be selected to generate a 'Quality Report' of the data set. The information contained in which displays numerical values of various statistical errors in three dimensions for both GCPs and CKPs. This 'Rematch and Optimize' process can be re-run any given amount of times to continually improve the accuracy of the data and to minimise any errors. Equally, additional runs can be completed with different

64

configurations of GCPs and CKPs to compare error statistics for these different configurations.

Analysis

For this study, the outputs from eleven different GCP target scenarios were analysed. The configurations of these scenarios can be broken into three different categories (as detailed in Table 4) where:

1) A set amount of GCPs were chosen in a configuration that enabled them to be fully distributed across the study area without any bias, i.e. as evenly distributed as possible.

2) GCPs were selected at either the 'edge' or the 'centre' of the study area.

3) GCPs were selected based on the number of images they appeared in (\geq 10 images, or \geq 15 images).

The quality reports of each of these scenarios were then compared to produce a quantifiable comparison between the different GCP and CKP configurations. The key statistic which this analysis focused on was the Route Mean Squared Error (RMSE) of the checkpoints in the X, Y and Z directions. This RMSE statistic gives information regarding the standard deviation of the residuals (i.e. how far from the regression line data points are). Therefore, the RMSE error can indicate how concentrated a data set is around the line of best fit. When considering this statistic in relation to checkpoints, a lower RMSE would indicate a more accurate and reliable configuration.

In these processing scenarios, CKPs are used as a 'check' to display how accurate any given configuration of GCPs is. For each scenario, the point cloud produced is created with reference to all listed GCPs. However, it is the CKPs which should be analysed at this stage in processing to determine if the number and configuration of GCPs is appropriate, since they are independent from the SfM-MVS processing. Thus, if the RMSE of the CKPs for any given run is low, this is a reflection that the GCP configuration should be fairly accurate with low residual errors. 66 4.1.3 Results

Scenario	Number	Number of	Description	Configuration
Number	of GCPs	CKPs		Category
1	0	56	0 GCPS	1
2	5	51	5 GCPs distributed evenly	1
3	9	47	GCPs shown in \ge 15 images	3
4	10	46	10 GCPs distributed evenly	1
5	15	41	15 GCPs distributed evenly	1
6	20	36	20 GCPs distributed evenly	1
7	27	29	GCPs shown in \ge 10 images	3
8	27	29	GCPs in "centre' configuration	2
9	28	28	Equal Amount of GCPs and CKPs	1
10	29	27	GCPs in 'edge' configuration	2
11	56	0	All accurate targets listed as GCPS	1

Table 4. Information related to the 11 different target configurations, sorted by the number of GCPs in ascending order. The second column displays the number of 'accurate' CKPs. This makes reference to the fact 15 targets have been deemed inaccurate due to the fact they have not been calibrated in any aerial images - these targets have consequentially been omitted from the analysis. A brief description of each run is then detailed, followed by a 'category' number which identifies the three different types of configurations.



Figure 14. A comparison of RMSE values in all three dimensions. Triangles denote the 'edge' scenario, and diamonds denote the 'centre' configuration for each direction. The graph displays how error values vary depending on the number of GCPs used, and the spatial direction.

4.1.3.1 Z axis error

With reference to Figure 14, one can note that, in general, RMSE decreases in all dimensions as the number of GCPs increase. This trend is perhaps the most pronounced for the Z axis. This is because there are higher absolute values for RMSE, due to an inherent lack of accuracy in the vertical axis. Vertical accuracy is known to be harder to quantify correctly by GNSS technologies in comparison to horizontal accuracy. This is because it is easier for satellites to triangulate a horizontal point accurately on the earth's surface, in comparison to a vertical point some distance above the surface (Uren and Price, 1994). For test scenarios where there is a higher number of GCPs, RMSE in the Z axis is comparable to that of the X and Y axis'. However, when there are 10 or less GCPs, Z axis errors become a lot greater than those of the other two dimensions. This is as to be expected for RMSE when there is fairly high uncertainty, due to a lack of georeferencing constraints, in the dataset. Because of these relatively high errors for lower GCP configurations, the Z axis dataset has a much higher range in values than the X and Y axis' - highlighted by Figure 14.

4.	1.	3.	2	Category	1 /	GCP	scenarios
	•••		_	oucesor,		001	3001101103

	RMSE, m				
Number					
of GCPs	X Axis	Y Axis	Z Axis		
28	0.0410	0.0363	0.1570		
20	0.0528	0.0524	0.1460		
15	0.0405	0.0474	0.1302		
10	0.0549	0.0566	0.2757		
5	0.0611	0.0505	0.3679		
0	1.2278	0.7517	20.1878		

Table 5. RMSE in the X, Y and Z direction. Listed by the number of GCPs in descending order.

Figure 14 shows six different target scenarios for all three dimensions, each of which are based on having a set number of GCPs. The configurations of these points have been chosen in such a way that they should be spread across the study area as evenly and broadly as possible, to capture the landscape in representative

way i.e. with no preference to edge or centre GCPs. When comparing RMSE to the number of GCPs, there is not necessarily a simple negative linear trend. One can observe that when GCPs are spaced in such a manner, RMSE decreases for scenarios between 5 and 15 GCPs for both the X and Z axes. Following this point, values for RMSE do not decrease any further for GCP scenarios containing both 20 and 28 GCPs. In fact, in certain scenarios the RMSEs become notably higher e.g. the X axis RMSE for the 20 GCP configuration.

When considering the Z axis error, it must be noted that it is not solely dependent on the number and configuration of GCPs. Other factors can influence this error, such as inherent errors associated with the RTK-GNSS survey. There is therefore an additional error metric associated with the survey data obtained for the GCPs themselves. As with the error on CKPs, the Z axis error on surveyed GCPs is notably harder to constrain due to its vertical component. Additionally, there is a higher likelihood of human error associated with obtaining this parameter. This is more so the case when surveying on uneven topography, where the pole of the handheld RTK-GNSS is liable to move during any single point measurement. An example of when this might happen is on a poorly consolidated silty channel bed, where the surveying pole is liable to sink into the bed during the measurement. However, consistency and surveying rigour was discussed prior to all field data collection to minimize this human error parameter.

The interpretation of Figure 14 is slightly more complex when considering the Y axis, one can consequentially refer to Table 5 for more detail regarding precise error values. Between 5 and 20 GCPs error fluctuates up and down and shows no distinct trend. However, the amount they vary is fairly minimal, with there only being a 0.009 m difference in RMSE between the highest (10 GCPs) and the lowest (15 GCPs) values. Considering this, 15 GCPs is therefore the most optimum scenario when considering 20 GCPs or less. For 28 GCPs there is a slight decrease in RMSE, however this means the addition of 13 targets for only a 0.011 m drop in error. Therefore, depending on the survey specification, the deployment of more than 10 additional targets would perhaps not be a worthwhile given the time, money and manpower involved for a small decrease in error.

4.1.3.3 Category 2 GCP Scenarios

With reference to Table 6, one can note that 29 GCPs are used in the 'Edge' scenario, and 27 GCPs are used in the 'Centre' Scenario. While considering this, one can then look to Table 6, which quantifiably compares the differences between the two configurations in terms of RMSE in all three dimensions. The table displays that all errors in the 'Centre' configuration are consistently higher than those of the 'Edge', despite the two having a similar number of GCPs. The table displays that the greatest directional difference is along the X axis; with RMSE in the 'Centre' configuration being 83.6% greater than the 'Edge'. Although the numerical value of difference in the Z direction is higher, Z axis errors are naturally higher, and the relative percentage difference is comparative to the X axis; and is in fact lower by 1.7%. The table also shows that errors in the Y direction are notably lower than in the X for both scenarios. When comparing the difference between the two errors statistics, they are 36.6% greater along the X axis than in the Y.

This pattern of Y axis errors being lower than X, was also observed by Javernick et al. (2014)'s study of the errors associated with SfM technologies in a shallow braided river setting. Here, GCPs were set out in a regular gridded format across the survey extents, and were not altered throughout the analysis. With this configuration, the team found that Y axis RMSE was consistently 0.01-0.02 m less than that of the X axis. They posited this to be due to the longer longitudinal dimensions of the study area in a North-South direction, as opposed to East-West. When considering this fact, and the applicability of the River Feshie site also in these rough geographic orientations, one could posit this to also be the reason for reduced Y axis errors in this analysis.

The overall conclusion on Category 2 scenarios is that the 'Edge' configuration produces overall lower errors, and consequentially more accurate and representative results. When considering the two as fixed extremes at two ends of a spectrum one can state that the 'Edge' configuration would be the preferable option. However, GCP configurations are not necessarily fixed to either extreme. Subsequently, the most ideal configuration would most likely be situation were the edges of a study area have a good coverage but there are additional points in

the centre so as to not lose accuracy there, and to maintain a representative overall coverage.

RMSE, m	X Axis	Y Axis	Z Axis
Edge	0.032	0.031	0.080
Centre	0.059	0.051	0.146
Difference	0.027	0.020	0.066
Difference as a % of			
'Edge'	83.60%	64.39%	81.90%

Table 6. RMSE comparison between 2 GCP scenarios; 'Centre' were GCPs are chosen in the central area of the study area only and 'Edge' where GCPs are chosen to cover the edges of the study area only. The third row lists the numerical value of difference between the two scenarios for each axis, followed by this difference as a percentage of the 'Edge' scenario - the more accurate of the two.

4.1.3.4 Category 3 GCP Scenarios

This subsection of the analysis investigates the hypothesis that, 'the more UAV images an individual target appears in, the better constrained that target is'. If the Pix4D georeferencing procedure was based on a scenario where the GCP targets used were observed in a relatively high number of aerial images, it is likely to result in a higher accuracy post-processing product than if the targets were to be observed in fewer images?

					Number
RMSE, m	X Axis	Y Axis	Z Axis	Scenario	of GCPs
≥ 10 Images	0.0468	0.0498	0.1271	A	27
≥ 15 Images	0.0798	0.0761	0.1855	В	9
Difference between A					
and B	0.0330	0.0263	0.0583		18
Difference as % of A	70.54%	52.85%	45.89%		
10 GCP Scenario	0.0549	0.0549	0.0549	C	10
Difference between B					
and C	-0.0249	-0.0212	-0.1306		1
Difference as % of B	-31.23%	-27.83%	-70.41%		

Table 7. RMSE in all three dimensions for three GCP scenarios; A-C. Additionally, a numerical and percentage comparison of A and B, and B and C is listed.

70

To test this hypothesis, two different GCP configurations were run; one where targets were only marked as GCPs if they appeared in 15 or more images, and one were targets were marked if they appeared in 10 or more images. However, as one might expect, there were a lower number of targets which appeared in \geq 15 images compared to \geq 10 images. This subsequently resulted in a relatively large disparity between GCPs in the two test scenarios. The \geq 15 images run contained only 9 GCPs, whereas the \geq 10 images run contained 27 GCPs - a reduction in GCPs of two thirds. When considering this, GCP scenario C - which contains 10 GCPs - was added to Table 7, to allow for a more realistic comparison. In scenario C, GCPs were chosen purely to be distributed in a representative way across the landscape, and the number of images in the target appeared in was not a factor.

With reference to Table 7, one can observe that RMSE has increased (i.e. accuracy has decreased) for all axis' from scenario A to B. However, this increase has not been consistent throughout all three axes', with the decrease in the Z axis being less than in the horizontal plane. One can assume that the primary reason for this consistent increase is the reduction in the number of GCPs, and is not a direct result of the targets being observed in more images. In an ideal scenario, one would compare scenarios where a fixed number of GCPs had been observed in \geq 10 images, and subsequently \geq 15 images. However, completing multiple additional drone flights to test this assumption fully would not be an efficient use of resources, and is outside the scope of this analysis. Therefore, if we assume the reduction in RMSE is solely a result of the reduction GCPs, a lower percentage change in the Z axis becomes a notable result. Prompting the conclusion that, if a target is observed in more images it should be more accurate in the Z axis; assuming the number of GCPs used does not decrease substantially. This is not the same for the horizontal axis', especially the X, where RMSE has increased in line with the decrease of GCPs and results do not appear to be as significantly influenced by the number of images they appear in.

Table 7 also shows a comparison of scenarios B and C, which considers the influence of target layout. With reference to the Z axis, one can see that RMSE has decreased substantially between B and C despite the fact a similar number of GCPs are used. One can therefore assume this change is due to the configuration of GCPs. In scenario B points are clustered into an area which is observed in many
72

images, whereas scenario C covers the study area broadly and in a much more representative way. Showing that, the configuration of GCPs has a much higher influence on the accuracy of results for the Z axis, in comparison to the number of images a GCP is shown in. This conclusion is also true for the X and Y axis' but to a lesser extent.

In summary, for scenario 3, we can state that: (i) the number of images a target appears in has little influence on its accuracy in the horizontal plane; (ii) the number of images a target appears in has a notable influence in the Z axis, but the number of targets and their layout must also be considered; and (iii) for a given number of GCPs, the layout of the points is the primary control on RMSE, most notably so in the Z axis, and the influence of layout greatly outweighs the influence of the number of images a target appears in.

One must however note that the above conclusions have been drawn from a limited number of GCP scenarios. To ensure total confidence in these statement numerous additional GCP scenarios should be processed to ensure results are consistent. Most notably, an analysis should be done which considers the number of images targets appear in with a consistent number of GCPs.

4.1.4 Defining a Best Practice Survey Protocol

This chapter has included a thorough analysis of the influence of GCPs on georeferencing errors. In order to achieve the most accurate georeferencing results one must consider: the number of GCPs used, the layout of GCPs, and number of images the GCPs appear in. For the River Feshie field site, c. 15 GCPs is the optimum number to be deployed. These should be located so that the edges of the study area have a good coverage, as well as additional points in the centre to maintain an acceptable overall coverage. When viable, targets should be observed in as many images as possible so as to best constrain the Z axis error. However, if there are pressures on time and resources this parameter should not be made a priority, depending on the scale and resolution of the project. An example of a good configuration for 15 targets might be: 5 targets positioned on the middle of the braidplain in triangular configurations with those positioned on the banks. However, defining restricted and precise location for such targets is not necessary or realistic in a dynamic fluvial environment, where targets must

preferentially be located in locations that are less likely to undergo geomorphic change. The conclusions made from this work should be applicable for use in other SfM-MVS analyses. This analysis provides quantifiable evidence that, for this site, deploying a larger number of GCPs than 15 is counterproductive in terms of cost, time and efficiency.

The above work defines a structured survey protocol that was used in the subsequent June 2018 field campaign, with the aim of minimising errors relative to the time spent positioning and surveying targets. In the June 2018 field campaign, 20 targets were deployed - 13 GCPs, and 7 CKPs. It was determined that only 13 GCPs were necessary as the configuration of these was sufficiently well distributed - an overall good coverage with preference to the edge of the study area. Additionally, this allowed more CKPs to check the accuracy of the configuration. The configuration of these is displayed in Figure 15 below, and the errors associated with the new configuration are listed in Tables 7 & 8.



Figure 15. June 2018 field survey target layout, as per best practice protocol defined above.

June 2018 Target Layout

Check Point Name	Error X [m]	Error Y [m]	Error Z [m]
t002	0.0509	-0.0570	0.0477
t006	0.0463	0.0716	0.0289
t007	-0.0220	-0.0354	-0.0790
t014	-0.0082	-0.0446	0.0308
t015	-0.0244	0.0060	0.0016
t017	0.0497	0.0274	-0.0662
t018	0.0009	0.0031	-0.0390
Mean[m]	0.0133	-0.0041	-0.0108
RMSE	0.0346	0.0421	0.0481

 Table 8. CKP errors associated with June 2018 Targets displayed in Figure 15 - individually and then mean and RMSE values.

RMSE, m	No. of GCPs	X Axis	Y Axis	Z Axis
2017, 'Edge'	29	0.032	0.031	0.080
2017, 15 GCPs	15	0.0405	0.0474	0.1302
2018, Final	13	0.0346	0.0421	0.0481

 Table 9. Best 2017 configurations compared to the final 2018 configuration - a direct comparison of errors.

Table 8 and 9 show that the best practice survey protocol implemented in 2018 resulted in an overall reduction in errors in comparison to the 2017 survey - most significantly in the Z Axis. The Z axis RMSE associated with the 2018 survey is lower than any error associated with the 2017 survey data - even when more than double the GCPs were used in 2017. Only one Z axis RMSE was less than 0.1 m for the 2017 - the 'Edge' scenario with contained 29 GCPs. This best practice survey protocol has decreased this by almost half - highlighting a significant advancement in the constraint of this parameter. RMSE in both horizontal axis' are broadly similar for both the 2017 and 2018 surveys. However, on average the 2018 survey has lower errors for both the X and Y axis'. Especially when one considers the lower number of GCPs in comparison to some of the 2017 GCP scenarios. These results highlight the vast improvement in error results when using a structured survey protocol. It is therefore advised that all subsequent surveys carried out in a similar fluvial setting should adopt this best practice survey protocol.

4.2 Bathymetric Reconstruction

4.2.1 Introduction

After obtaining a comprehensive set of aerial imagery for the study area, it is important that both wet and dry topography is accurately represented in the DEM of the catchment. The SfM workflow detailed above aimed to provide the best representation possible of the dry topography i.e. with the lowest horizontal and vertical positioning error values possible. However, this workflow and its associated vertical error values were only applicable for dry topography.

As detailed in the SfM methodology above, Pix4d processes aerial imagery and RTK-GNSS point data in three stages. After the third and final stage of processing, the software will output a DSM that should have representative and accurate values for surface topography, and point elevations for all dry surface areas. However, DSMs generated using SfM do not accurately predict bathymetry due to the refraction of light through a water column. It was therefore decided that a bathymetric correction would be applied to the SfM data in all wetted areas, based on Woodget et al. (2014)'s method. As was previously noted, their analysis achieved acceptably low error values when analysing similar SfM aerial imagery, at a comparable scale, in a riverine environment. It must also be noted that this method only considers refraction in the vertical dimension. However, as the input data is comprised exclusively of overhead imagery the horizontal component of refraction is assumed to be negligible.

This chapter details a bathymetric reconstruction that was completed for the June 2017 topographic data, with three primary aims: (1) to produce a DEM which was representative for both wet and dry areas; (2) to identify where the bathymetric correction is working well, verses when it is working poorly; and, (3) in areas where bathymetry is poorly modelled to posit reasons for this and to suggest how this could be improved in the 2018 survey.

4.2.2 Methodology

During the 2017 and 2018 field surveys, RTK-GNSS data points were collected across the braidplain. Each of these points were labelled with a code: in channel

77

(IC), water surface edge (WSE), or dry topography (Topo). For this methodology, the WSE point shapefile was used as a component of the bathymetric correction workflow to define elevations of the channel's wetted perimeter. Such values can be obtained from other methods - such as the DEM - but this is deemed the most accurate method if the dataset is available. To maintain accuracy, this point dataset should ideally contain survey points that are longitudinally spaced at a sufficiently high density to represent changes in water levels. When surveying this parameter in the field, it was ensured that each data point was taken precisely where the dry and wet topography met. A good example of an ideal survey location was on a large clast which has a clear line showing where the clast is submerged in water. Data points were taken on both the right and left edge of the river, generally parallel to one and other to avoid inaccurate interpolation of values when water surface elevations do not correlate across the channel. The IC dataset was then also used in this analysis as a 'check' to relate observed elevation values to those predicted by the bathymetric correction.

For this analysis, ArcMap was used to manipulate datasets and reconstruct bathymetric values. The process involved a number of sequential steps which are detailed in the text below. Firstly, the orthomosaic file produced as an output from Pix4d was used to create a single polygon shapefile that included all wetted areas within the survey extents. This area was traced visually using the aerial image, and noting contrasting colour and shade differences between wet and dry areas. A triangular irregular network (TIN) was then created by using the wetted area shapefile as the extent, and the WSE point values to extract elevations. The TIN was then converted to a raster type dataset, which displays data in a gridded format with a specified cell size. The cell size for the 2017 survey was 0.1 metres, and for the 2018 survey 0.2 metres. The differing resolutions were a result of the processing time available for the Pix4d analysis. Due to the timings of this thesis there was a longer time period over which to process the data obtained in June 2017, in comparison to the 2018 survey. The original DSM produced in Pix4d was then subtracted from the new WSE raster, to obtain initial values for depth. However, a portion of these values were negative. As these are not physically plausible, a conditional statement was applied to the dataset to make these values null. The conditional stated that, 'if the pixel value in the raster is negative, return 0, or if it is positive return the raster value.' This produced an additional

78

raster containing positive depth values only, to which a refraction correction factor was applied. This was done using the Raster Calculator tool, by stating that all values in the raster should be multiplied by 1.4 - the value obtained from Woodget et al (2014) to account for the effect refraction across the air-water interface. The resultant dataset contains corrected depth values however these are not related to the elevations in the landscape. This dataset was therefore subtracted from the original WSE raster. Finally, the Mosaic to New Raster tool was used to merge the original SfM DTM with the corrected bathymetric raster.

4.2.3 Results

4.2.3.1 A Comparison to Observed Data

To determine the validity of this correction for in channel values one can look to the IC RTK-GNSS point dataset to compare observed values to those predicted by the above correction. To do this, the point dataset was loaded in ArcMap and the points converted into a TIN and then a raster as per the methodology above. The differences between the bed level rasters produced from the refraction corrected SfM DSM and the RTK-GNSS survey data were then compared by differencing the two to produce a comparison raster file - Figure 16. Figure 17 displays a histogram of the data values included in this comparison raster file, and highlights the proportion of errors greater than or less than 10 cm. Table 10 contains the main statistical parameters relating to the data in both Figure 16 & 17, highlighting generally how errors are spread and consequentially how well the bathymetric correction has worked. If one converts the data displayed in the histogram in Figure 17 into percentage values, the surface area represented by errors of less than 10cm equates to c. 80%.



Bathymetric Correction 2017. Observed - Predicted Errors

Figure 16. 2017 Aerial survey data overlain by bathymetric error statistics (SfM-GNSS). Errors above and below 10cm are displayed. Circled areas highlight high errors due to lack of WSE survey data. Zoom shows an example of where bathymetric correction has worked poorly, and therefore RTK-GNSS survey should be focused.



Figure 17. Histogram of the data contained in Figure 16, the legend displays the values contained in each of the 13 bins (metres).

Min:	-4.9877
Max:	1.6926
Mean:	-0.0625
SD:	0.1019

Table 10. Main statistical values from Differencing Raster created in ArcMap.

4.2.3.2 Errors

From an initial look at the output from Figure 16, it is clear to see that there are higher errors associated with areas where there is limited or no WSE RTK-GNSS survey data (circled areas). This is as to be expected, as the bathymetric correction was performed using this information. It therefore makes sense that in areas where this data were not available, reconstruction of depths would be less accurate. To correct for this problem, these areas were removed from the further analysis. This was deemed acceptable as all the main anabranches were still included in the analysis, as the areas removed only equated to a few minor isolated pools. After the polygon file was edited and these areas were removed, the steps detailed above were replicated to produce a more representative DEM raster.

Figure 16 highlights errors which are \pm 0.1 metre. This value was chosen as 10 cm is the approximate error associated with an RTK-GNSS survey (Bangen et al., 2014), so considering errors under this threshold would be futile. When taking a closer look at these errors it becomes clear that they are broadly associated with areas of similar characteristics. As one might expect from the literature reviewed in the previous chapter, the correction did not work as well for deeper areas, such as deep pools of scour around fallen trees (as is shown in the zoomed area). This is likely as the apparent depth observed from SfM technologies, coupled with the correction factor of 1.4, does not produce depths deep enough when considering pools which are on the order of a metre or more deep.

Another characteristic setting in the study area which showed higher bathymetric error values, were areas with overhanging vegetation, or where vegetation existed in the channel. It can be challenging for SfM technologies to discern the difference between the elevation of the ground, verses the point elevation on the top of a shrub or tree. As a limited number of trees exist on the braid plain, where the reach scale analysis was taking place, it was decided that it was not necessary to remove all vegetation from the SfM analysis, as this can be a time-consuming process which was not deemed efficient when considering the aims of this study. A limited number of errors exist where Pix4d has identified the elevation of vegetation as the elevation of the channel's apparent depth. A way to mitigate this issue is to define the wetted channel polygon boundaries precisely, so that overhanging vegetation is largely omitted from the bathymetric analysis. However, this might result in the modelled extent of the channel differing from the actual extents in reality. Another way to account for these errors would be to focus a supplementary RTK-GNSS survey in areas where overhanging vegetation is likely to obscure aerial photography from observing the location of the water's surface edge.

4.2.3.3 2018 Bathymetric Reconstruction

From the results of this bathymetric correction, based on the data from the June 2017 field survey, it can be stated that the adaptation of Woodget et al. (2014)'s method for bathymetric reconstruction produces acceptable magnitudes of errors

when considering the scope of this analysis on the River Feshie. However, the above analysis has served the purpose of highlighting how this method can be best implemented when coupled with an RTK-GNSS survey - detailing the characteristic areas where survey effort should be focused. Additionally, highlighting where potential errors may occur when using this type of methodology.

Despite the above efforts to constrain a field protocol which will aid the accuracy of bathymetric correction, the bathymetric reconstruction for the 2018 field survey data showed higher inaccuracies in comparison to the previous year. It is hypothesised that this is primarily down to the fact survey data were obtained at very low flows, meaning that the refraction correction method worked poorly - as is stated in the literature. This is highlighted in Figure 18a&b which show the highest errors (denoted by a red colour) are largely represented by areas of consistent shallow water, such as on riffles. Considering these results, the limitations of this method are detail in Section 7.1.2 later in the thesis.

Another potential reason for error associated with both datasets is the omission of refraction in horizontal directions. Dietrich (2017) states that despite using imagery parallel to the horizontal plane, points can be still be distorted in all coordinate directions. A methodology for incorporating this influence is still being developed in the literature, and uncertainties still exist regarding how much each coordinate direction is influenced by the process of refraction. As such, this factor has not been incorporated in the River Feshie's bathymetric analysis. However, it should be noted as a point for future study when more literature becomes available regarding these techniques.

82

Bathymetric Correction 2018. Observed - Predicted Errors.



Figure 18a. Bathymetric reconstruction for 2018 field survey data. Black rectangle denotes zoom in section displayed in Figure 18b.



Figure 18b. Zoom in of anabranch bend highlighting how shallower waters relate more to higher error values.



Figure 19. Histogram of data contained in Figure 18a, legend contains the values contained within each bin.

4.3 <u>Conclusion</u>

This chapter aimed to answer the first objective of this thesis: to generate high quality DEMs of the braided reach from 2017 and 2018 field survey data. Firstly, a SfM survey protocol was developed based on 2017 data. Results showed that ground control targets should be positioned so that the edges of the survey area are preferentially represented, while still maintaining a good overall coverage of the study area. This analysis also concluded that c. 15 targets should be used as Ground Control Points (GCPs) for a reach scale analysis. Subsequently, this survey protocol was used to generate the 2018 DEM, which was characterised by RMSEs on check points of 0.035, 0.042, and 0.048 for the X, Y and Z axis' respectively. These results represent a reduction in vertical error from the 2017 dataset, and highlights the benefit of this analysis.

5.1 Introduction

This chapter details how the first component of a coupled two stage modelling process was carried out. The chapter states how input files were obtained, alongside how the CLiDE model was set up. The chapter follows a calibration process, as the model is edited to try and best represent reality. The results of each model run are displayed, as well as a concluding final output run.

Chapter 4 details the methodology through which a high-resolution topographic product was obtained for a braided reach of the River Feshie. This DEM was produced to be used in a modelling framework that aims to simulate geomorphic change along the c. 2 km long reach of the River Feshie. To do this, a discharge input was needed at the reach's upper bound. As no long-term flow gauge record exists at this location, it was decided that a catchment scale model of the River Feshie would be created and calibrated to produce a dataset containing discharge values to supply to the reach scale geomorphic change model. CLiDE - the catchment scale model used - was based on inputs from rainfall data, as well as containing catchment specific parameters which characterised the River Feshie's overall hydrological system. This method satisfies the need for an input into the reach scale model, as well as providing a component of the novel 'rainfall-toreach' methodology for the modelling procedure. By integrating the two scales, predicted geomorphic change observed in the reach scale model will be a function of observed rainfall that is input into a rainfall-runoff model. This is because rainfall inputs will influence river discharge in the catchment scale model. These discharge outputs will then be used as an input into the reach scale model, which will act to drive geomorphic change.

5.2 <u>Methodology</u>

5.2.1 Digital Elevation Model of the Feshie Catchment

To produce a catchment scale rainfall runoff model a DEM representing the topography of the catchment was required. The base DEM file was obtained from Digimap - an Ordnance Survey five metre resolution digital terrain model (OS Terrain[™] 5 DTM, 2018). A DEM of the Feshie catchment was then extracted using

the Hydrology Tools in ArcMap. The first step of this was the use the 'Fill' tool, to hydrologically correct in any depressions in the DEM. These depressions might be real, or artefacts in the DEM related to its resolution or how it has been created. This was to ensure that when water is following a path downstream it does not get stuck in such depressions, and ensures that there is always a downstream path to the edge of the DEM from any pixel, i.e. removing any internally draining depressions (Zhu et al., 2013). The 'Flow Direction' tool was then used to find the direction of steepest descent to a neighbouring pixel, with the assumption that water will follow the gravitational path of least resistance. The 'Flow Accumulation' tool was then used to calculate the cumulative number of adjacent cells draining downslope into each pixel, as displayed in Figure 20.



Determining the accumulation of flow

Figure 20. Cellular approach to determining the flow direction and accumulation incorporated in ArcMap Tools (Jenson and Domingue, 1988; Tarboton et al., 1991).

From this, a map of values was created for the number of upstream contributing pixels, which was converted into an upstream contributing area by multiplying by the surface area of one pixel in the raster calculator tool. The 'Stream Order' tool was then used to produce the stream network with a reasonable number of streams flowing in the correction direction (Strahler, 1952). A threshold contributing area of 25000 m² was assumed for channelised flow. This produced a reasonable representation of the drainage network when compared to the topography. The resulting stream network and associated data layers were used to extract the River Feshie catchment required for this study. The catchment outlet was chosen to correspond with the same location as the SEPA Feshiebridge Flow Gauge (SEPA, 2018). This was to allow a direct comparison of modelled discharge exiting the catchment, to the discharge reading from the SEPA flow

88

gauge. The catchment outlet was identified on the stream network and the 'Watershed' tool identified the spatial extent of the catchment draining to that node. The resulting watershed raster file was then converted into a two-dimensional polygon file, and provided to the 'Extract by Mask' tool to generate a 5-metre resolution DEM for the Feshie catchment.

The resolution of this DEM was deemed too fine for simulating catchment hydrology over annual to decadal timescales, and would result in intractable model run times given the available computational resources. After resampling the DEM and running short experimental model simulations using CLiDE at 5, 10, 20, 50 and 100 m resolutions, a cell size of 100 m resulted in model run times that were acceptable and provided a total number of pixels less than the maximum suggested by the CLiDE model developers (< 500, 000 pixels). At this scale, enough topographic detail is retained to capture catchment-scale hydrology, whilst the model run time is appropriate to allow simulations over a 10-year timeframe. The weigh up between time and resolution was key to this decision when considering the length of time available for this thesis.

A bedrock DEM (the elevation of the bedrock-soil interface) is also required as an input for CLiDE to define the soil/superficial deposit thickness across the model. As there was no high quality, spatially distributed data available concerning the thickness of the soil and debris in the Feshie catchment, a uniform value of 1 metre was chosen to represent the depth to bedrock across the entire area. This value was informed by the BGS Superficial Thickness Map; a UK wide raster dataset which contains values for superficial thickness at a 1 km resolution gridded in a hexagonal distribution (Lawley and Garcia-Bajo, 2010). To create the bedrock DEM the ArcMap 'Raster Calculator' tool was used to subtract 1 metre from all pixel values in the catchment DEM. However, one must note that in reality the value for soil/debris depth would likely be a lot more variable. Unfortunately, to obtain enough information to represent this accurately in the model it would require costly and labour intensive geophysical ground penetrating techniques - which are outside the scope and budget of this thesis. The potential effects of this assumption on the model simulation and its outputs will be discussed in more detail later in Section 7.2.1.

It must be noted that for the catchment scale analysis of the River Feshie slope processes were not considered, as this analysis was purely to obtain catchment hydrology. These processes are only applicable when considering the reach scale modelling, and the process of sediment transport and geomorphic change. However, as DEMs focus on the floodplain as opposed to the hillslopes, and the timescales of modelling are too short to include any substantial soil creep, the significance of these processes is thought to be minimal.

5.2.2 Hydrology Inputs

A number of additional files that relate specifically to key hydrological processes which occur within the catchment were then created. These files were all sampled to the same extents and resolution as the catchment DEM detailed above.

5.2.2.1 Evapotranspiration

CLIDE requires data inputs from the Meteorological Office Rainfall and Evapotranspiration Calculation System (MORECS; Hough and Jones, 1997). The Met Office initiated the collection of information for this dataset in 1961 in the hope that it would result in the more accurate prediction of modelled river flows, as well as enabling more effective planning and assessment of drainage catchments across the UK. The dataset covers the entirety of the UK and consists of monthly values at a resolution of 40 km² (Hough and Jones, 1997). At this resolution, the dataset should capture the seasonal variation in evapotranspiration rates.

The three files associated with this MORECS dataset are the Potential Evapotranspiration (PE) grid file, the PE list file, and the PE numbers file. The first contains a single distributed asci-grid file, where individual nodes contain an identifier which relates a polygon shapefile (displaying 40 km² grid squares across the UK) to the gridded interpolation of evaporation data. The second contains a list of time stamped, monthly averaged values for PE (kg m⁻² s¹) relating to each of the identifiers. The third contains a list of PE identifiers required from the PE grid file. This file exists purely to reduce the computational cost of searching through an entire list of PE identifiers for the entirety of the UK (Barkwith and Coulthard, 2013). In the case of this study area, only two PE identifiers were listed.

5.2.2.2 Soil type and soil hydrological properties

Another UK-wide dataset that is required as an input from CLiDE is the data file associated with the HOST. This file, as detailed previously in the thesis, breaks vegetation and soil type down into 31 different classes, and defines the properties of each in relation to hydrology. Each class has a known baseflow index (-), field capacity (m), and wilting point (m). The dataset contains values for 1 km grid squares across for the whole of the UK. The model requires one file containing the HOST dataset in asci-grid format, extracted for the extents of the study area. CLiDE is then programmed to attribute values for baseflow index, field capacity and wilting point based on these class identifiers, following Boorman et al. (1995)

5.2.2.3 Landuse

90

A third input that is based on an existing UK-wide dataset is the Landuse data parameter. As with the HOST dataset, the Landuse raster file, which can be viewed in ArcMap, contains 1 km gridded data. Each cell is labelled under one of 10 class types as per Morton et al. (2011); each of which contain varying vegetation properties. CLiDE subsequently uses this input file to populate information needed for its surface partitioning model (SLiM), which requires values for: the rooting depth (m), depletion factor (-), and crop coefficient for each Landuse class.

5.2.3 Initial Hydrological Conditions

5.2.3.1 Precipitation

The sole input of water to the CLiDE model is precipitation, provided as a Daily Rainfall rate (mm/day). This data must be provided in a subfolder named 'Rainfall', which is located in the same folder as the model executable. Within this folder, there should be multiple files containing total daily rainfall, and each file should be an ascii-grid format named: rainYYYYD.asc, where YYYY refers to the year and D refers to the day (1-366). This input can be distributed across the catchment, or be represented by a single value that will be dispersed evenly. In an ideal situation, distributed rainfall data would be used. However, as the Feshie catchment is situated in a reasonably remote location there is a distinct lack of consistent rain gauge data. In recent years, a number of rain gauges have been installed in the catchment as part of a monitoring project led by Dundee University. However, due to the challenges of gauging in extreme weather

conditions, and issues with accessibility, data from these gauges is patchy and unreliable over the 10-year timespan required.

Considering this, it was decided that the best dataset to use was the Catchment Averaged Daily Rainfall Values from the National River Flow Archive (NRFA). These values were derived from the Centre for Ecology and Hydrology's Gridded Estimates of Areal Rainfall (CEH-GEAR) dataset (Keller et al., 2015; Tanguy et al., 2016). The initial dataset contained 1 km gridded data generated from daily observed rainfall data available from the Met Office, which was developed in accordance to the guidance provided in the British Standards Institution's Guide to estimate areal rainfall (British Standard Institution, 2012). The rainfall grids which were produced used the 'natural neighbour' interpolation method (Sibson, 1981), including a normalisation step based on average annual rainfall over the period of 1961 to 1990. These gridded values were then converted into catchment averaged rainfall values, to produce a more representative overall value for rainfall related to a specific catchment as opposed to on an arbitrary grid. Within this process, an averaging procedure was used to define catchment boundaries. In which, a grid cell was considered within a given catchment when more than half of its area was inside the boundary i.e. 500 m² (Tanguy et al., 2016). For further detail about how these catchment products were created see Keller et al. (2015).

As well as files containing gridded rainfall data, CEH have also produced a dataset which indicates how accurate the data might be on any given day, for any given catchment. This dataset contains the value of the distance to the closest rain gauge (m) used to interpolate rainfall at that grid point. In the file, the catchment average of this distance (m) is provided next to the value for catchment averaged rainfall (mm). This distance value can then be used as a proxy for uncertainty. Where the distance is larger, the more uncertain the rainfall value will be and the opposite will be true for the corollary.

Seasonality can also be included in CLiDE but this was not deemed to be an important initial step in calibrating the catchment's hydrology. Parameters for which were therefore left to their defaults initially. Additional tabs which include detail on Sediment, Vegetation, Slope Processes and Climate Factors are also included in CLiDE. However, values on these tabs were not edited during the

catchment analysis as these processes would not contribute significantly to the catchments hydrology when considering the landscape with no geomorphic change.

5.2.3.2 Soil moisture

CLiDE also requires inputs for both initial Soil Moisture Deficit (SMD), and initial Near Surface Soil Storage (NSSS; both in mm). These files allow the hydrological model to be initialised from steady state conditions. As per the CLiDE Manual (Barwith et al., 2013), the input files created initially contained zero values for every node within the extents of the study catchment (i.e. the landscape is initiated as entirely dry). By running the hydrological model to steady state, one can subsequently use the model outputs for SMD and NSSS as inputs for later transient runs. The process describing how to reach a steady state system is detailed in the Spin Up section later in the text.

5.2.3.3 Groundwater

Initial groundwater conditions (gridded elevation of the water table) are required as inputs to the model. In CLiDE, water is provided to the groundwater by recharge via the SLiM water partitioning code, and is removed as baseflow by the surface water routing routine incorporated in Lisflood. More detail regarding the specific code and hydraulic processes replicated in this file can be found in the CLiDE Manual (Barkwith and Coulthard, 2013). This module is initialised using groundwater head levels, and is constrained by the hydraulic conductivity and specific yield of the aquifer. These constraints can be specified as either uniform or distributed across the catchment. As no subsurface detail currently exists to produce a distributed ascii file for the Feshie Catchment, the assumption was made that these parameters are uniform across the study area. Hydraulic conductivity and specific yield were set to their default values, 10 and 0.7 respectively, in the initial Spin Up run. These values were used as purely as calibration parameters while attempting to match observed and modelled baseflow. Additionally, there is an option to vary both of these parameters' 'multipliers' in the graphical user interface (GUI). For the initial run both values were set to 1, i.e. the value should remain as inputted and not be increased by any factor.

As there was no observed information pertaining to the location of the water table, groundwater heads were varied across the catchment in situation which was though to best represent reality. When considering the assumption that groundwater flows from areas of high elevation to low elevation, as per the depth of the water table. One can consider that it might be appropriate to have lower groundwater head values on hillcrests, while also keeping them at a value similar to the DEM in valley floors where groundwater will likely accumulate. To do as such, a calculation was performed to the elevation values of the DEM using the 'Raster Calculator' tool. Where, for the lowest elevation of the DEM, groundwater was kept equal to the DEM elevation. While, for anything greater than the lowest elevation, the pixel value was halved and then added to the minimum DEM value - Equation 13. Creating an input file which mirrored the elevations in the landscape, but where groundwater relief was dampened.

$$z_w = z_{min} + \frac{1}{2}(z - z_{min})$$
 Equation 13

5.2.3.4 Hydrological Boundary Conditions

A final input the model requires is a Hydrologic Boundary file that allows the surface water and groundwater boundary conditions to be specified. This file takes the catchment DEM, and defines all pixels in this file with either values of 11 or 12. All pixels within the DEM will be listed as 11, unless they are specified as a 'river outflow node', which is represented by the value 12. When a pixel is specified as an outflow node, CLiDE will extract the value for surface water flowing across the pixel and record it in a 'Surface Water Discharge' output file - an ascii file which can be examined and manipulated in a text editor. In the case of this study, the river outflow is specified by one pixel at the downstream boundary of the catchment; where the outlet shapefile exists. As the DEM has a 100-m resolution one node was appropriate, but in other cases where DEM resolution is finer, more than one node could also be applicable.

5.2.4 Calibration

5.2.4.1 Flow Separation

When considering a catchment's hydrological system, surface and subsurface water flows should be treated as two separate entities as they account for different portions of a river's hydrograph. Groundwater flows will contribute primarily to the baseflow, which remains reasonably constant with time in comparison to storm flow. Baseflow is defined as the sustained input of water to the river channel from the ground-water system (Focazio and Cooper, 1995). In contrast, surface water flows are those that will be reflected in storm peaks in a hydrograph because of direct surface runoff into the river system. This stormflow will be present during storms, and for a short period of time following the event. A hydrograph can therefore represent purely baseflow, during inter-storm periods, or baseflow plus storm flow when the river is in a time of flood (Figure 21).



Figure 21. Components of a typical storm hydrograph, where quickflow represents stormflow (Erickson and Stefan, 2008).

Due to the fact these two hydrological entities behave in such different manners one must calibrate each process independent of the other. This facilitates a simpler assessment of the influence of each of these water fluxes to the hydrological system as a whole. Additionally, this will negate any issues which may arise from autocorrelation of calibrating more than one variable. The fewer number of variables which are calibrated at a single time, the more confident one can be in the result of the calibration. The CLiDE manual advises the user to calibrate surface flow initially, followed by baseflow. However, for this study, the calibration stage was completed in reverse order. The logic behind this decision was that the baseflow is always present within the system, and the storm flow is,

94

at times, present on top of this. Stormflows values are harder to predict if baseflow values are incorrect as the two are coupled together in times of flood. Therefore, if baseflow is calibrated to a reasonable standard initially, one would hope that the calibration of storm flows following this would be easier than if baseflow values had not been known.

When both ground and surface water flows are turned on, CLiDE produces a surface water outflow file which contains quantitative values of total discharge exiting the catchment at the outlet node or hydrological boundary. It was therefore necessary to perform a baseflow separation to the model's output dataset to observe and analyse the two as separate entities. This was done by a series of steps, following the method by Piggott el al. (2005), an adaptation of the UKIH method (Institute of Hydrology, 1980). This method is based on periods of flow that are assumed to be composed entirely of baseflow. To identify these periods, the modelled dataset is divided into 5 day segments and the minimum values within each segment are selected. The low point for each segment, y_i , is then compared to neighbouring segments - Equation 14. If the value for y_i satisfies the equation, the point is identified as a low flow data point. The daily baseflow contribution to a river is then estimated by linearly interpolating between all the low flow data points which remain. This method allows any erroneous points which might increase the standard deviation of the data to be removed.

$$0.9y_i < min(y_{i-1}, y_{i+1})$$
 Equation 14

To create a baseflow separated time series, where baseflow and surface water flow and two separate entities, the baseflow value at each point must then be removed from the total modelled flow. This same method can then be performed to the gauge data so both datasets are directly comparable. The values for baseflow and peak flow for both the model and the SEPA gauge, were then compared on two separate graphs (Figures 22-26 below) which plotted discharge (m^3d^{-1}) , against time.

5.2.4.2 Approach to Calibration

The following results section will contain a calibration procedure which aims to better tune the catchment model so it is more accurately representing the 96

situation in reality. This calibration will consist of running the model for a given length of time, analysing the outputs, and then determining where the model is performing well and/or poorly. This procedure will then inform the input parameters for the subsequent run, i.e. the calibration runs aim to highlight any errors included in the model parameterisation. The aim of this calibration process is that, after multiple calibrations runs the model will show a better likeness to reality than it did after the initial spin up run. A similar approach is then used for the reach scale modelling detailed in Section 6.3 of this thesis.

5.3 <u>Results</u>

5.3.1 Groundwater Spin Up (1)

The model was therefore initially run with the 'groundwater only' box ticked for a period of 10 years, from 2001 and 2011. This specific time period was chosen to correspond with the calibration period for reach scale modelling. This was because high resolution DEMs are present for the braided reach from 2000 to the present day. This analysis therefore considers this time period, with an aim to calibrate the model from 2001 to 2011 and then to test the model beyond these dates. Additionally, 10 years is estimated as an appropriate length of time to allow a modelled system to equilibrate to a steady state. However, one must note that this is only an estimate, and the time taken to reach steady state groundwater values will vary dependent on a specific catchment's characteristics.

As well as allowing a time period over which groundwater can equilibrate, this initial run served a dual purpose in allowing input files which initially contained zero values to be populated with modelled data. As well as the groundwater file, the NSSS and SMD parameters - both of which directly relate to groundwater - were also populated with modelled data as the system equilibrated to a steady state with time. The final output files from this 10 year spin up were then used as inputs into subsequent runs, where both groundwater and surface water were active hydrological entities.

To determine whether 10 years was a sufficient amount of time to allow the modelled system to equilibrate, the outputs from this run were visualised and analysed using a scatter plot which displayed groundwater heads as a function of

time. When observing this data, one would expect a normal amount of inter annual fluctuation, but an overall exponential trend of groundwater tapering to a roughly constant value - considered to be its steady state.

5.3.2 Surface Water Spin Up (2)

Following the previous 10 year run, a second spin up run was initiated over the same time period, where all input files (bar the three listed above) were the same as the first. Additionally, all model parameters were kept the same to maintain consistency. This second run allowed groundwater values to equilibrate further, as well as commencing a steady state spin up of the surface water system.

5.3.3 Calibration 1 (3)

When analysing the plots which were produced by the Spin Up 2 model run (Figure 22) it was evident that there was a large disparity between the baseflow of the model and the gauge data. Modelled baseflow was shown to be consistently too high, overestimating flow by an average of c. $500,000 \text{ m}^3\text{d}^{-1}$. Guidance from the user manual was then considered, and to try to reduce baseflow the hydraulic conductivity multiplier was increased from 1 to 2.5. The aim of this being to increase the rate of water flow into the river, and increase rates of transport out of the system - allowing more efficient groundwater drainage because of a lesser gradient between high and low groundwater heads, reducing the time lag from recharge to baseflow return (Barkwith and Coulthard, 2013). In addition to this, rainfall interception values were doubled (from 1 to 2) consistently over the course of the year. The aim of this being to reduce the amount of water being directly fed into the groundwater system, again with a hope that this would reduce the baseflow discharge.

For calibration runs, it was determined it would not be effective in terms of time and computation processing power to run 10 year simulations. Model runs were therefore run over a two-year period from the start of 2010, to the end of 2011. Additionally, outputs from the previous model run such as groundwater, NSSS and SMD were used as inputs into the model as these parameters have been 'spun up' and now represent reasonably constant and stable values. Again, allowing for shorter run times over a lesser period of modelled time.

98 5.3.4 Calibration 2 (4)

Despite the fact hydraulic conductivity was increased, baseflow discharge did not fall as anticipated. This demonstrated that a relatively high amount of water was being transported into the river system and the value was not decreasing with time, as was initially hoped. Baseflow discharge, displayed in Figure 23, can be seen to more than double in value - the antithesis of what had been hoped for in this calibration. However, as this simulation was performed over a shorter period compared to the spin up run, one cannot be sure that this effect would not level out with time. Despite this, it would not be effective time-wise to run this simulation infinitely to observe if it eventually levelled out. The peakflow discharge values for the model also increased dramatically, and became more dissimilar to the gauge data. Considering this, it was decided to reverse both of the parameters edited in the initial calibration. The hydraulic conductivity multiplier was decreased from 2.5 to 0.5, and the rainfall interception was changed from 2 to 0.5 for every month. Additionally, every other parameter remained as previously.

5.3.5 Calibration 3 (5)

Reversing the values of the above two parameters had the desired affect and decreased baseflow discharge to a value more comparable with the gauge dataset - see Figure 24. In some months the data values matched up reasonably well, but there was a distinct overall trend that the modelled data flow was set c. 100,000-200,000 m³d⁻¹ too low. When considering the above, changes were made to the model input values whereby hydraulic conductivity was decreased further from 0.5 to 0.35. Another disparity between the baseflow plots, is that the model is displaying a more subdued version of what the gauge data is displaying. The gauge shows more fluctuation from month to month, whereas levels in the model remain somewhat consistent. This shows that the model is not reacting as rapidly and to the same degree as the natural system. To try to bring the model more in line with the observed gauged data, the specific yield multiplier was edited. This parameter determines the groundwater head response to an incoming or outgoing flux of water. Therefore, a groundwater system with a high specific yield will show a smaller response to the influx or removal of water in comparison to a catchment with a low specific yield. Considering this, one would want to lower the specific

yield so the groundwater system had a larger response to an influx of water. For the following calibration, the specific yield multiplier was decreased from 1 to 0.5.

If one then looks to the peakflow plot displayed in Figure 24, it can be observed that modelled peaks are generally lower than those of the gauge data. This same trend was present in the last calibration run, however this time to a much lesser extent. Rainfall interception should therefore have been decreased to allow more rainfall to be transported directly into the river, which should subsequently be reflected in peakflow discharge values. This in turn might also increase the variability of the baseflow, via peaky rainfall input into the groundwater system alongside the decrease in specific yield. Unfortunately, in error this parameter was increased from 0.5 to 0.75 - the opposite of what was desired for this step in the calibration. The following outputs were subsequently reviewed with this fact in mind.

5.3.6 Calibration 4 (6)

The modelled baseflow data were now reasonable well constrained, with both the model and gauge displaying a mean discharge of c. $350,000 \text{ m}^3\text{d}^{-1}$ - Figure 25. However, the baseflow plot from the model outputs was still more subdued than that of the gauge. To try to resolve this, specific yield was decreased further to 0.25, to allow the groundwater to react even more sharply to inflows and outflows of water.

Despite the above error regarding the rainfall interception parameter, modelled and gauge peakflow values appear to match up better than the previous model calibration outputs. Perhaps displaying that the primary control on this value is specific yield as opposed to rainfall interception. Considering these results the rainfall interception was kept at a value of 0.75. The modelled values were still marginally too low, but it was hoped that the decrease in specific yield should act to correct this further.

5.3.7 Output Run (7)

The modelled values of daily baseflow discharge from Calibration 4 (Figure 26) appear marginally more similar to the gauged flows, in comparison to the previous

100

model simulation. Fluctuations in the data are slightly more pronounced than previously, but still not as prominent as the gauge baseflow.

When considering the peakflow values, the model now appears to be producing flows greater than those observed by the gauge - the corollary of what was true for the results of the last calibration. This trend is not entirely consistent throughout the datasets, but is true for the majority of the peaks. However, the amount the model is greater than the gauge is, in most cases, marginal.

At this stage in the calibration process it was decided that the model was producing results comparable enough to the observed gauge data that next stage of model processing could begin. It can be noted here that calibration could have continued for many further runs, tuning the model further in an attempt to more closely to predict the gauge data. However, when considering the scope and timescale of this masters' thesis it would have not been efficient to do so. Additionally, multiple additional 10-year runs could take up to a week at a time to process so the option of completing longer calibrated runs was dismissed.

The next step in this modelling framework was to therefore the link the catchment scale rainfall runoff model, to a reach scale geomorphic change model. In order to do this, the model was run for an additional 10-year period, again from 2001 to 2011, as per its calibrated parameters. The final output run was therefore run with all parameters as per Calibration 4, excluding the hydrological boundary file. This file edited to include four additional outlets nodes which related the catchment to the reach scale model. These nodes spanned the width of the valley floor at the upper extent of the reach scale model. They were positioned in such a way so that any water flowing through that portion of the valley would be accounted for. The outflow from these four nodes was then summed up to represent the total influx of water to be put into the reach scale model at its upper bound. Furthermore, it can be noted that the addition of these nodes meant that the catchment model downstream of this boundary becomes somewhat futile; as water is being extracted and removed from the model mid catchment. Downstream values will consequently be inaccurate and unrepresentative due to this, and are no longer of relevance to the following modelled analysis.

5.4 Figures

Parameter:		Hydraulic Conductivity Multiplier	Specific Yield Multiplier	Rainfall Interception	Time Period (Years)
Run Number	Description				``````````````````````````````````````
1	GW Spin Up	1	1	1	10
2	SW Spin Up	1	1	1	10
3	Calibration 1	2.5	1	2	2
4	Calibration 2	0.5	1	0.5	2
5	Calibration 3	0.35	0.5	0.75	2
6	Calibration 4	0.35	0.25	0.75	2
7	Output Run	0.35	0.25	0.75	10

Table 11. Varying parameterisation for catchment scale modelling (via CLiDE). The four variables that were altered as displayed, all other parameters remained consistent.



Figure 22. Surface water Spin Up outputs. Top plot represents the storm flow hydrograph, and bottom plot represents baseflow hydrograph (these plots are consistent in this from Figure 22 - 26).







Figure 24. Calibration 2 outputs.



Figure 25. Calibration 3 outputs.



Figure 26. Calibration 4 outputs.

5.5 Conclusion

The purpose of this chapter was to obtain a discharge input for the CL reach scale model. As well as completing the objective of this thesis to create and calibrate a catchment scale model of the River Feshie. The rainfall-runoff model was first calibrated over a 10-year period from 2001 to 2011. The work aimed to predict river discharge from rainfall inputs via the CLiDE hydrological model. The model was calibrated using seven simulations, within which various variables were adjusted to allow the model to better represent reality. The discharge prediction from the calibrated model was assessed by comparing to observed flow gauge data for both base and storm flow hydrographs. For the calibrated simulation, both the model and the gauge displayed a mean baseflow discharge value of c. 325,000 m³d⁻¹. Even though subtle variations in the predicted and observed baseflow hydrographs didn't match up perfectly, the model was deemed to be representing the overall magnitude and dynamics of the baseflow relatively well. The same was true for the calibrated stormflow. Although the model was unable to perfectly replicate stormflow peaks, perhaps due to the lack of spatially variable rainfall data within the catchment. Despite this, the final model output displayed peaks of similar magnitudes and frequencies.

6.1 Introduction

This chapter aims to produce a calibrated reach scale model by using input data detailed in the previous chapter. Alongside this, the chapter outlines how addition input data was obtained including a thorough description of how the input sediment data was collected and analysed. As with the catchment scale modelling, a calibration process with subsequent stages wa completed. Yearly DEM outputs from the final model run were then analysed using geomorphic change detection software.

For the reach scale modelling portion of this thesis the model CL is used. CL was used as opposed to CLiDE as it represents a stripped back version of the catchment scale model. They share the same base code, but differ in the fact that CL is not able to model groundwater flow, and groundwater discharge to rivers. As this process is already calibrated within the catchment scale system, it is thought to be appropriate that it was excluded from the reach scale model. The benefit of CL is that has quicker run times, and less input parameters and files to calibrate.

To initiate the process of geomorphic change along a braided reach of the River Feshie, a discharge input file is needed which was obtained from the catchment scale model. Obtaining this parameter from CLiDE means that the reach scale modelling is being forced by the behaviour of the catchment scale model, allowing for an encompassing analysis of the River Feshie catchment as a whole. This data file contains daily discharge which is supplied to the reach scale model at its upper bound. This flow of water through the modelled system, coupled with CL's ability to represent sediment transport processes aims to simulate representative quantities of geomorphic change along the c. 2 km braided reach this study focuses on.

6.2 <u>Methodology</u>

6.2.1 Digital Elevation Models

CL firstly requires a DEM of the reach extent one wishes to model, and a bedrock DEM of the same area. For the initial spin up run a DEM of the reach, obtained from a field survey in early summer 2000 was used (Brasington et al., 2003). As

104

per Table 2, one can see that there have been multiple surveys of the braided reach of the River Feshie - the extents of which vary spatially. Therefore, to allow for a direct geomorphic comparison, a smaller clipped area which is present in all the DEMs will be used when comparing DEMs- the extent of the 2002 DEM. However, the original extents of the 2000 DEM were used in the modelling process to allow for as much observed information to be retained as possible. The outputs from the model were then cropped when subsequent analysis was taking place.

The 2000 DEM was obtained via photogrammetric methods and contained a final output resolution of 0.5 metres (Brasington et al., 2003). Despite the fact that resolution would be optimum for the geomorphic change analysis, when the size of the area is considered this produces too high a number of cells to be processed by CL. Anticipating this problem, the resolution of the model was coarsened to 2 metres. At this resolution the model was running at a rate of half a model year per day, meaning a 10-year simulation would take 20 days to run. Unfortunately run time can only be decreased by increasing the resolution or decreasing the extent surveyed. Therefore, for the initial 10 year spin up the resolution was coarsened further to 4 metres - the implications of this decision are discussed in detail in Section 7.2.2.1.

To allow modelled geomorphic change to occur freely without any resistance from bedrock, the bedrock depth for the reach scale model was set to 10 metres below topography. As the river is characterised by a highly dynamic system which has the ability mobilise large amounts of sediment which have accumulated in the valley floor over thousands of years, this value could in fact be accurate. However, with no geophysical information to make an informed decision, the value is purely theoretical and has been chosen so that the model can perform to its full capacity with regard to sediment transport. So that it is not inhibited by a bedrock depth which is too shallow, meaning the system will run out of potential sediment to transport. However, there could be additional implications to this decision which are discussed in detail later in Section 7.2.2.2.

1066.2.2 Bulk Sediment Analysis

6.2.2.1 Introduction

Caesar requires information on sediment size distribution, which can include up to nine separate grain sizes. During the June 2018 field campaign, a bulk sediment analysis was carried as per the method detailed in Hoey (2014). For this, two sites - one at the upstream end of the reach, and the other at the downstream - were sampled (Figures 27-29). The site chosen for the first analysis was on the upstream end of a lateral bar on the western bank of the river. At high flow this geomorphic unit would have represented a mid-channel bar, as a dry river channel was present cutting across the bar's western edge. The site chosen was therefore selected where this river would be flowing at a time of high flow, but in its current state represented the upstream end of a lateral bar. This site was thought to accurately represent the overall sedimentary characteristics of the bar as a complete geomorphic unit.

When considering a bulk sediment analysis for a gravel bed river, the size of the sample is important. If one wishes to obtain a sample which is reliable and meaningful in an environment that includes a wide grain size distribution from cobbles to silt, sample size can be relatively high in comparison to an environment only containing fine sediment (< 4 mm). Additionally, a larger sample size is also preferable when dealing with more poorly sorted material (Hoey, 2004). Several guidelines exist which relate to optimum sample size. Church et al. (1998) state that the mass of the largest particle should be equivalent to c. 0.1% of the total sample mass. However, this can lead to unmanageably large sample sizes when one is considering sampling modern fluvial gravels. Consequentially, Church et al. (1987) recommend using a 1% criterion if the maximum particle size of a unit is greater than 32 mm. This criterion was applied to sampling on the River Feshie.

6.2.2.2 Field Data Collection

For the analysis, sediment was excavated from a 2 m^2 area in two layers - a surface, and a subsurface. The depth of each layer was defined by the length of the intermediate axis (b axis) of the largest clast found in that layer. As per Church et al. (1987) the mass of material (M_s) excavated should be equal to 100 times the mass of the largest particle (M_{Max}) when surveying coarse material. However, to

avoid overly large sample sizes they have recommended a cap for M_s of 1000 kg, as when sample sizes exceed this volume the increase in reliability is not comparable to the increase in manual time and effort.

The following methodological description details how sediment was removed, weighed and categorised for a single bulk sediment sample. This process was replicated four times, for both upstream and downstream, surface and subsurface sites.

Sediment was excavated by hand and measured using a Wolman template (Wolman, 1954), which categories sediment by half phi sizes. Sediment would be placed on one of two tarps, depending on if its intermediate axis was less than, or greater than 90 mm/6.5 phi. After one ton of sediment was excavated, 10% of the total - in this case 100 kg - was taken from the < 90 mm pile and treated as a subsample (Hoey, 2014). This subsample was extracted randomly and without bias. The material on the tarp was divided into half, and then into guarters, and from this a random 100 kg was taken. This subsample was then sub-dived into all lower size classes within the phi scale, down to 4 phi/16 mm. Any material smaller than 16 mm was then placed into a separate pile, a c. 4 kg of which was then taken away for sediment sieving in a lab environment. The data obtained from this subsample was then assumed to be representative of the total amount of sediment < 90 mm. As per Hoey (2014) this method is said to produce errors in keeping with the scope of this study. This method was therefore preferential in terms of labour and time spent in the field. Material > 90 mm was weighed during the excavation process as fewer particles amounted to the same weight, making them easier to weigh. The largest size class surveyed on the River Feshie was 8 phi or 256 mm.


Figure 27. Upstream bulk sediment sample size. A) Photo of unit sampled, looking downstream

from other bank of the river. B) Photo of unit sampled looking downstream. C) Aerial view of unit makeup - exact location sampled.





Figure 28. Downstream bulk sediment sample site. A) Example of excavated sediment on tarp > 90 mm pile. B) View of sampled unit, looking upstream. C) View of sampled unit looking downstream.

Location of Bulk Sediment Sample Sites



Figure 29. Location of upstream and downstream sediment sample sites displayed on the 2018 aerial imagery of the modelled reach.

112 6.2.2.3 Lab Analysis

The following lab analysis first consisted of drying the bag of material brought back from the field site in an oven. The material was dried at a temperature of 200°C, for a period of 24 hours. Prior to this, the material was weighed and a wet weight was noted. After the material had been dried a second weight was noted and the two compared. This processes allows for the drying and removal of moisture and organic matter which is liable to cling to smaller grains and unrepresentatively increase weight values. After the material was dry, sediment sieves were used to divide the material into six smaller size classes, and a < 2mm residual class. Hand sieves were used for size classes down to 4 mm, and an Octagon Digital electromagnetic sieve shaker was used for all smaller size classes. It must be noted that during this analysis it was not possible to obtain a 11.2 mm sized sieve (-3.5 phi). In place of this, a 11-mm sieve was used. Ideally, all size classes should be even and follow the phi scale however, due to the fact this would have a fairly insignificant influence on the final dataset it was deemed acceptable to make this substitution. The following calculations that were then used to analyse the data were altered to account for this change in size.

6.2.2.4 Statistical Analysis

Following the above field and lab data collection, a dataset was now present which contained eight or nine values for the mass of sediment retained in the field, and seven additional weights for dry masses retained in the lab. These data values were input into an excel spreadsheet for statistical analysis, where sample sizes were scaled by the volume of the median-sized grain - see Appendix 1. This was done so that results could be applied to any log-normal grain size distribution (GSD). The dataset also produced statistical values of bias and precision, which represent the reliability of estimated percentiles of bulk GSD. Bias, in this instance, is defined as the systematic deviation between sample and population values, and is quantified as the difference between the mean sample value of the pth percentile (ψ p), and the equivalent population value. Precision is defined as the systemates in the face of random sample-to-sample scatter, and is quantified as the between sample standard deviation of ψ p. In additional to this, the statistical analysis also included equations which produced values for the sample size necessary to obtain a specific precision and bias (Ferguson and

Paola, 1997; Bunte and Abt, 2001). This analysis primarily focuses on the precision statistic, as when one is considering a large sample size bias is generally 0.

6.2.2.5 Results

Graphs showing the precisional error on the main parameters which characterise the GSD for each site are shown in Figures 30 & 31 below. Where the parameter represents the nth percentile of the dataset, i.e. parameter 5 is equivalent to D5 of the dataset. Which in the case of the upstream surface 5.04 mm. The error brackets denote the precisional error (in mm) on each of the parameters listed. It is clearly demonstrated that as the parameter increases, so does the precisional error. This is because there is a much higher confidence level related to smaller grain sizes when a consistent amount of sediment has been excavated. To decrease the precisional error for D84-95 a much larger sample size would have been necessary, on the order of 10's of tonnes - see Appendix 1 for more detailed statistics. Appendix 1 contains the raw data for all four sites, and all additional statistical parameters which were calculated for the data.



Precisional Error - Downstream Sub-Surface



Figure 30. Precisional error at downstream site - surface and subsurface. Error bars denote how constrained a parameter is.



Figure 31. Precisional error at upstream site - surface and subsurface. Error bars denote how constrained a parameter is.

6.2.2.6 CAESAR-LISFLOOD Input

For input to CL, all sediment statistics must be defined as per the depth of the active layer. This is defined as the layer of material where a continuous interchange of particles occurs between a sediment bed and the flow over it during periods of bedload transport. This surface layer provides the source for the bedload, and thus influences its transport rate and composition. The effect of this layer is to act to equalize the mobility of grains contained in the substrate, and is achieved by a combination of two effects. The first being the effect of particle sheltering (e.g. Einstein (1950); Proffitt and Sutherland (1983)) and the second being the evolution of a bias in the surface layer toward coarser grains relative to the substrate. As per Hoey and Ferguson (1994)'s sediment routing equation (15), which

considers individual grain size classes, the thickness of the active layer can be defined as L_a (L).

$$(1-\lambda)\frac{\partial L_a F_i}{\partial t} = -\frac{\partial (q_T p_i)}{\partial x} + E_i \left(\frac{\partial q_T}{\partial x} + (1-\lambda)\frac{\partial L_a}{\partial t}\right) \qquad Equation \ 15$$

Additional parameters include, *Fi*, *Pi*, and *Ei* which denote proportions by volume of material in the ith size class in the active layer, bed load, and exchange sizes, respectively. The solution to Equation 15 depends on the definition of the active layer thickness, L_a . In this instance, L_a is defined as a constant function of the active layer grain size, the default value being 2D₈₄. Note that the dependence of L_a on surface D₈₄, and hence on F_i , makes the solution of (15) necessarily iterative.

The percentage of mass retained for each GSD site was then compiled into a spreadsheet which coupled the surface and subsurface measurements for both sites to produce averaged 2D₈₄ values for both - as per Table 11. These values were then subsequently used to relate the grain size data obtained to the depth of the active layer. However, CL only allows a GSD for a single site to be input into the model. It was therefore decided that the upstream site should be used, as this is where the sediment data is initially input into the model. Additionally, the site where data were obtained for the upstream GSD analysis was within the hydrological boundary nodes created for the Output run of the catchment scale model.

116

Grain Size (mm)	256	181	128	91	64	45	32	23	16	10	8	9	4	3	2
Grain Size (phi)	8-	-7.5	-7	-6.5	-9	-5.5	-5	-4.5	-4	-3.3	-3	-2.5	-2	-1.5	-1
Sub-surface (D)	0	2.546	3.932	9.91	13	13.636	13.636	10.818	8.636	6.5	2.935	4.332	3.537	3.461	3.121
Surface (D)	0	5.071	13.493	16.143	31.574	14.953	7.701	5.134	2.653	1.089	0.55	0.664	0.412	0.302	0.261
2D84 (D)	0	4.367	10.825	14.404	26.392	14.586	9.357	6.72	4.322	2.598	1.215	1.687	1.284	1.183	1.059
Cum % finer (D)	100	95.633	84.808	70.404	44.012	29.427	20.07	13.35	9.028	6.429	5.214	3.527	2.242	1.059	0
Sub-surface (U)	0	2.591	4.093	10.876	29.919	13.437	8.958	7.704	5.554	4.035	2.387	3.14	2.603	2.499	2.20288
Surface (U)	1.754	9.978	10.039	14.524	18.783	11.325	8.977	7.458	4.281	3.024	1.719	2.452	2.079	2	1.60878
2D84 (U)	1.522	9.003	9.254	14.042	20.252	11.604	8.974	7.49	4.449	3.157	1.807	2.543	2.148	2.065	1.68719
Cum % finer (U)	98.478	89.475	80.221	66.179	45.926	34.323	25.348	17.858	13.409	10.252	8.444	5.901	3.753	1.687	6.44E-15

Table 12. Grain size data collected for Surface and Sub-Surface layers of two sites - Downstream and Upstream.

The data displayed in Table 12 splits up the sediment load into 15 different grain size proportions, however CL only accepts nine different sediment data inputs. The data were then combined into nine sediment populations, where multiple size brackets were summed and listed within an averaged size bracket. Smaller fractions were preferentially combined into one, as those were deemed less significant. The resultant grain size figures which were used in final CL model are displayed in Table 13 below.

	Grain Size	Proportion
	(m)	
Size 1	0.0029	0.0591
Size 2	0.0068	0.0435
Size 3	0.013	0.0761
Size 4	0.0275	0.1646
Size 5	0.045	0.116
Size 6	0.064	0.2025
Size 7	0.091	0.1404
Size 8	0.128	0.0925
Size 9	0.2185	0.1053
		SUM: 1

Table 13. Averaged grain size proportions input into CL.

The GSD shown in Table 12 represents the material which the reach is made up of, and is currently existing within the modelled area. Unfortunately, there was no information available which quantified a sediment influx into the reach scale model and measuring such a parameter is out with the scope of this thesis. It was therefore decided that the sediment being output at the end of reach should be recirculated back into the upstream input. The implications of this will be discussed in more detail in the following discussion Section 7.2.2.3.

6.3 Results

6.3.1 Spin Up (1)

Similar to the catchment scale model, Caesar requires a 'spin up' in order to equilibrate to a steady state system. This is relation to the sediment transport parameter, which is liable to have initial fluctuations that are not generally representative of the natural system. Depending on the makeup of sediment in a river's reach, a series of floods might have to pass through the landscape before the sediment load incorporated in the model is broadly representative of reality. For example, there might be a spike in the proportion of fine sediment initially output as the river is preferentially transporting this (Peakall et al., 1996). Additionally, as sediment is being recirculated the system needs time for this process to equilibrate as multiple sediment pulses pass through the system. This spin up therefore needs to be completed to obtain representative sediment outputs from the model's downstream outlet node.

The initial spin up run took the DEM from 2000 and ran for over 7 years, theoretically representing the system in spring 2007 when the run had completed. However, CL purely states the amount of time the model has run for and is not related to inputs at a specific date. As one can see from Figure 32, cumulative sediment output from this run is less for the same discharge for the first 2.5 years of the model run. This implies that in this period the system was preferentially transporting fine material before it began to equilibrate and larger particles were transported as fines were removed. One can also observe the evolution of the sediment transport system in Figure 33, where total sediment output increases with time and better reflects the magnitudes of storm waves travelling through the system.

After the sediment parameter of the model had been reasonably stabilised other model parameters were investigated. The aim of this calibration process was to tune the model to a stage where it would represent similar magnitudes of geomorphic change to what was happening in reality. With an additional aim of being able to represent the general appropriate locations, and similar characteristics of geomorphic change. Such as similar areas of erosion and aggradation, and comparable rates of bank migration. These comparisons were

made both qualitatively and quantitatively in ArcGIS and via GCD software (Wheaton et al., 2010).

Geomorphic change is shown in Figures 34-40 on reach scale maps of elevation difference, where colours of red and green show areas of erosion and deposition respectively. These 'Elevation Difference' maps display the difference between the initial input DEM and the final DEM after geomorphic change has occurred. For simplicity change is only denoted as positive or negative regardless of magnitude, as the analysis focuses on broad areas of change as opposed to precise qualitative values.

For the initial spin up run the parameters were set in accordance to advice from the CL Manual (Coulthard et al., 2013). The specific parameters that will be key to this study are included in the sediment tab. All parameters which have been altered during the calibration process are displayed in Table 14.

A parameter which this calibration focuses on is the lateral erosion rate - a key component of braided river modelling. The CL manual states that values of 0.01 to 0.001 are appropriate for the lateral erosion rate when considering braided rivers. A value of 0.005 was chosen to represent an initial estimate. However, when running the model for multiple years it became evident that this was not an appropriate value with regard to the morphological evolution of the planform. A realistic representation of the system existed initially, however this realism degraded with time. Islands and banks were eroded rapidly, and the channel filled with sediment - producing a flattened landscape. After some investigation, it became evident that the lateral erosion rate was set too high, and the model was acting to always try and erode banks to create a flatter, wider channel - Figure 34. The parameter was therefore decreased for subsequent calibrations.

Par Run Number	ameter: Description	Max Erode Limit	In Channel Lateral Erosion	Lateral Erosion Rate	Number of Passes	Minimum Discharge (Q)	Time Period (Years)
1	Spin Up	0.02	20	0.005	100	0.04	7
2	Calibration 1	0.02	20	0.0001	100	0.04	11
3	Calibration 2	0.075	30	0.0001	50	0.15	7
4	Calibration 3	0.01	25	0.0002	100	0.04	11
5	Calibration 4	0.01	25	0.001	100	0.04	11
6	Calibration 5	0.01	25	0.0005	100	0.04	3
7	Calibration 6	0.01	25	0.0005	100	0.04	11

 Table 14. Summary of parameters altered in calibration process for reach scale model.



Figure 32. How discharge (blue) and sediment output (orange) vary with time.



Figure 33. Discharge verses sediment output. Colours correlate to sediment output before and after a 2-year cut off, i.e. the initial sediment pulse verses once the system has run to a steady state.



Figure 34. Spin Up model run outputs for both a high (after 1 year of model time) and low (after 5 years of model time) flow events. Progression of model through time displays increased erosion and channel flattening.

6.3.2 Calibration 1 (2)

For the following calibration run all parameters remained the same bar the lateral erosion rate, which was set to 0.0001. The outputs from this run confirmed the above assumption that this was the parameter was the primary influence on the geomorphic behaviour of the model. The outputs of this model showed lateral erosion rates more comparable to reality than in the previous model, but with time these still appeared higher than normal. Additionally, at low and medium

flow conditions the modelled channel represented more of a meandering single channel river as opposed to a braided system - Figure 35.



Figure 35. Calibration 1 model outputs. High flow event occurred after 6 years of model time, low flow event occurred after 4 years of modelled time. Model displaying appropriate braiding behaviour at high flows but represents a single thread channel at low flows.

6.3.3 Calibration 2 (3)

The following calibration therefore hoped to increase channel divergence and braiding intensity at lower flow conditions. Additionally, it aimed to keep the influence of these parameters more consist with time. To do this, two scenarios were modelled - Calibrations 2 and 3. The first, Calibration 2, was set to run with

the recommended parameters given by Ziliani et al. (2013) who used CL to model the evolving geomorphology of a braided reach of the River Tagliamento, Italy. An explanation of this study and its comparison to the River Feshie is detailed in Section 7.2.2.5. The calibration process completed by Ziliani et al. (2013) is the most rigorous of its kind when considering reach scale modelling of braided river geomorphology using CL. It was therefore deemed appropriate to use the parameters defined in the final calibration stage of this analysis as a starting point for the River Feshie analysis, a reach of similar characteristics.

As per Ziliani et al., (2013), the max erode limit was set to 0.075, compared to 0.02 in the previous run. Additionally, the in channel lateral erosion was increased to 30 and the number of passes decreased to 50. In keeping with Ziliani et al. (2013), the Min Q value was set to 0.15 to keep consistency in using these parameters. However, for grid cell sizes of 4 m² the model does not recommend Min Q to be set to such a high value. Despite the fact this method may appear abstract, the purpose of this calibration stage was twofold. Firstly, to supply the model with parameters which have been confirmed to be appropriate for a divergent and braided river. Secondly, to give insight into the differences and similarities between the two rivers, as well as demonstrating how specific parameterisation must be despite having the same broad characteristics.

6.3.4 Calibration 3 (4)

Alongside this model run Calibration 3 was started on a different desktop machine in run in parallel with Calibration 2. The parameters for this mirrored that of Calibration 1, but with some slightly altered parameters. The in channel lateral erosion was increased slightly from 20 to 25, as was the lateral erosion rate from 0.0001 to 0.0002. The purpose of this was to alter the parameters which generally define braiding and divergence, with an aim to make the channel more dynamic and representative of the natural braided system.



126 Calibration 2, 3 Years Model Time Calibration 2, 7 Years Model Time

Figure 36. Calibration 2 model outputs. 3 years of model time (low flow conditions), and 7 years of model time (average-high flow conditions). Modelled system is largely unrepresentative of reality - isolated pools of water.



Calibration 3, 4 Years Model Time Calibration 3, 8 Years Model Time

Figure 37. Calibration 3 model outputs. 4 years' model time (average-high flow conditions), and 8 years of model time (low flow conditions). Model lacks divergent braided behaviour with time.

6.3.5 Calibration 4 (5)

The outputs from the above two runs yielded different results. Calibration 2 produced outputs relatively unrepresentative of the natural system. The channel was primarily dry with focused patches of deep water - Figure 36. This is likely the result of parameters which were not tailored to the specific site and acts as an example to display how site specific the process of parameterisation can be.

For Calibration 3, the modelled system lacked the characteristic divergence of a braided system and was performing exceptionally poorly at low flow where the

system was represented by a single channel or isolated pools of water. It was therefore decided to edit the lateral erosion rate parameter in Calibration 4 to try and simulate behaviour more similar to that of a natural braided river. This parameter was increased to 0.001 and all other parameters remained as of Calibration 3.

6.3.5.1 Persistent Low Flow Issue

In addition to this it was becoming evident that in all model runs water depths were much lower than they should have been in reality. The amount of water entering the reach at its upper bound did not equal the amount being output at its lower bound - displayed through the fact that the channel was consistently under predicting depth values. This trend, however, wasn't consistent throughout time. Model outputs showed that during high flow events the model was able to better predict braid patterns, but the landscape at low flows was entirely unrepresentative of the natural River Feshie system. However, over both high and low flows it was evident that the model appeared to be draining a substantial proportion of the river's discharge into the subsurface. At this stage in the calibration process it was unclear as to how much subsurface drainage was being influenced by the parameterisation of lateral erosion and braiding intensity. It was therefore decided that the model calibration should continue to try and reproduce the general characteristics of the braided River Feshie, but generally focusing on the channel characteristics during storm flow events. This decision was made based on the work by Ziliani et al. (2013) who aimed to model geomorphic change of a braided river reach by only modelling the fluvial system where discharge was greater than 100 m³s⁻¹ - the approximate discharge needed to initiate the movement of sediment in the Tagliamento River. The potential reasons for this subsurface drainage phenomenon, and the implications of calibrating the geomorphic model whilst focusing on storm flow events are discussed in more detail in Section 7.2.2.2.



Figure 38. Calibration 4 model outputs. High flow event occurred after 6 years of model time, low flow event occurred after 4 years of model time. Unrepresentative behaviour at low flows, and too much erosion of channel banks creating a flattened landscape.

6.3.6 Calibration 5 (6)

At a very low lateral erosion rate the modelled river system represents a single channel which has limited to no divergence. However, when lateral erosion is set too high the model erodes banks too rapidly and the landscape fills in and becomes flattened. This was the case for Calibration 4 (Figure 38), much the same to the Spin Up run, lateral erosion was set too high and the floodplain was rapidly flattened out as the result of an unnaturally high lateral erosion rate. This

worsened the effects of drainage into the subsurface, as water could easily access and drain into the entirety of the floodplain. The outputs of this model run therefore show a flattened landscape with unrepresentatively low values for water depth - at low and high discharge values. Considering the broad extremes which appear to exist for both too high, and too low values of lateral erosion, an adjusted rate of 0.0005 was subsequently chosen for Calibration 5 as an aim to find a middle ground between the two scenarios.



Calibration 5, 1.5 Years Model Time Calibration 5, 3 Years Model Time

Figure 39. Calibration 5 model outputs. 1.5 years of model time (average flow conditions), and 3 years of model time (low flow conditions). Braided behaviour being adequately represented at higher flows, but lacking likeness at low flow conditions.

The outputs from Calibration 5 are those which represent the closest likeness to reality in comparison to previous model runs. This is primarily the case for high flow events, where the channel can be seen to be exhibiting the dynamic channel switching behaviour observed in reality. As well as producing realistic rates of erosion and deposition. However, Calibration 5 only ran for a period of 3 years as it was run on a standard desktop computer with suboptimal processing power. Therefore, for Calibration 6 the same parameters were used but the model was run on a higher-powered computer.



Calibration 6, 2 Years Model TimeCalibration 6, 7 Years Model Time

Figure 40. Calibration 6 model outputs. 2 years of model time (average flow event), and 7 years of model time (average flow event). Model working well for initial few years of run time, system degrades with time as landscape erodes and flattens.

Figure 40 displays the results of this model which ran for the maximum amount of time which is 100,000 minutes - equating to approximately 11 years. It became evident that with time the model parameterisation worked less well. As time progressed, the DEMs the model was producing were less representative of reality. The model initially represented a braided system with appropriate amounts of erosion and deposition (as displayed by model outputs after 2 years). However, as the system progressed and multiple floods had progressed through the system, banks were entirely eroded by the river and the channel was infilling resulting in a flattened landscape (shown by model results after 7 years) - similar to the Spin Up (Figure 34). Despite the fact model parameterisation was not at an optimum level, it was decided that the calibration process had reached as far as it could within the timeframe of this thesis. The following chapter will therefore consider and discuss the geomorphic model at this stage in the calibration process, noting how the parameterisation process could progress had the research had longer timescales

6.4 Geomorphic Change Detection

6.4.1 Overview

The final calibrated model, as per the parameters included in Calibration 6, was then used for a subsequent geomorphic change detection analysis alongside observed survey data. This analysis aimed to address the second objective of this thesis which was to quantify the morphological change of the c. 2 km braided reach. As was observed from the outputs of the final reach scale modelling, for the first few years of model run time, output broadly represented a system similar to the River Feshie in times of flood - with regard to fact it the model is simulating a braided river of similar dimensions. However, it is then necessary to compare these results to the DEMs of the reached surveyed as observed data to see how the model directly compares to reality. Key questions concerning this comparison are:

Is geomorphic change happening in similar areas in the model verses reality?
 Is the model representing similar volumes of geomorphic change to reality?

DEMs from the model outputs, and observed survey data can both be displayed in ArcMap. CL outputs files at a given time increment - in the case of this study 1000

minutes was used. The model outputs files which give values of: water depth; topographic elevation values (the evolved DEM); and the elevation difference (from the DEM at a given time-step, to the one input at the start of the model run). These files can be viewed and analysed in ArcMap to give both qualitative and quantitative results of change. To view change, various DEMs can be compared. DoDs where produced for each successive time increment for both the model and observed data. This was done by subtracting the older DEM from the more recent DEM to create a map of the reach which had values of positive and negative elevation change. The resultant DoD can then also be compared visually in Figures 41-46 denoted by the title 'Modelled DoD', where red colours denote surface lowering and blues represent surface raising.

Observed DEMs from 2000 to 2007 (excluding 2001, Table 2) were then compared with model outputs extracted to correspond to the dates of observed data. Here, a direction comparison was made between the model and observed data each time increment where an observed DEM existed (Figures 41-16 - 'Direct Comparison').

The third subplot contained within the following figures was created with Wheaton et al. (2010)'s GCD software (see Section 2.4.1). This is thought to be a more informative tool to compare change, as survey error metrics can be incorporated into the DoDs. For the case of the observed DoD, which were derived via an RTK-GNSS survey, a constant error threshold of 0.1 m was applied across the DoD as per (Williams, 2012). A constant error surface was created for each individual DEM. A probabilistic threshold of 80% (which incorporates this error surface) was then applied to the DoD comparison.

This method was used for the comparison of observed data only, as there is no way to quantify and determine the potential errors associated with the CL modelling. For this analysis, the assumption was therefore made that all modelled DEMs were accurate, as per their outputs from CL. Because there is no additional information regarding the location and magnitudes of errors which may be included in the model it would be unrepresentative to include this in a GCD analysis - despite the factor such errors might exist.

134 6.4.2 Results

The following Figures display the progression of geomorphic change from 2000-2007 as per observed survey data and model outputs.



Figure 41. 2002-2000 change detection for (A) observed survey data, (B) CL model outputs, (C) and a direction comparison of modelled to observed data for 2002.



Figure 42. 2003-2002 change detection for (A) observed survey data, (B) CL model outputs, (C) and a direction comparison of modelled to observed data for 2003.



Figure 43. 2004-2003 change detection for (A) observed survey data, (B) CL model outputs, (C) and a direction comparison of modelled to observed data for 2004.



Figure 44. 2005-2004 change detection for (A) observed survey data, (B) CL model outputs, (C) and a direction comparison of modelled to observed data for 2005.



Figure 45. 2006-2005 change detection for (A) observed survey data, (B) CL model outputs, (C) and a direction comparison of modelled to observed data for 2006.



Figure 46. 2007-2006 change detection for (A) observed survey data, (B) CL model outputs, (C) and a direction comparison of modelled to observed data for 2007.

Regarding the Modelled DoDs, initially geomorphic change can be observed to be clearly linked to specific locations and geomorphic units, and appears representative and consistent (Figure 41). However, this ability to visually observe change, i.e. removal of sediment on the outer edge of a meander bend, degrades with time as the outputs from CL become distant from the situation in reality consequentially meaning DoDs look patchy and indistinct. From 2003 onwards Modelled DoDs appear very subdued and smoothed out, and by comparison one can see this is not the case with the observed DoDs. One can also look to the direct comparison between the model and reality to see precisely how much the two differ, where this difference becomes more pronounced with time.

The thresholded DoDs show what is theoretically the change that has occurred in reality. Where it can be observed that the greatest change has occurred on the eastern half of the modelled reach. This is represented by the model initially to some extent, but this matching of results is not consistent with time. The threshold DoDs give an indication of how subtle or substantial change over time has been. For example Figure 43 and 45 display less change over their time periods in comparison to previous years. It could potentially be argued that this is displayed in the model, but it is likely that this is just a result of the model displaying a subtler change with time regardless of what is occurring in reality.

6.5 Conclusion

The discharge outputs from the final CLiDE run were used, alongside bulk sediment data, as inputs to stimulate geomorphic change in CL. The model was parameterised over one spin up, and six calibration runs. After the final calibration stage the model could simulate the inherent behaviour of a braided system well at high flows, for the initial 1-2 years of model run time. However, once multiple floods were simulated, CL outputs degraded for both high and low flows. This outcome highlights the potential difficulties concerning modelling braided river morphodyanmics using CL.

The morphological change which occurred as a result of model simulations was quantified. This was done using DoDs for both observed (surveyed) and modelled DEMs, as well as a direction comparison between the modelled and surveyed DEMs for a set time increment. This quantification of geomorphic change was largely influenced by how well the model was behaving; as the model's affinity to reality degraded with time, so did the representation of appropriate geomorphic change. Initially, observed and modelled geomorphic change took place in similar areas, and was represented by similar magnitudes. However, after multiple years of model run time the observed outputs did not correlate well with the geomorphic change the model was exhibiting.

7.1 Topographic reconstruction and change analysis

7.1.1 SfM

7.1.1.1 Target distribution

The conclusions draw from the GCP analysis are similar to those found by Ves' (2016) work on the Swindale Valley in Cumbria. This analysis considered an area approximately 500m² in size, containing 31 ground control targets spaced in a defined hexagonal layout. This is a smaller area than the Feshie analysis, and has more regimented layout of targets than those of the 2017 Feshie campaign. Ves' analysis considered 3 distinct configurations (1) edge, (2) centre, and (3) random and performed 7 iterations for each, yielding 21 different georeferencing solutions. A subsequent analysis of these GCP test scenarios showed that the best layout solution was using either random or edge configurations. Results also showed that the precision of the dataset passes 1 'Ground Sampling Distance' (GSD) for the edge and random configurations after using at least 19 GCPs, but for the centre pattern, precision only passes this threshold when all 27 GCPs are used.

These results further strengthen the conclusion that the configuration of GCPs has a high influence on their accuracy and confirms that positioning targets at the edges of a study area and in a broader random distribution is the best protocol to follow. Ves' work used a very regimented approach to deploying targets. This is in comparison to target placement on the River Feshie site, which had more of a broad encompassing layout. With comparable error statistics, it can be stated that it is perhaps not necessary to position targets in a defined and restricted layout. Additionally, this is not always feasible in a complex field area.

Tonkin and Midgley (2016) carried out a similar analysis but to a greater scale, where a total of 101 GCPs were deployed. As with this thesis, various test were carried out based on number and configuration. Results of this work stated that DSM quality was significantly improved when more than four GCPs were used, but errors did not significantly decrease below this threshold when an appropriate spatial configuration was used. Additionally, they showed that for any given number of GCPs, errors were higher when GCPs were clustered in the same spatial

area. This was shown by a correlation between vertical error and the distance to the nearest GCP. Therefore in order to obtained the lowest error values, GCPs should be broadly spaced across the study area.

7.1.1.2 Check Point Bias

An additional parameter which wasn't incorporated in this thesis' analysis was the influence of a checkpoint location bias. This phenomenon would be most influential for test scenarios were GCPs are located based on a set spatial configuration. For example, if GCPs are located exclusively at the edges of a study area all the CKPs will be located in the centre. As stated in previous methodology chapter, CKPs and their associated errors are checks on the performance of a given GCP configuration. Prompting the question, are CKPs which are spatially biased to an opposing area appropriate for representing the accuracy of GCPs in a given scenario? The answer to this question is unknown, and therefore were this work to exclusively focus on this portion of topographic analysis further tests should be carried out. These tests might include CKPs which have fixed locations across the entire study area, while altering GCP locations much as this work has done. A comparison of these additional tests to the results obtained in this study would answer the question, does CKP location bias influence errors on GCPs?

7.1.1.3 Alternatives to use of Ground Control Points for georeferencing

Until recently, ground control has generally been used to georeference imagery, define camera calibration parameters, and remove any artefacts of optical distortion which occur in a topographic produce (Carbonneau and Dietrich, 2017). However, in recent years, an alternative georeferencing method has emerged which is becoming increasingly popular as technology evolves. This method is known as Direct Georeferencing (DG) and is a method where control is provided through measurements of camera orientations alone (Förstner et al., 2013). The primary benefit of this method is that it allows for aerial survey over hazardous or inaccessible terrain. Precision measurements from this survey method have been seen to be on the order of 0.1m, where a survey-grade GPS has been synchronised with image capture. However, most current consumer-grade UAVs are only equipped with low-quality GPS antennas meaning that direct georeferencing is not feasible. Despite this, higher quality GPS systems are increasingly becoming
available as options installed on UAVs (e.g. the new DJI Phantom 4), meaning that in the future this technology is likely to become commonplace. Consequently, it will be key to understand the differences between survey performance when using GCPs versus direct georeferencing (Figure 47). This will be integral to optimising future survey strategies aimed at quantifying topography and topographic change (James, Robson, and Smith, 2017).

A study by Carbonneau and Dietrich (2017) aimed to assess the performance of DG, in comparison to the use of GCPs. Their findings concluded that DG workflows result in higher levels of errors in comparison to current SfM GCP surveys. DG outputs generally resulted in a doming effect on the end topographic product, as well as other surface deformations. Whilst acknowledging these consistent errors, the study developed a novel approach of characterising them so resultant products could still be of use while considering their associated errors. The study concluded that DG technologies still pose significant challenges as well as opportunities. When considering the cost of technologies, DG poses many advantages in comparison to costly and labour intense RTK-GNSS surveys. However, they are limited in their applications due to the level of accuracy they can currently produce. Additionally, when one considers the cost and labour involved in SfM surveys with GCPs, the minimal additional labour becomes preferable to produce lower error statistics. Despite this, if the surface deformation errors currently associated with DG can be rectified the technology could provide a key development in automated data collection. With the right development, this form of data acquisition could allow for high quality, meso-scale elevation data to be obtained at a low cost.



Figure 47. Accuracy of GCPs verses Direct Georeferencing (James at al., 2017).

7.1.2 Limitations of Bathymetric Correction

The bathymetric correction performed for the data obtained from the June 2017 field survey was used as a basis to define a best practice survey method for June 2018. It was concluded from the analysis of the 2017 dataset that topographic modelling of deeper channel areas and areas in close proximity to vegetation were poorly constrained compared to other areas. This analysis therefore guided RTK-GNSS data collection for the subsequent field campaign which focused a survey effort in these areas. This generally resulted in lower error results in comparison to the 2017 survey in these type areas. Despite this, there was a pervasive issue with the 2018 as a whole due to the water's level at the time of survey. It was concluded from the results that consistently low water depth over features such as riffles resulted in erroneous data due to the fact the refraction correction was

over predicting depth. As the system is characterised by a large proportion of this type of feature, the error was pervasive across of much of the reach. This is therefore a limiting factor of SfM technologies, as this type of analysis is only applicable for mid-range depths with low turbidity (Woodget et al., 2014).

In the case of the River Feshie, the accuracy of survey results therefore depends on timing and water depth. When the river is at median flow conditions this type of survey method is thought to be applicable as deeper pools are only characteristic of a few sections of the river with limited spatial extent. Additionally, there is a limited spatial extent of particularly shallow water during these flow conditions. A reality small survey effort coupled with a bathymetric correction can therefore produce acceptable results.

The conclusions from this analysis suggest that a SfM analysis coupled with a bathymetric correction is still an applicable survey method for the site if survey timing is considered. Additionally, for the most optimum results a complimentary RTK-GNSS survey should be carried out which has a focused effort to survey deeper areas, so that when coupled with the overall SfM-MVS survey the resultant DEM can be deemed accurate, with low error values across the breadth of the study area. However, this solution is only applicable for settings such as the River Feshie, where the majority of the fluvial landscape is characterised by mid-range, low turbidity water. Additionally, as survey timing is so constrained it reduces the feasibility and ease of such survey method.

7.2 <u>Rainfall Runoff and Geomorphic Change Modelling</u>

7.2.1 Catchment Scale

7.2.1.1 Resolution and Representation of Reality

DEM

The catchment scale modelling of the river was performed at a reasonably coarse scale of 100 m^2 . The modelling process for this investigation took a 'fining down' approach, by modelling catchment scale processes at a much coarser resolution than when considering a single river reach. This method was chosen due to limitations in the scope and timescales of this year long research project. The

main aim of the catchment scale modelling was to produce a simulation of the river's overall hydrology based on inputs of rainfall data. As these files were also at a coarse resolution (1 km²) it would have been trivial to create a significantly finer resolution DEM, when rainfall is the primary input driving this model. On the corollary, it was important not to make the DEM too coarse as internal slope and hydrological processes which occur on a grid cell basis must still be allowed to drive the system in a representative manner.

Bedrock DEM

An assumption was made within the modelling process that bedrock had a consistent depth of one metre below the exposed topography. However, it is unreasonable to think that this would be the case in reality. One would assume the value to be lower on the hill crests where weathering has allowed exposure of bedrock, and hillslope instabilities have resulted in the loss of material. Whereas in the valleys the value would likely be higher, where debris has accumulated at the base of slopes and conditions are more favourable to the stabilisation of thicker soil layer less affected by the influence of physical and chemical weathering. However, this assumption was deemed appropriate as this parameter is thought not to be particularly significant when modelling the processes of sediment transport on hillslopes. This factor is only likely to be influential when considering its effect on hydrology, a parameter which was calibrated within the process anyway. Therefore, by calibrating the hydrology of the system, any potential error in the influence of bedrock depth has been compensated for through parameterisation. Obtaining quantitative measurements of bedrock depth is beyond the scope of this master thesis; geophysical techniques such as ground penetrating radar, would be need which would be both expensive and time consuming.

Rainfall Input Files

The parameter which results in the highest level of uncertainty within the catchment scale modelling is the rainfall input files. The files used as inputs into CLiDE were catchment averaged files obtained from the NFRA, based initially on rainfall data obtained from the Met Office. The Met Office rainfall data is the only long term and consistent rainfall record which is currently available for the River Feshie catchment. As mentioned in the methodology section, in more recent years

a network of rain gauges have been installed and maintained within the catchment, but at this time they do not have enough data to be useful in any medium to long term monitoring study. Subsequently, the Met Office data were the best option for this study. Unfortunately, there is not a Met Office rain gauge installed within the Glen Feshie valley meaning that the data used in this study was obtained from other rain gauges in adjacent valleys which were anywhere from 9 to 33 km distance away from the catchment. Table 15 below displays the relative percentages of the different distances to the rainfall gauges used for the precipitation data values. Where the 'Distance to Gauge' parameter is defined precisely as, 'the value of the distance to the closest rain gauge (in metres) used to interpolate rainfall at that grid point'. As one can see, the highest proportions of the pie chart are the lower distances (from 9 KM to 13 KM) with only very minor portions of the data being obtained from a distance greater than this - never more than 0.5% of the dataset.

Distance	Count	Percentage
to Gauge		
(M)		
9000	4302	21.42%
10000	11690	58.19%
11000	2416	12.03%
12000	1046	5.21%
13000	396	1.97%
14000	30	0.15%
18000	61	0.30%
19000	68	0.34%
20000	3	0.01%
21000	15	0.07%
22000	30	0.15%
33000	31	0.15%
Total	20088	100.00%

Table 15. Distance to gauge as both a count and a percentage.

This 'Distance to Gauge' value can then be used as a proxy for uncertainty of the precipitation values. Where disparities between the discharge of model outputs versus observed flow gauge data could be the result of erroneous precipitation data that are feeding the model. This parameter is even more significant in a setting such as the Cairngorm National Park where mountainous terrain and variable relief mean that individual valleys and, in turn their river catchments, can have specific micro climatic conditions. This means that the timings of storm flow events shown in the modelled river system might not correlate directly with

what the observed flow gauge is showing; especially in relation to shorter more localised rainfall events. The chance of this being the case is additionally increased the greater the distance the rainfall gauge is away from the River Feshie catchment. In addition to this, the presence of snowmelt in the catchment means that hydrograph peaks might occur without any rainfall occurring in the catchment. Additionally, if snow is present rainfall events are likely to be accentuated in the storm hydrograph as rainfall causes snow to melt at a faster rate than dry conditions.

The CLiDE hydrological model aimed to represent comparable magnitudes of storm flow events occurring in the catchment, and aims to be able to largely replicate any persistent longer term storm events. However, due to the resolution of the precipitation data used in this analysis it is unrealistic to assume that the model will be able to replicate the precise timing of shorter rainfall events. Additionally, it is possible that smaller short term events which are shown in the river's observed flow gauge record might be completely omitted from the model simulation due to being solely constrained to the Glen Feshie. Figure 48 displays some potential examples of these differences in discharge as a result of precipitation.



Figure 48. Peak flow discharges from final model output run. Where red circles denote areas where modelled and gauged storm flow peaks do not match up well with one and other.

Other additional uncertainties which may be introduced because of the rainfall input file are those related to defining the boundaries of the River Feshie catchment. These are two-fold, the first being related to the interpolation of gridded rainfall values to those averaged by catchment. Whereby data from a grid square is included in the catchment averaged value if 50% of more of its area falls within the catchment. The effect of this is significant when considering smaller catchments, and areas where there is a large amount of rainfall variability. This could potentially be significant within the Feshie catchment had the precipitation data files been at a higher resolution. However, as this was not the case the effect

of this catchment averaged interpolation is thought to be negligible; due to the factors discussed above. Additionally, the catchment is large enough that the effect of including/excluding a portion of the precipitation (on a km grid scale) should be generally inconsequential on the errors associated with the total values.

The same can be said when considering the exact boundaries of the CLiDE DEM catchment input file verses the catchment boundaries over which precipitation values have been determined. The two catchments have the same overall size and shape, however the precise boundaries of the files do not match up perfectly. This is because the catchment for this study was chosen based solely on the location of the SEPA Feshiebridge gauge, whereas the complete River Feshie catchment would end c. 1 km further downstream where the River Feshie meets the River Spey. This was done so that the discharge output of the catchment scale model could be related directly to the flow gauge data. Therefore, it is thought that the direct benefit of this comparison outweighs the error associated with the difference in extent of the catchment boundaries. Additionally, this difference results in a very small disparity in terms of spatial extent - on the order of less than 1 km². Again, this effect is even more so less significant when one considers the overall accuracy of the rainfall data for the catchment. However, had this study been performed with higher resolution precipitation data, the above parameters should be quantified to improve the robustness of the results.

7.2.1.2 Environmental and Landscape Change

The Landuse input file contains 1 km² gridded data for majority of the UK, which breaks landuse in to one of 10 broad classes. Generally, this information is reasonably coarse when considering the analysis of a single catchment. Also, in the case of this catchment, information obtained to populate the dataset is more likely to have a higher error associated with it due to the remoteness of Glen Feshie, in comparison to more densely populated and easily accessible areas. The broad classes which the landuse file gives certain areas of the catchment are largely applicable - this can be confirmed by multiple field visits to the site. However, these parameters are reasonably coarse and do not vary in either space or with time.

When considering the above, it is important to highlight the potential influence of the recovery of Caledonian Pine within the catchment over recent years, and the effect this is having on catchment's hydrology. Vegetation rewilding has occurred primarily on the floodplain and lower slopes of the catchment, and much of the upper barren or heather covered slopes have been uninfluenced by changes in the deer population. One can subsequently estimate the proportion of the catchment which has seen change due to the replantation effort from Google Earth satellite images through time. When this was carried out, approximately 30% of the catchment's spatial extent was covered by a combination of existing woodland, and new developing woodland resulting from rewilding. New replantation accounted for around 5-10% of this area. However, it must be noted that this analysis of area was derived from a purely qualitative approach for the purposes of this study, and a more detailed investigation of land use cover should be completed if any broader conclusions were to be drawn from such statistics. Considering the extent of these areas, and a 100-m resolution the impact of the growth of these plantations should have a fairly negligible influence on the overall model outputs. Additionally, it would be hard to precisely quantify the extent of the replantation's influence on model results - especially within the scope of a year long study. Therefore, as with the influence of bedrock depth, the hydrology of the catchment has been calibrated whilst including this land use parameter. The calibrated model therefore accounts for this variation in land use. Considering the above, this is deemed appropriate for this study, but in an ideal situation a more detailed look at of how land use influences surface and subsurface water flows would be ideal.

7.2.1.3 Calibration

In an ideal scenario the calibration stage of the hydrological modelling would have included more runs than were carried out. However, as there was a time limiting factor on this investigation, only seven CLiDE model runs were carried out to constrain the model's hydrological predictions. This is the same number of model runs that were completed by Barkwith (2015) on the Eden Valley catchment, indicating that this is an appropriate number when considering the scope of this study. When performing a calibration it is best practice to only alter one variable at a time to avoid autocorrelation of opposing variables. However, due to time constraints this wasn't always possible. Model runs took anywhere from 24 hours

to 2 weeks to complete, therefore in the timeframe provided it was sometimes necessary to alter multiple variables in a single model run. Table 11 shows when this was the case.

The catchment calibration process considered two components: baseflow and storm flow. Whilst considering the above information which pertains to the precipitation input files, one can note that calibration of the storm flow parameter was likely to be more challenging than that of the baseflow because if the rainfall component of the hydrological cycle was not captured, there would be no way to transport this effect into the river system where it would be represented as a storm flow peak. It was therefore determined that the calibration process should aim to replicate the overall frequency and magnitude of storm flow peaks, as opposed to trying to perfectly match up peak timings with the flow gauge data. When considering this re-evaluated aim, and observing the results of the catchment scale modelling it can be concluded that the CLiDE model adequately simulated high flow magnitudes. One must however note that this statement is someone qualitative, as although based on quantitative values no robust statistical analysis has been performed to objectively state how similar flow data from the model and the gauge are. Despite this, considering the aims and timescales of this thesis, this level of analysis has been deemed acceptable. When the broader errors associated with rainfall runoff modelling are considered, a rigorous statistical test is not deemed necessary. This is because it is unlikely both gauge and model datasets will ever perfectly mirror one and other, and it is therefore unrealistic to perform a statistical test which ranks data based on such.

Hydraulic Conductivity

In Calibrations 1 and 2 there was a fairly substantial difference between the model and gauge outputs in terms of both base flow and storm flow. This was due to an unrepresentative value for hydraulic conductivity - a parameter which determines the flow at which water moves through the subsurface. The model was initially producing outflows that were dramatically higher than those recorded at the flow gauge for both base and storm flows. This was corrected for by decreasing the value of hydraulic conductivity so the modelled system transported subsurface water slower than it had previously done. This produced model outputs which were more representative of the way the natural River Feshie system was behaving in terms of quantitative values of discharge. However, both components of the flow still needed to be constrained further to better represent reality.

Specific Yield

To do this the specific yield and rainfall interception parameters were altered, to try and represent the rate and percentage input of precipitation into the catchment's hydrological system correctly. Questions were asked as to how much rainfall was transported into the river system during a storm, and what were the timescales of this? The parameter of specific yield acts to characterise how sensitive the subsurface hydrological system, specifically groundwater, is to inflows and outflows of water. Where a high specific yield means that the groundwater head's response to a flux of water in or out of the system will be smaller and less subtle than if a system was said to have a low specific yield.

It can be noted from Figures 25 that baseflow values with time were a lot more subdued for the model outputs, in comparison to the gauged discharge values. This issue was combatted through a reduction in specific yield multiplier; initially by half and then by half again. This parameterisation aimed to increase the groundwater's sensitivity to rainfall inputs so that changes in precipitation were reflected more notably in the river's baseflow as well as its storm flow. This reduction in specific yield worked to make discharge values more comparable, up until the point where they were limited by the quality of the precipitation input files. The output of Calibrations 4 (Figure 26) shows a trend where, some peaks in gauged baseflow are not represented by the model whatsoever, whereas others match up well. The question then arises as to whether the remaining disparities between the baseflow trend lines are a result of errors in the precipitation data, or whether it is an error regarding the parameterisation of the model? Considering the known error in the rainfall data files, it is fair to assume that this is contributing somewhat to this disparity, but the extent to which they are is difficult to accurately quantify.

At this stage in the calibration process it was determined that the model was reproducing the overall trend of baseflow to an appropriate standard; mean discharge over a two-year period can be observed to be a similar value for both datasets. Additionally, despite the fact subtle fluctuations in the two trend lines

don't exactly match up, the model displays a reasonable representation of reality in terms of general magnitude and frequency of high and low baseflows.

Rainfall Interception and its Link to Flash Flooding

The rainfall interception parameter influences the storage of water within a catchment's hydrological system, and the lag time of transportation of water from when it falls as precipitation verses when it reaches the river system. Increasing the interception parameter will mean a system is less flashy, and less responsive to changes in precipitation. The River Feshie is notoriously known to be a characteristically flashy river, due to its mountainous terrain which results in periods of snowmelt, as well as contributing to more frequent and heavier rainfall events. A large influx was water into the river coupled with a well-connected hydrological network with limited storage has resulted in many flash flooding events over the River's history. Such events have seen a marked increase in water level over time periods on the order of an hour or two (Macdonald, 2016).

Considering this, it was therefore appropriate to decrease the amount of rainfall interception within the catchment with the aim to allow the modelled system to act in a flashier manner in relation to its stormflow peaks. This alteration had the desired effect, whereby the magnitude and frequency of high flow peaks for both the modelled and gauged flows matched up well post calibration. Again, one must note that not all peaks correlate directly to one and other due to omitted rainfall events. However, it is thought that the modelled system post calibration retains an acceptable level of likeness to reality in the fact that it can reproduce a representative storm flow hydrography based on the rainfall inputs it is given.

One aside which must be noted in relation to the rainfall interception parameter is that CLiDE has the ability to account for seasonality. This parameter can be varied on a monthly basis to account for the cyclical trends which happen as a result of deciduous woodland. However, due to the resolution of the rainfall input data it was determined that trying to incorporate seasonality into the modelling framework would be futile and hard to constrain especially when considering the scope of the study. Ideally this parameter would be included in a finer resolution study, where input data has a higher level of accuracy.

7.2.1.4 Conclusions on Hydrological Model Performance

As set out in the aims of this thesis, this investigation hoped to calibrate a catchment scale numerical model (CLiDE) to predict catchment hydrology and rainfall runoff using inputs from rain gauge data. This aim was set out so that the subsequent reach scale model had an input that was directly related to rainfall as opposed to supplying the model with flow data that was devoid of any catchment link. The above analysis highlights the complications involved in performing modelling through this 'top down' approach, as any subsequent models rely on the accuracy of the initial precipitation data. Therefore, the accuracy of the data obtained from any analysis will be a function of the accuracy and spatial and temporal resolution of the precipitation input files.

This analysis satisfies the above aim in that, the catchment scale modelling performed using CLiDE has provided a modelled system and subsequent outputs which represent the hydrology of the River Feshie catchment. For the specifications of this study, the resolution and accuracy of this data is deemed acceptable as the aim of this component of the modelling framework is to drive the successive reach scale model in a representative manner. Therefore, the intention of this wider catchment model is to obtain flow values which are broadly representative of the characteristics of the River Feshie's hydrological system, in terms of both a consistent baseflow and storm flow peaks. However, it is not crucial that the model represents specific storm flow peaks on specific days, as long as the magnitude and frequency of storm peaks over a given time period is generally consistent with what is happening in reality. The above model satisfies this by providing a calibrated system which has the ability to replicate storm flow peaks consistent with precipitation data provided, which are comparable to those observed. Additionally, it can replicate equivalent values of mean baseflow as well as providing a similar overall trend and magnitude to the flow gauge dataset. Caution must be noted when considering this system to be fully calibrated. Due to various time and data restrictions, this calibration represents a best effort in the timescales provide. However, were this model to be used in subsequent analysis' further work would need to be done to define the system as fully calibrated and representative of reality.

The desired accuracy and resolution of a catchment scale model will depend on an individual study's specifications. The level of certainty inherent to this dataset might not be applicable for other studies where analysis focuses exclusively at the catchment scale, but considering the scope and scale of this analysis it is deemed appropriate.

7.2.2 Reach Scale

7.2.2.1 Resolution and Representation of Reality

DEM

At the reach scale, DEM resolution should be finer than that at a catchment scale: as the spatial extent of the modelled area decreases the resolution should increase. However, very high resolution modelling uses up a vast amount of computational power, and timescales of model runs are dramatically increased when the resolution of a model is increased. When considering the size of the area, CAESAR's modelling capabilities, and the timescales available for the analysis, a grid resolution of 4 metres was used. As the River Feshie contains a reasonably high proportion of coarse sediment, and geomorphic features in the landscape span tens to hundreds of metres in length, it was decided that this resolution would still maintain an acceptable amount of detail. However, were this analysis to be replicated in a study with broader scope, a finer resolution would always be desirable. Unfortunately, for this project it would not be possible to perform a full and complete analysis using a finer resolution DEM with the given timescales.

A higher resolution DEM was obtained by the SfM analysis, had this project been given a longer timescale the modelling should be redone using the higher resolution model and the outputs compared to back up or contrast conclusions made from this analysis. The presence of this finer resolution dataset should also be considered for future analysis which considers the time series of high resolution DEMs of the River Feshie. However, due to time and computational limitations it was unrealistic to use this DEM to its full potential in this thesis.

Hydrological Processes

There are however broader implications of having a DEM at a 4-m resolution. It has been stated by Williams et al. (2013) that resolution of a fluvial model will have impacts on the accurate prediction of depth, velocity, and flow routing. However, these three parameters are influenced at different scales. Their study on the Rees modelled a river reach at a variety of scales - from one to six metres (in one metre increments). Across these scales the overall prediction of depth patterns remained consistent; where the Mean Error (ME) value for all six simulations was close to zero. From the results of this investigation one can then assume that DEM resolution does not substantially influence errors associated with reconstructing depth values. However, this is less of the case when considering predicted velocity values. Here, Williams et al. (2013) found that as grid cell size increased, so did the errors associated with the prediction of these values and subsequently the precision of the results. This was the case for anything greater than a resolution of two metres - the optimum level with a ME closest to zero. Their results state that at a grid resolution of 4 m, ME equates to -0.03 ms⁻¹, and SDE is 0.36 ms⁻¹. As no observed values of velocity were obtained for this analysis on the River Feshie one can assume ME and SDE error statistics would be on a similar scale, as both the Rees and the Feshie are dynamic braided systems with similar channel widths.

DEM resolution will also determine how many grid squares per channel width there are, and therefore how accurately models can represent the hydrodynamic flow routing processes which are occurring across a river channel. Williams et al. (2013) found that, like velocity, flow routing is influenced negatively by increasing grid size, and that this effect is shown most significant in smaller anabranching channels. The width of channels varies quite substantially in the River Feshie from those at a sub metre scale, to tens of metres. Width can vary spatially and with time as channels widen during high flow events and narrow at low flows. In times of flood, smaller channels can merge with those adjacent to them and can consequently more than double in width over relatively short timescales. This dynamic nature, typical of a braided river system, is notoriously hard to incorporate into fluvial modelling, as individual channel threads have different dimensions and characteristics whilst still being incorporated into one fluvial system. Therefore, meaning a fluvial model which works well for the main stem of the channel might not be as applicable at modelling the hydrodynamic processes in a river's smaller anabranching streams.

When a model contains a consistent grid resolution it is generally thought that the larger main stems of the channel are represented more accurately than anabranches, as the ratio of grid resolution to channel width decreases. As CAESAR does not have the ability to spatially vary the resolution of grid squares this statement is true for the geomorphological modelling of the River Feshie. This is in comparison to other 2D hydrodynamic models that can vary the resolution of their model grid or mesh such as CaFloodpro (Gibson et al., 2016) and MIKE21FM (Mackay et al., 2015).



Figure 49. Predicted depths for different grid resolutions, as indicated above each map (Williams et al., 2013).

Flow Routing Directions

One of the potential challenges of using CL to model braided river morphology is how the combined model routes flows on a grid cell basis. The model's two components route flow via differing methods which led to complications when merging to the two into the CL planform. CAESAR had the ability to route flow in eight directions, whereas LISFLOOD-FP only routes flow in four. The Moore flow routing method incorporated in the combined model allows interactions in all neighbouring directions. However, because of this, Coulthard et al. (2013) state that in CL a finer grid resolution might be necessary to represent flow routing processes accurately, especially when considering narrow diagonal channels. This is important when considering braided river morphology, as a single reach can contain many narrow anastomosing anabranches. However, due to an inability to

158

change the resolution of the DEM spatially, grid resolution remained consistent for both wide and narrow channels in the River Feshie reach scale model. This therefore might have contributed to the model not being able to fully replicate a braided channel morphology, as flow routing in narrow anabranches was not being represented accurately. For subsequent research the model should either be run at a finer grid resolution, or a different morphodynamic model should be used.

Hydrological Boundary

An additional suggestion for why the modelled system had notably and consistently low water levels is the placement of the hydrological boundary nodes. These nodes link the catchment scale hydrological model to the reach scale geomorphic change model. These nodes were placed so that they spanned the width of the valley where the upstream extent of the braided reach. However, they failed to include the addition of two notable tributaries which input water into the reach. The first (labelled Tributary 1 in Table 16) is the larger of the two and inputs water at the upstream end of the reach - c. 250 m downstream from the reach's upper bound. This tributary feeds water from a smaller valley which is situated to the West of the main Glen Feshie. During the 2018 field campaign it was noted that this tributary stream was completely dry and not contributing any water to the main river during low flow. However, it has not been constrained what conditions allow this stream to dry out and for what percentage of the year it contributes water to the main river channel. From visits to the site during the wet season it has been observed that this tributary does have the potential to transport a significant amount of water during high flows. However, Table 16 displays the area of land which Tributary 1 drains, and what percentage of the total River Feshie drainage area that accounts for (3%). It was therefore decided, that because of its minimal contributing area, and its disconnect from the main river at time of low flow it was appropriate to omit this tributary for the analysis.

The effect of this decision should have a very limited on the impact of geomorphic modelling at low flows, and is considered not to be the reason for the unrepresentatively low water levels. At high flows, the discounting of this input into the reach scale model may have at a minor impact but, as the model has been unable to be fully calibrated to a stage where one can make predictions about geomorphic change based on its outputs, the omission of this stream is said to

have a limited impact on the results of this analysis. The same can be said for the omission of Tributary 2 and, as can also be observed in Table 16; it has an even smaller contributing area than Tributary 1.

	Total Area (Km²)	Tributary Watershed as % of Total	
Entire Watershed	240	100.00%	
Tributary 1	7.2751	3.03%	
Tributary 2	4.8385	2.02%	

Table 16. Area and percentage of area for the main catchment watershed, and two smaller tributaries.

7.2.2.2 Assumption of Bedrock Depth

It is thought that the primary reason as to why the reach scale model is behaving unrepresentatively is because of the assumption of bedrock depth, and how water is being routed through the subsurface. Bedrock depth was assumed to be 10 m below the surface, so as to not restrict the river system's ability to erode the river's bed and banks. However, it became evident throughout the calibration process that a large proportion of the river's discharge was draining into the subsurface. This is thought to be because of the thick sediment layer comprising the bed of the river. The situation has likely been worsened by the fact that the sediment input into the model is reasonable coarse, in comparison to a river bed comprised of silt or sand as there is more interstitial space between the grains, into which water can drain. The composition of the bed, coupled with the thickness of the layer created a highly permeably substrate which is draining the river system at unrepresentative rates. In reality, this layer will vary with depth; becoming more consolidated with less porosity and permeability. So, despite the fact it is plausible to have a sediment layer this thick, CL is portraying an unrepresentative situation in the fact that the layer's characteristics don't change with depth. This modelling is therefore limited to either allowing the model erode to its fullest extent while over representing subsurface drainage or capping the model's ability to erode but more accurately representing subsurface drainage. As it was more important that the model was accurate during high flows the former scenario was chosen. However, had this analysis had a larger scope with a longer timescale an effort would have been made to better calibrate this parameter as

there is likely a balance between the depth of bedrock and the amount of subsurface drainage, somewhere between 10 m and the level of the surface.

7.2.2.3 Implications of Sediment Recirculation

As a geomorphic change model, CL requires a sediment input. As no observed data were obtained as part of this analysis - due to time constrains and labour necessary to do so. Additionally, the use of historic data meant observed data were limited to that obtained at the time. It was decided that the model should recirculate the sediment which was being output at its downstream node. However, one must note the limitations and uncertainties that arise as a result of using this method. An experiment by Parker and Wilcock (1993) tested two different sediment input scenarios in a flume experiment, where they ran sediment through the system until it reached an equilibrium state. They compared the differences between feeding the flume with 'new' sediment at its upstream bound, verses recirculating sediment from the end to the start of the flume. This experiment was carried out for both uniform and distributed sediment grains in a flume with set dimensions, so processes occurring in every scenario obey the same laws of flow and sediment transport. For uniform sediment, the final equilibrium scenarios reached were equivalent to one and other, and were independent of their initial conditions; despite the fact flow and sediment constraints differed. However, this was not the case when grain size was varied. In the scenario were sediment was fed into the upstream boundary, the amount of sediment and its distribution must be specified by the user and must be identical to the bed load characteristics at the model's final equilibrium stage. This suggests that the equilibrium state is independent of the initial conditions - in fact the corollary is true. This is in comparison to a recirculating scenario which has a varied grain size distribution. Where neither the bed load, nor the bed surface material can be specified in advance. Demonstrating that the final equilibrium state of the model will depend on the initial conditions - a statement which is true for this analysis.

These finding therefore means that, generally, transport relations obtained using one of the methods will not be comparable to the other. This is because of a difference in the governing constraints between the two whereby, if fractional transport is scaled by the grain size distribution, the two different methods can

produce different transport relations despite having the same sediment characteristics. To maintain consistency, transport should therefore be scaled by the size distribution of the bed surface. However, as no direct observed sediment input data is available, a quantitative comparison of the above two methods in the setting of the River Feshie is out with the reach of this analysis. Nonetheless, one should still consider the above conclusions draw by Parker and Wilcock (1993) which highlights that geomorphic change which occurs as a result of sediment transport will be sensitive to the initial conditions input into the model, i.e. factors such as the grain size distribution. However, as the reach being analysed is less than 2 km in length the differences between these two methods might not be significant in comparison to modelling flow along the entire length of the River Feshie. There is no simple way to determine how much of an influence this parameter is having on the reach without further analysis which is out with the scope of this investigation; nevertheless, the mechanism by which sediment is provided to the model is noteworthy when observing model outputs and drawing conclusions.

Additionally, it can be stated that the flume scenarios conducted by Parker and Wilcock (1993) cannot be expected to fully replicate natural sediment transport processes. Furthermore, natural channels are thought never to completely reach an equilibrium state. This statement is especially relevant to the River Feshie - a dynamic river.

7.2.2.4 Bulk Sampling

Field Method

The methodology used for bulk sampling followed a field method (Hoey, 2014) and a subsequent statistical analysis (Ferguson and Paola, 1997). Regarding the former, it is worth considering the variation between the two samples collected and if any broader trends can be extrapolated from these observations. A suggestion might be to obtain more observed data here, however each sediment sample took ³/₄ of a day to complete with 5 people excavating and weighing sediment. Therefore, when weighing up the cost-effectiveness of the amount of time spent obtaining this data versus the information obtained, it was deemed inefficient to carry out more bulk samples. Having a sample at the upstream and

downstream bounds of the reach scale model was thought to fairly represent the sediment characteristics which were included for this analysis. Tables 16 compares sediment statistics averaged for both sites. This data shows that the two sites are comparable with a D50 value within 0.001 m. However, it can be noted that there is a broader spread in the grain size distribution of the upper site, which contains a higher D84 value and a lower D10 value. However, these differences are on the order of a few cm's. From these results it can be shown that there is limited change in GSD along the braided reach. Reinforcing the fact that completing further bulk sediment samples would likely not be worthwhile for this analysis. However, one can note that this statement has been drawn on limited data and in a study with a longer timeframe or wider scope it might be worthwhile completing more bulk sediment samples.

(Metres)	Upstream	Downstream
D10	0.0096	0.0173
D50	0.0686	0.0692
D84	0.1475	0.1255
Active Layer Thickness	0.2949	0.2511

Table 17. A comparison of sediment statistics for the two bulk sediment sites sampled.

Statistical Analysis

The statistical analysis of the bulk sediment data used the method detailed in Ferguson and Paola (1997). This analysis focuses on the statistics of precision and bias in a dataset based on the amount of sediment information obtained, and its characteristics. A slight caution of using this approach is that the equations used to produced statistical parameters are based on an assumed normal distribution. These parameters will only be valid if an underlying normal distribution in φ -units can be assumed for the surveyed area - i.e. there is no skew in the dataset. This is a broad assumption and is not generally true for most GSDs. However, assuming there is no major skew in the dataset (as is the case for this investigation) the conclusions drawn from these statistics should be reasonably valid and give a good overview of bias and precision.

From their analysis, Ferguson and Paola (1997) concluded that smaller sample sizes are biased in that they over represent the fine fraction of sediment in a GSD. If larger grain sizes are marginally less frequent in a sediment unit, it is likely that

a small sample size will amplify this phenomenon and display it unrepresentatively in the GSD results. These conclusions highlight the importance of choosing a representative survey site, and having a large enough sample size - as was the case for the River Feshie analysis.

In addition to this, the paper also noted that bias is more prominent for poorly sorted grains, in comparison to a GSD with a low standard deviation (i.e. well sorted). This point was deemed significant when considering the bedload sediment of the River Feshie - a poorly sorted gravel bed river. As per the results of their analysis, Ferguson and Paola (1997) stated that the mass of the sample required to avoid bias for the D95 particle size is around two orders of magnitude larger than that required for the D50. The same is also true when considering sample size and precision. One can therefore note that in Figures 30 & 31 the precisional error is notably large when considering D84-D95 parameters, as errors increase non-linearly when bed material is poorly sorted. As per the equations in Ferguson and Paola (1997), sample size would need to be increased to 10's of thousands of Kgs in some cases to result in zero bias and precision. This can be displayed clearly by Figure 31 - the precisional error on the River Feshie upstream surface site. Here, for 0 precisional error on the D95 parameter, a sample size of c. 60,000 Kgs is needed - see Appendix 1. This is because of a poorly sorted surface sediment layer - i.e. there is a wide range in grain sizes. In cases such as this, obtaining 60 tones of sediment is unrealistic when considering the scope of this investigation. It is therefore deemed more appropriate to consider the precisional error on the D50 or median grainsize, which is within acceptable bounds for a gravel bed braided river environment.

These statistical parameters are useful for highlighting the sediment characteristics of the River Feshie - by displaying quantitatively how poorly sorted both the surface and subsurface sediment layers are, and how variable this sorting can be. Despite the fact mean values of sediment between both the upstream and downstream sites are broadly comparable, statistics regarding sorting do not so the same affinity to one and other. Additionally, precision statistics highlight how sure one can be about the GSD input into the reach scale model. The accuracy of this data is within keeping within the scope of this analysis, especially when considering CLs restricted ability to deal with distributed GSDs. For reach scale

modelling at the highest resolution GSD should be distributed both by size, and spatially across the catchment. However, it is only currently possible to distribute GSD by size and proportion - i.e. we can observe broad trends in sorting but these cannot be distributed spatially. If GSD was to be distributed spatially a large effort would have to be made regarding the bulk sediment sampling - on the scale of multiple weeks or months of data collection depending on available labour.

7.2.2.5 Calibration

Spin Up

The initial spin up run for CL worked well, the model behaved as expected. Fines were preferentially transported for the initial few years of the model run while the modelled system began to spin up and equilibrate to a more representative sediment transport model. This happened as it is likely that the sediment which the channel bed is comprised of is unrepresentatively fine. Fine sediment is therefore initially washed down the system as low to medium flows pass through the reach. Additionally, this preferential removal of fines will have been influenced by the choice to recirculate sediment. The system would have therefore needed a longer time to spin up in comparison to model where observed sediment data were being input at the upstream bound. Figures 32 & 33 display how the modelled system evolved through time, where one can observe sediment peaks correlating more directly with discharge peaks with time. However, it does appear this correlation is reasonably subtle, and a further statistical analysis of the numerical outputs of this data should be done to prove this correlation were the data to be used in further work.

Additionally, the sediment vs discharge plot appears to display a representative distribution. One question can be asked as to why this plot appears to have two distributions, where sediment output is higher (1) and lower (2) for the same discharge. An answer to this might perhaps be to do with the configuration of channels in the braid plain, certain anabranches may contain more mobile sediment than others. Therefore, as the system migrates through time sediment may become more readily available depending on its course. However, the quantification of this is beyond the reach of this analysis and has therefore not been performed.

166 Representation of Reality - Calibration 1

For the initial calibration run the model had too high lateral erosion which resulted in the landscape becoming flattened as banked eroded and channels filled. This process was likely worsened by the simplification of equations relating to fluid dynamics included in CL. The model is current unable to represent momentum as per real rivers, and therefore is unable to produce natural geomorphic units such as pools and riffles. Because of this, the model has been found by others to have a tendency to create flatter channels in comparison to natural systems (Coulthard et al., 2013). This should be noted as a limitation of the model, and be taken into account when considering its applications. For very high resolution analysis which focuses on specific geomorphic units this model might not be applicable. However, for a study such as this which aims to quantify the overall geomorphic change in terms of broad areas of deposition and erosion, CL is an appropriate model to help answer such aims.

A Comparison to Ziliani et al. (2013) - Calibration 3

Perhaps the most significant analysis which uses CL in a gravel bed braided river setting is that by Ziliani et al., (2013). In their study, the team used CAESAR to model the fluvial morphodynamics of a 33 km long braided reach of the River Tagliamento (Italy). The overall aim of the study was to evaluate the performance of the model against observed survey data, as well as providing a general overview of the applicability of this kind of reduced complexity CA model in this setting. The scope of this paper is much broader than this thesis - with a longer distance being surveyed, and a much more in depth calibration process. To decrease model run time a threshold discharge was chosen over which geomorphic change would be modelled. There were 226 days above this threshold discharge value meaning the model ran for this number as opposed to 2957 days had there been no threshold. This highlights CLs influence on channel morphodynamics during low flow conditions, and therefore reinforces that the calibration process should primarily focus on how the model is performing during high flow. However, to have confidence in this perceived lack of geomorphic change at low flows, a model simulation should be run for both high flows and for all flows, and the outputs compared via a DoD

7.2.2.6 Timescales of Reach Analysis

On the outset of this research project the initial aim was to calibrate a reach scale model from 2000-2011 which would then be used to make predictions about geomorphic change for the subsequent 7 years, i.e. up until the present day. However, it was decided based on the outputs from model calibrations that this was beyond the limits of the current CL modelling. DEM datasets were present from 2013, 2017 and 2018 meaning that the CL model would have to be able to run over a four-year time period to be able to compare modelled to observed DoDs. The current calibrated model after two or three years was beginning to degrade guite substantially from the situation in reality, and the output DEMs were largely unrepresentative - displayed a more flattened landscape than the case in reality. It was therefore decided that this analysis should not aim to model geomorphic change over this time period, as results would likely have limited gravitas. This thesis therefore focuses more on the limitations of braided river reach scale modelling at the multi-event scale, and provides advice for future studies when carrying out similar analysis'. However, still aims to contain a thorough parametrisation of a reach scale model despite the model outputs not directly matching up with reality.

7.2.3 Geomorphic Change Detection

7.2.3.1 Error Surfaces

The assumption of having a consistent error surface of 10 cm on all RTK-GNSS derived DEMs has the potential to impact change analysis. All observed DEMs were given this error surface, as there is generally an inherent error associated with RTK-GNSS survey. For the case of the River Feshie this was generally less than 10 cm but having this limit encompasses all data which is likely to be erroneous. These error surfaces were purely used when creating DoDs via the GCD toolkit developed by Wheaton et al. (2010). For this analysis, the error surfaces were created to be applied to a DoD when a propagated error threshold of 80% was used. This couples any potential error associated with the data with a percentage of uncertainty to obtain a DoD that aims to exclude any noise or erroneous data.

However, the drawback of this approach is that it can remove any subtle changes in topography. This is significant when considering braided rivers due to their small vertical relief in comparisons to their horizontal extent. A small change in

elevation can have significant impacts on channel morphology, i.e. a substantial anabranch can develop as a result of a few cm's of erosion during a high flow event. The outputs from the GCD used for the analysis of the River Feshie are coarse and patchy. This is likely a result of several factors including the initial DEM resolution, poorly constrained CL outputs, and broad encompassing error values obtained from using the GCD toolbar. However, despite the lack of detail, these outputs appropriately display the broad overall trends of surface lowering and increasing as is displayed in Figure 41-46.

It must be noted that this type of analysis was only performed for the observed datasets, as modelled dataset were assumed to be true. This is purely because we cannot relate a quantifiable error to the model outputs, especially one that is spatially distributed. However, such errors must be qualitatively reviewed and considered especially when model outputs represent a situation distant from reality.

8 Conclusion

The aim of this thesis was to develop and assess an end-to-end modelling framework for the braided River Feshie. This framework consisted of two main components; a rainfall runoff model (CLiDE) and a morphological change model (CAESAR LISFLOOD). In summary, the model used catchment rainfall-runoff predictions to simulate geomorphic change of a c. 2 km long braided reach. Predictions from the geomorphic change model were compared to observed patterns of geomorphic change.

To create the end-to-end framework, the first objective of this thesis was to generate high quality (cm precision) DEMs of the braided reach from 2017 and 2018 field survey data. The first step of this analysis was to develop an SfM survey protocol based on 2017 data. Results showed that ground control targets should be positioned so that the edges of the survey area are preferentially represented, while still maintaining a good overall coverage of the study area. For the spatial extent of this study area it was found that c. 15 targets should be used as Ground Control Points (GCPs). This survey protocol was used to generate the 2018 DEM, which was characterised by RMSEs on check points of 0.035, 0.042, and 0.048 for the X, Y and Z axis' respectively. The constraint of error in the vertical axis is notable and indicates the success of the survey protocol. The generation of DEMs using SfM links into the broader framework of this thesis as high quality DEMs are required for the CL reach scale modelling component. However, 2017 and 2018 DEMs were not used for reach scale modelling, because the numerical modelling effort focused upon parameterising the model for the period 2000 to 2007 rather than verifying the model using data from 2017 and 2018. Despite this, this work shows how high-resolution DEMs can be generated that could be used to verify numerical model predictions of geomorphic change.

To obtain a discharge input for the CL reach scale model, a rainfall-runoff model of the River Feshie catchment was first calibrated over a 10-year period from 2001 to 2011. Many previous studies that have modelled braided river geomorphic change have focused on measured discharge as a representation of catchment hydrology. However, this work aims to predict river discharge from rainfall inputs that are input into a parameterised CLiDE hydrological model. The model was calibrated using seven simulations that involved two initial spin up stages, four calibration stages within which various parameters were adjusted, followed by a

final calibrated output model run. The discharge prediction from the calibrated model was assessed by comparing to observed flow gauge data for both base and storm flow hydrographs. For the calibrated simulation, both the model and the gauge were displaying a mean baseflow discharge value of c. 325,000 m³d⁻¹. Despite the fact that subtle variations in the predicted and observed baseflow hydrographs didn't match up perfectly, the model was deemed to be representing the overall magnitude and dynamics of the baseflow relatively well. The same was true for the calibrated stormflow. Although the model was unable to perfectly replicate stormflow peaks, perhaps due to the lack of spatially variable rainfall data within the catchment. Despite this, the final model output displayed peaks of similar magnitudes and frequencies.

The discharge outputs from the final CLiDE run were then used, alongside bulk sediment data, as inputs to stimulate geomorphic change in CL. The model was parameterised over one spin up, and six calibration runs. Parameters such as lateral erosion, and maximum erosion limit, were varied to try and simulate channel divergence and typical braiding behaviour in the reach scale model. After the final calibration stage the model could simulate the inherent behaviour of a braided system well at high flows, for the initial 1-2 years of model run time. However, once multiple floods were simulated, CL outputs degraded for both high and low flows. The parametrisation of CL included in this thesis contained two main limitations. The first, a disproportionate amount of water drained into the subsurface, meaning that in low flow conditions surface flows were not adequately predicted. Second, models tended to erode and smooth topography with time, meaning that after three years of run time model outputs lacked robustness. The CL parameterisation and modelling thus highlight potential difficulties concerning modelling braided river morphodyanmics using CL. The model can represent braided channel behaviour appropriately over shorter timescale of 1-2 years or for individual high flow events, but not for multiple consecutive high flow events.

The morphological change which occurred as a result of model simulations was quantified for a consistent section of the braided reach which had been surveyed from 2000 to 2018. This was done using DoDs for both observed (surveyed) and modelled DEMs, as well as a direction comparison between the modelled and surveyed DEMs for a set time increment (i.e. the time survey data were collected). This quantification of geomorphic change was largely influenced by how well the

model was behaving; as the model's affinity to reality degraded with time, so did the representation of appropriate geomorphic change. Initially, observed and modelled geomorphic change took place in similar areas, and was represented by similar magnitudes. However, after multiple years of model run time the observed outputs did not correlate well with the geomorphic change the model was exhibiting.

Initially, this thesis had aimed to calibrate the CL geomorphic change model over a period of 10 years from 2000 to 2010. The parameter set from this would then be used to make predictions about geomorphic change between 2017 and 2018, using data from the high-resolution datasets from field surveys. However, as further efforts are needed to improve the parameterisation of the current CL model over extended time periods, this aim was beyond the scope of the current model's capacities. Despite this, as these high quality topographic datasets exist any future effort to compare geomorphic change along this reach will be aided substantially.

Overall, this thesis shows how repeat topographic surveys can be used as input and verification data for flood risk modelling that incorporates geomorphic change, as well as displaying how much of an important component these inputs are. Additionally, this work emphasises the challenges and uncertainties associated with the accurate representation of geomorphic change within fluvial modelling. The conclusions drawn from this analysis therefore highlight the pressing need to better constrain sediment transport and geomorphic change in flood risk models so that predictions made from such models are robust.

The results of this thesis highlight the challenges associated with implementing an end-to-end modelling framework, especially within a restricted timeframe. Uncertainties associated with catchment scale modelling can feed into results produced at a reach scale. These uncertainties have the potential to be amplified in the subsequent stage of modelling, decreasing the robustness of final collective outputs. A question can therefore be asked, is the benefit of an end-to-end framework outweighed by the increase in cumulative uncertainty throughout the process? The answer is complex and situation dependent, but the results of this work suggest that depending on a study's aims and timescales, discrete spatial modelling may be the preferable option.

- Allen, R.G., Pereira, L.S., Raes, D. and Smith, M., 1998. Crop evapotranspiration-Guidelines for computing crop water requirements-FAO Irrigation and drainage paper 56. Fao, Rome, 300(9), p. D05109.
- Anderson, J.D., 1992. Governing equations of fluid dynamics. In Computational fluid dynamics (pp. 15-51). Springer, Berlin, Heidelberg.
- Arnold, J.G., Allen, P.M., Muttiah, R. and Bernhardt, G., 1995. Automated base flow separation and recession analysis techniques. Groundwater, 33(6), pp.1010-1018.
- Ashmore, P.E. and Church, M., 1998. Sediment transport and river morphology: a paradigm for study. Gravel-bed Rivers in the Environment, pp.115-148.
- Attal, M. 2017. Structural Analysis of Rocks and Regions 2017 Topographic analysis.
- Bangen, S.G., Wheaton, J.M., Bouwes, N., Bouwes, B. and Jordan, C., 2014. A methodological intercomparison of topographic survey techniques for characterizing wadeable streams and rivers. Geomorphology, 206, pp.343-361.
 Barkwith, A. and Coulthard, T.J., 2013. CLiDE version 1.0 user guide.
- Barkwith, A., Hurst, M.D., Jackson, C.R., Wang, L., Ellis, M.A. and Coulthard, T.J., 2015. Simulating the influences of groundwater on regional geomorphology using a distributed, dynamic, landscape evolution modelling platform. Environmental Modelling & Software, 74, pp.1-20.
- Bates, P., Trigg, M., Neal, J. and Dabrowa, A., 2013. LISFLOOD-FP user manual. Code release, 5(6).
- Bates, P.D., Horritt, M.S. and Fewtrell, T.J., 2010. A simple inertial formulation of the shallow water equations for efficient two-dimensional flood inundation modelling. Journal of Hydrology, 387(1-2), pp.33-45.
- Bergeron, N. and Carbonneau, P.E., 2012. Geosalar: Innovative remote sensing methods for spatially continuous mapping of fluvial habitat at riverscape scale. Fluvial remote sensing for science and management, pp.193-213.
- Beven, K.J., Kirkby, M.J., Schofield, N. and Tagg, A.F., 1984. Testing a physicallybased flood forecasting model (TOPMODEL) for three UK catchments. Journal of Hydrology, 69(1-4), pp.119-143.
- Bird, S., Hogan, D. and Schwab, J., 2010. Photogrammetric monitoring of small streams under a riparian forest canopy. Earth Surface Processes and Landforms, 35(8), pp.952-970.
- Boorman, D.B., Hollis, J.M. and Lilly, A., 1995. Hydrology of soil types: a hydrologically-based classification of the soils of United Kingdom. Institute of Hydrology.
- Blanckaert, K., Wei, M., Heyman, J., Li, D. and Schleiss, A.J., 2015. Secondary Flow in Sharp Open-Channel Bends: Experiments And Modeling. In Proc. of Int. Congress IAHR 2015
- Brasington, J., Langham, J. and Rumsby, B., 2003. Methodological sensitivity of morphometric estimates of coarse fluvial sediment transport. Geomorphology, 53(3-4), pp.299-316.
- Brasington, J., Rumsby, B.T. and McVey, R.A., 2000. Monitoring and modelling morphological change in a braided gravel-bed river using high resolution GPSbased survey. Earth Surface Processes and Landforms: The Journal of the British Geomorphological Research Group, 25(9), pp.973-990.
- Briggs, L.J. and Shantz, H.L., 1912. The wilting coefficient and its indirect determination. Botanical Gazette, 53(1), pp.20-37.
- British Standard Institution. 2012 Acquisition and management of meteorological precipitation data from a raingauge network Part 4: Guide for the estimation of areal rainfall. London, BSI Standards Limited, 14pp. (BS 7843-4:2012, CEH Project no: C04215)

- Bunte, K. and Abt, S.R., 2001. Sampling surface and subsurface particle-size distributions in wadable gravel-and cobble-bed streams for analyses in sediment transport, hydraulics, and streambed monitoring. Gen. Tech. Rep. RMRS-GTR-74. Fort Collins, CO: US Department of Agriculture, Forest Service, Rocky Mountain Research Station. 428 p., 74.
- Carbonneau, P., Fonstad, M.A., Marcus, W.A. and Dugdale, S.J., 2012. Making riverscapes real. Geomorphology, 137(1), pp.74-86.
- Carbonneau, P.E. and Dietrich, J.T., 2017. Cost-effective non-metric photogrammetry from consumer-grade sUAS: implications for direct georeferencing of structure from motion photogrammetry. Earth Surface Processes and Landforms, 42(3), pp.473-486.
- Carrivick, J.L., Smith, M.W. and Quincey, D.J., 2016. Structure from Motion in the Geosciences. John Wiley & Sons.
- Carter, W.E., Shrestha, R.L. and Slatton, K.C., 2007. Geodetic laser scanning. Physics Today, 60(12), p.41.
- Church, M., Hassan, M.A. and Wolcott, J.F., 1998. Stabilizing self-organized structures in gravel-bed stream channels: Field and experimental observations. Water Resources Research, 34(11), pp.3169-3179.
- Clapuyt, F., Vanacker, V. and Van Oost, K., 2016. Reproducibility of UAV-based earth topography reconstructions based on Structure-from-Motion algorithms. Geomorphology, 260, pp.4-15.
- Coulthard, T.J. and Macklin, M.G., 2003. Modeling long-term contamination in river systems from historical metal mining. Geology, 31(5), pp.451-454.
- Coulthard, T.J. and Van De Wiel, M.J., 2017. Modelling long term basin scale sediment connectivity, driven by spatial land use changes. Geomorphology, 277, pp.265-281.
- Coulthard, T.J. and Wiel, M.J.V.D., 2006. A cellular model of river meandering. Earth Surface Processes and Landforms: The Journal of the British Geomorphological Research Group, 31(1), pp.123-132.
- Coulthard, T.J., Lewin, J. and Macklin, M.G., 2005. Modelling differential catchment response to environmental change. Geomorphology, 69(1-4), pp.222-241.
- Coulthard, T.J., Macklin, M.G. and Kirkby, M.J., 2002. A cellular model of Holocene upland river basin and alluvial fan evolution. Earth Surface Processes and Landforms, 27(3), pp.269-288.
- Coulthard, T.J., Neal, J.C., Bates, P.D., Ramirez, J., de Almeida, G.A. and Hancock, G.R., 2013. Integrating the LISFLOOD-FP 2D hydrodynamic model with the CAESAR model: implications for modelling landscape evolution. Earth Surface Processes and Landforms, 38(15), pp.1897-1906.
- D'ambrosio, D., Di Gregorio, S., Gabriele, S. and Gaudio, R., 2001. A cellular automata model for soil erosion by water. Physics and Chemistry of the Earth, Part B: Hydrology, Oceans and Atmosphere, 26(1), pp.33-39.
- Dietrich, J.T., 2016. Riverscape mapping with helicopter-based Structure-from-Motion photogrammetry. Geomorphology, 252, pp.144-157.
- Dietrich, J.T., 2017. Bathymetric structure-from-motion: extracting shallow stream bathymetry from multi-view stereo photogrammetry. *Earth Surface Processes and Landforms*, 42(2), pp.355-364.
- Einstein, H.A., 1950. The bed-load function for sediment transportation in open channel flows (Vol. 1026). Washington, DC: US Department of Agriculture.
- Erickson, T.O. and Stefan, H.G., 2008. Baseflow Analysis of the Upper Vermillion River, Dakota County, Minnesota.

- 174
- Estrela, T. and Quintas, L., 1994, March. Use of a GIS in the modelling of flows on floodplains. In 2nd International Conference on River Flood Hydraulics (Vol. 177, p. 190). Wiley, Chichester, UK.
- Ferguson, R.I. and Paola, C., 1997. Bias and precision of percentiles of bulk grain size distributions. Earth Surface Processes and Landforms: The Journal of the British Geomorphological Group, 22(11), pp.1061-1077.
- Feurer, D., Bailly, J.S., Puech, C., Le Coarer, Y. and Viau, A.A., 2008. Very-highresolution mapping of river-immersed topography by remote sensing. Progress in Physical Geography, 32(4), pp.403-419.
- Focazio, M.J. and Cooper, R.E., 1995. Selected characteristics of stormflow and base flow affected by land use and cover in the Chickahominy River basin, Virginia, 1989-91. US Department of the Interior, US Geological Survey.
- Förstner W, Wrobel B, Paderes F, Fraser CS, Dolloff J, Mikhail EM, Rajikietgumjorn W. (2013). Analytical photogrammetric operations. In Manual of Photogrammetry, McGlone JC (ed). American Society for Photogrammetry and Remote Sensing: Bethesda. In Manual of Photogrammetry, McGlone JC (ed). American Society
- Gibson, M.J., Savić, D.A., DJordjević, S., Chen, A.S., Fraser, S. and Watson, T., 2016. Accuracy and computational efficiency of 2D urban surface flood modelling based on cellular automata.
- Hall, J. W., Tran M., Hickford, A. J. and Nicholls R. J., 2016. The Future of National Infrastructure: A System-of-Systems Approach. Cambridge: Cambridge University Press.
- Hancock, G.R., Willgoose, G.R. and Evans, K.G., 2002. Testing of the SIBERIA landscape evolution model using the Tin Camp Creek, Northern Territory, Australia, field catchment. Earth Surface Processes and Landforms: The Journal of the British Geomorphological Research Group, 27(2), pp.125-143.
- Hicks, D.M., 2012. Remotely sensed topographic change in gravel riverbeds with flowing channels. Gravel-Bed Rivers: Processes, Tools, Environments, pp.303-314.
- Hirabayashi, Y., Mahendran, R., Koirala, S., Konoshima, L., Yamazaki, D., Watanabe, S., Kim, H. and Kanae, S., 2013. Global flood risk under climate change. *Nature Climate Change*, *3*(9), p.816.
- Hoey, T.B. and Ferguson, R., 1994. Numerical simulation of downstream fining by selective transport in gravel bed rivers: Model development and illustration. Water resources research, 30(7), pp.2251-2260.
- Hoey, T.B., 2014. The size of sedimentary particles. In A practical guide to the study of glacial sediments (pp. 66-91). Routledge.
- Horritt, M.S. and Bates, P.D., 2002. Evaluation of 1D and 2D numerical models for predicting river flood inundation. Journal of hydrology, 268(1-4), pp.87-99.
- Hough, M.N. and Jones, R.J.A., 1997. The United Kingdom Meteorological Office rainfall and evaporation calculation system: MORECS version 2.0-an overview. Hydrology and Earth System Sciences Discussions, 1(2), pp.227-239.
- Howard, A.D., 1992. Modelling channel migration and floodplain sedimentation in meandering streams. Lowland floodplain rivers, pp.1-41.
- Howard, A.D., 1996. Modeling channel evolution and floodplain morphology. Floodplain processes, pp.15-62.
- Hunter, N.M., Bates, P.D., Neelz, S., Pender, G., Villanueva, I., Wright, N.G., Liang, D., Falconer, R.A., Lin, B., Waller, S. and Crossley, A.J., 2008.
 Benchmarking 2D hydraulic models for urban flood simulations. In Proceedings of the institution of civil engineers: water management (Vol. 161, No. 1, pp. 13-30). Thomas Telford (ICE publishing).

- Institute of Hydrology, 1980. Low Flow Studies. Research Report, Wallingford, Oxon.
- Israelson, O.W.; West, F.L. (1922). "Water holding capacity of irrigated soils". Utah State Agricultural Experiment Station Bull. 183: 1-24.
- Jackson, D.L., 2000. JNCC Report No. 307 Guidance on the interpretation of the Biodiversity Broad Habitat Classification (terrestrial and freshwater types): Definitions and the relationship with other Habitat classifications. REPORT-JOINT NATURE CONSERVATION COMMITTEE, (307).
- James, M. and Robson, S., 2012. Straightforward reconstruction of 3D surfaces and topography with a camera: Accuracy and geoscience application. Journal of Geophysical Research: Earth Surface, 117(F3).
- James, M. and Robson, S., 2014. Mitigating systematic error in topographic models derived from UAV and ground-based image networks. Earth Surface Processes and Landforms, 39(10), pp.1413-1420.
- James, M., Robson, S. and Smith, M., 2017. 3-D uncertainty-based topographic change detection with structure-from-motion photogrammetry: precision maps for ground control and directly georeferenced surveys. Earth Surface Processes and Landforms, 42(12), pp.1769-1788.
- James, M., Robson, S., d'Oleire-Oltmanns, S. and Niethammer, U. (2017). Optimising UAV topographic surveys processed with structure-from-motion: Ground control quality, quantity and bundle adjustment. Geomorphology, 280, pp.51-66.
- Javernick, L., Brasington, J. and Caruso, B., 2014. Modelling the topography of shallow braided rivers using Structure-from-Motion photogrammetry. Geomorphology, 213, pp.166-182.
- Javernick, L., Hicks, D., Measures, R., Caruso, B. and Brasington, J., 2015. Numerical Modelling of Braided Rivers with Structure-from-Motion-Derived Terrain Models. River Research and Applications, 32(5), pp.1071-1081.
- Jenson, S. K., and J. O. Domingue. 1988. "Extracting Topographic Structure from Digital Elevation Data for Geographic Information System Analysis." Photogrammetric Engineering and Remote Sensing 54 (11): 1593-1600.
- Kasprak, A., Brasington, J., Hafen, K., Williams, R.D. and Wheaton, J.M., 2018. Modelling braided river morphodynamics using a particle travel length framework.
- Keller, V. D. J., Tanguy, M., Prosdocimi, I., Terry, J. A., Hitt, O., Cole, S. J., Fry, M., Morris, D. G., and Dixon, H. (2015). CEH-GEAR: 1 km resolution daily and monthly areal rainfall estimates for the UK for hydrological use, Earth Syst. Sci. Data Discuss., 8, 83-112.
- Keller, V.D.J., Tanguy, M., Prosdocimi, I., Terry, J.A., Hitt, O., Cole, S.J., Fry, M., Morris, D.G. and Dixon, H., 2015. CEH-GEAR: 1 km resolution daily and monthly areal rainfall estimates for the UK for hydrological and other applications. Earth System Science Data, 7(1), pp.143-155.
- Kinzel, P.J., Legleiter, C.J. and Nelson, J.M., 2013. Mapping river bathymetry with a small footprint green LiDAR: applications and challenges 1. JAWRA Journal of the American Water Resources Association, 49(1), pp.183-204.
- Kirkham, M.B., 2005. 8-Field Capacity, Wilting Point, Available Water, and the Nonlimiting Water Range. Principles of Soil and Plant Water Relations, pp.101-115.
- Lague, D., Launeau, P., Michon, C., Gouraud, E., Juge, C., Gentile, W., Hubert-Moy, L. and Crave, A., 2016, April. A geomorphologist's dream come true: synoptic high resolution river bathymetry with the latest generation of airborne dual wavelength lidar. In EGU General Assembly Conference Abstracts (Vol. 18, p. 12238).

- Lancaster, S.T. and Bras, R.L., 2002. A simple model of river meandering and its comparison to natural channels. Hydrological Processes, 16(1), pp.1-26.
- Lane, S., Richards, K. and Chandler, J., 1994. Developments in monitoring and modelling small-scale river bed topography. Earth Surface Processes and Landforms, 19(4), pp.349-368.
- Lane, S.N. and Carbonneau, P.E., 2008. High resolution remote sensing for understanding instream habitat. Hydroecology and Ecohydrology: Past, Present and Future. Chichester: John Wiley & Sons, Ltd, pp.185-204.
- Lane, S.N., 1998. The use of digital terrain modelling in the understanding of dynamic river channel systems. Landform monitoring, modelling and analysis, pp.311-342.
- Lane, S.N., Tayefi, V., Reid, S.C., Yu, D. and Hardy, R.J., 2007. Interactions between sediment delivery, channel change, climate change and flood risk in a temperate upland environment. Earth Surface Processes and Landforms: The Journal of the British Geomorphological Research Group, 32(3), pp.429-446.
- Lane, S.N., Westaway, R.M. and Murray Hicks, D., 2003. Estimation of erosion and deposition volumes in a large, gravel-bed, braided river using synoptic remote sensing. Earth Surface Processes and Landforms: The Journal of the British Geomorphological Research Group, 28(3), pp.249-271.
- Lejot, J., Delacourt, C., Piégay, H., Fournier, T., Trémélo, M.L. and Allemand, P., 2007. Very high spatial resolution imagery for channel bathymetry and topography from an unmanned mapping controlled platform. Earth Surface Processes and Landforms: The Journal of the British Geomorphological Research Group, 32(11), pp.1705-1725.
- Lewin, J., 2013. Enlightenment and the GM floodplain. Earth Surface Processes and Landforms, 38(1), pp.17-29.
- Luhmann T., Robson S., Kyle S., Harley I., 2006. Close Range Photogrammetry: Principles, Methods and Applications. Whittles, Caitness. Rajikietgumjorn W. 2013. Analytical photogrammetric operations.
- Lyra, G.B., de Souza, J.L., da Silva, E.C., Lyra, G.B., Teodoro, I., Ferreira-Júnior, R.A. and de Souza, R.C., 2016. Soil water stress co-efficient for estimating actual evapotranspiration of maize in northeastern Brazil. Meteorological Applications, 23(1), pp.26-34.
- Macdonald, P., 2016. Change in Glen Feshie. [online] Scottish National Heritage. Available at: https://www.eca.ed.ac.uk/sites/default/files/PMacdonald-ChangeInGlenFeshie-SNH-Nos-AW2016-PDF.pdf [Accessed 1 Dec. 2017].
- Mackay, C., Suter, S., Albert, N., Morton, S. and Yamagata, K., 2015. Large scale flexible mesh 2D modelling of the Lower Namoi Valley.
- Maddock, I., 1999. The importance of physical habitat assessment for evaluating river health. Freshwater biology, 41(2), pp.373-391.
- Manning, R., Griffith, J.P., Pigot, T.F. and Vernon-Harcourt, L.F., 1890. On the flow of water in open channels and pipes.
- Marcus, W.A., Fonstad, M.A. and Legleiter, C.J., 2012. Management applications of optical remote sensing in the active river channel. Fluvial Remote Sensing for Science and Management, pp.19-41.
- Milan, D.J., Heritage, G.L. and Hetherington, D., 2007. Application of a 3D laser scanner in the assessment of erosion and deposition volumes and channel change in a proglacial river. Earth Surface Processes and Landforms: The Journal of the British Geomorphological Research Group, 32(11), pp.1657-1674.
- Milan, D.J., Heritage, G.L., Large, A.R. and Fuller, I.C., 2011. Filtering spatial error from DEMs: Implications for morphological change estimation. Geomorphology, 125(1), pp.160-171.

- Moore, E.F., 1962, December. Machine models of self-reproduction. In Proceedings of symposia in applied mathematics (Vol. 14, No. 1962, pp. 17-33). American Mathematical Society New York.
- Morton, D., Rowland, C., Wood, C., Meek, L., Marston, C., Smith, G., Wadsworth, R. and Simpson, I., 2011. Final Report for LCM2007-the new UK land cover map. Countryside Survey Technical Report No 11/07.
- Motta, D., Abad, J.D., Langendoen, E.J. and Garcia, M.H., 2012. A simplified 2D model for meander migration with physically-based bank evolution. Geomorphology, 163, pp.10-25.
- Ocampo-Duque, W., Ferre-Huguet, N., Domingo, J.L. and Schuhmacher, M., 2006. Assessing water quality in rivers with fuzzy inference systems: A case study. Environment International, 32(6), pp.733-742.
- O'Connor, J., Smith, M.J. and James, M.R., 2017. Cameras and settings for aerial surveys in the geosciences: optimising image data. Progress in Physical Geography, 41(3), pp.325-344.
- Ordnance Survey. 2018. A Guide to Coordinate Systems in Great Britain. [online] Available at: https://www.ordnancesurvey.co.uk/docs/support/guidecoordinate-systems-great-britain.pdf [Accessed 1 Dec. 2017].
- OS Terrain[™] 5 DTM. 2018. https://digimap.edina.ac.uk/webhelp/os/data_information/os_products/os_ terrain5_dtm.htm: Ordnance Survey.
- Parker, G. and Wilcock, P.R., 1993. Sediment feed and recirculating flumes: Fundamental difference. Journal of Hydraulic Engineering, 119(11), pp.1192-1204.
- Pasculli, A., Audisio, C. and Sciarra, N., 2015. Application of CAESAR for catchment and river evolution. RENDICONTI ONLINE SOCIETA GEOLOGICA ITALIANA, 35, pp.224-227.
- Peakall, J., Ashworth, P. and Best, J., 1996. Physical modelling in fluvial geomorphology: principles, applications and unresolved issues. The scientific nature of geomorphology, pp.221-253.
- Piggott, A.R., Moin, S. and Southam, C., 2005. A revised approach to the UKIH method for the calculation of baseflow/Une approche améliorée de la méthode de l'UKIH pour le calcul de l'écoulement de base. Hydrological Sciences Journal, 50(5).
- Proffitt, G.T. and Sutherland, A.J., 1983. Transport of non-uniform sediments. Journal of Hydraulic Research, 21(1), pp.33-43.
- Lawley R. And Garcia-Bajo M., 2010. The National Superficial Deposit Thickness Model (version 5). British Geological Survey Internal Report, OR/09/049. 18pp.
- Ravazzani, G., Rametta, D. and Mancini, M., 2011. Macroscopic cellular automata for groundwater modelling: A first approach. Environmental Modelling & Software, 26(5), pp.634-643.
- Raven, E.K., Lane, S.N. and Bracken, L.J., 2010. Understanding sediment transfer and morphological change for managing upland gravel-bed rivers. Progress in Physical Geography, 34(1), pp.23-45.
- Rushton, K.R. & Eilers, V.H.M. & Carter, Richard. 2006. Improved Soil Moisture Balance Methodology for Recharge Estimation. Journal of Hydrology. 318. 379-399. 10.1016/j.jhydrol.2005.06.022.
- Schiff, J.L., 2011. Cellular automata: a discrete view of the world (Vol. 45). John Wiley & Sons.
- Sear, D.A. and Milne, J.A., 2000. Surface modelling of upland river channel topography and sedimentology using GIS. Physics and Chemistry of the Earth, Part B: Hydrology, Oceans and Atmosphere, 25(4), pp.399-406.

SEPA. 2018. SEPA Water Levels - Feshie @ Feshie Bridge. [online] sepa.org.uk. Available at:

http://apps.sepa.org.uk/waterlevels/default.aspx?sd=t&lc=234217 [Accessed 30 Jan. 2018].

- Shintani, C., 2016. Comparing photogrammetric and spectral depth techniques in extracting bathymetric data from a gravel-bed river.
- Sibson, R. and Barnett, V., 1981. Interpreting multivariate data. A brief description of natural neighbor interpolation, pp.21-36.
- Smith, M. and Vericat, D., 2015. From experimental plots to experimental landscapes: topography, erosion and deposition in sub-humid badlands from Structure-from-Motion photogrammetry. Earth Surface Processes and Landforms, 40(12), pp.1656-1671.
- Sólyom, P.B. and Tucker, G.E., 2004. Effect of limited storm duration on landscape evolution, drainage basin geometry, and hydrograph shapes. Journal of Geophysical Research: Earth Surface, 109(F3).
- Steduto, P., Hsiao, T.C., Fereres, E. and Raes, D., 2012. Crop yield response to water (Vol. 1028). Rome: fao.
- Strahler, A.N., 1952. Hypsometric (area-altitude) analysis of erosional topography. Geological Society of America Bulletin, 63(11), pp.1117-1142.
- Tanguy, M.; Dixon, H.; Prosdocimi, I.; Morris, D. G.; Keller, V. D. J. (2016). Gridded estimates of daily and monthly areal rainfall for the United Kingdom (1890-2015) [CEH-GEAR]. NERC Environmental Information Data Centre.
- Tarboton, D. G., R. L. Bras, and I. Rodriguez-Iturbe. 1991. "On the Extraction of Channel Networks from Digital Elevation Data." Hydrological Processes 5: 81-100.
- Thomas, R. and Nicholas, A.P., 2002. Simulation of braided river flow using a new cellular routing scheme. Geomorphology, 43(3-4), pp.179-195.
- Toffoli, T. and Margolus, N., 1987. Cellular automata machines: a new environment for modeling. MIT press.
- Toffoli, T., 1984. Cellular automata as an alternative to (rather than an approxima- tion of) differential equations in modeling physics. Phys. Nonlinear Phenom. 10 (1e2), 117e127.
- Tonkin, T. and Midgley, N., 2016. Ground-control networks for image based surface reconstruction: an investigation of optimum survey designs using UAV derived imagery and structure-from-motion photogrammetry. *Remote Sensing*, 8(9), p.786.
- Uren, J. and Price, W.F., 1994. Surveying for engineers. Macmillan International Higher Education.
- Van De Wiel, M.J., Coulthard, T.J., Macklin, M.G. and Lewin, J., 2007. Embedding reach-scale fluvial dynamics within the CAESAR cellular automaton landscape evolution model. Geomorphology, 90(3-4), pp.283-301.
- Veihmeyer, F.J.; Hendrickson, A.H. (1931). "The moisture equivalent as a measure of the field capacity of soils". Soil Science. 32 (3): 181-193.
- Von Neumann, J. and Burks, A.W., 1996. Theory of self-reproducing automata (p. 8). Urbana: University of Illinois Press.
- von Neumann, J., 1951. The general and logical theory of automata. Cereb. Mech. Behav. 1e41.
- Wang, Lei; Barkwith, Andrew; Jackson, Christopher; Ellis, Michael. 2012 SLiM : an improved soil moisture balance method to simulate runoff and potential groundwater recharge processes using spatio-temporal weather and catchment characteristics. [Lecture] In: 12th UK CARE Annual General Meeting, UK

Chinese Association of Resources and Environment, Bristol, UK, 8 Sept 2012. (Unpublished)

- Wechsler, S.P. and Kroll, C.N., 2006. Quantifying DEM uncertainty and its effect on topographic parameters. Photogrammetric Engineering & Remote Sensing, 72(9), pp.1081-1090.
- Westaway, R., Lane, S. and Hicks, D., 2003. Remote survey of large-scale braided, gravel-bed rivers using digital photogrammetry and image analysis. International Journal of Remote Sensing, 24(4), pp.795-815.
- Westaway, R.M., Lane, S.N. and Hicks, D.M., 2000. The development of an automated correction procedure for digital photogrammetry for the study of wide, shallow, gravel-bed rivers. Earth Surface Processes and Landforms, 25(2), pp.209-226.
- Westaway, R.M., Lane, S.N. and Hicks, D.M., 2001. Remote sensing of clear-water, shallow, gravel-bed rivers using digital photogrammetry. Photogrammetric Engineering and Remote Sensing, 67(11), pp.1271-1282.
- Westoby, M., Brasington, J., Glasser, N., Hambrey, M. and Reynolds, J., 2012. 'Structure-from-Motion' photogrammetry: A low-cost, effective tool for geoscience applications. Geomorphology, 179, pp.300-314.
- Wheaton, J.M., Brasington, J., Darby, S.E. and Sear, D.A., 2010. Accounting for uncertainty in DEMs from repeat topographic surveys: improved sediment budgets. Earth surface processes and landforms: the journal of the British Geomorphological Research Group, 35(2), pp.136-156.
- Wilcock, P.R. and Crowe, J.C., 2003. Surface-based transport model for mixedsize sediment. Journal of Hydraulic Engineering, 129(2), pp.120-128.
- Williams, R., 2012. DEMs of difference. Geomorphological Techniques, 2(3.2).
- Williams, R., Brasington, J., Vericat, D., Hicks, M., Labrosse, F. and Neal, M., 2011. Monitoring braided river change using terrestrial laser scanning and optical bathymetric mapping. In Developments in Earth Surface Processes (Vol. 15, pp. 507-532). Elsevier.
- Williams, R., Lamy, M.L., Stott, E. and Maniatis, G., 2017, December. A walk by the river: three-dimensional reconstruction of surface sedimentology and topography using wearable laser scanning. In AGU Fall Meeting Abstracts.
- Williams, R.D., 2014. Two-Dimensional Numerical Modelling of Natural Braided River Morphodynamics (Doctoral dissertation, PhD Thesis, Aberystwyth University, Aberystwyth, Ceredigion, UK).
- Williams, R.D., Brasington, J. and Hicks, D.M., 2016. Numerical modelling of braided river morphodynamics: Review and future challenges. Geography Compass, 10(3), pp.102-127.
- Williams, R.D., Brasington, J., Hicks, M., Measures, R., Rennie, C.D. and Vericat, D., 2013. Hydraulic validation of two-dimensional simulations of braided river flow with spatially continuous aDcp data. Water Resources Research, 49(9), pp.5183-5205.
- Wolman, M.G., 1954. A method of sampling coarse river-bed material. EOS, Transactions American Geophysical Union, 35(6), pp.951-956.
- Woodget, A.S., Austrums, R., Maddock, I.P. and Habit, E., 2017. Drones and digital photogrammetry: from classifications to continuums for monitoring river habitat and hydromorphology. Wiley Interdisciplinary Reviews: Water, 4(4), p.e1222.
- Woodget, A.S., Carbonneau, P.E., Visser, F. and Maddock, I.P., 2015. Quantifying submerged fluvial topography using hyperspatial resolution UAS imagery and structure from motion photogrammetry. Earth Surface Processes and Landforms, 40(1), pp.47-64.
180

- Young, A.R., 2006. Stream flow simulation within UK ungauged catchments using a daily rainfall-runoff model. Journal of Hydrology, 320(1-2), pp.155-172.
- Zhu, D., Ren, Q., Xuan, Y., Chen, Y. and Cluckie, I.D., 2013. An effective depression filling algorithm for DEM-based 2-D surface flow modelling. Hydrology and Earth System Sciences, 17(2), pp.495-505.
- Ziliani, L., Surian, N., Coulthard, T.J. and Tarantola, S., 2013. Reducedcomplexity modeling of braided rivers: Assessing model performance by sensitivity analysis, calibration, and validation. Journal of Geophysical Research: Earth Surface, 118(4), pp.2243-2262.

						-								
									Sample size for zero bias	Precision sample	ā	Lower recision Pr	Upper Pri ecision	ecision
2 SAMPLE Feshie Upstream Sub-Surface					_	Date	01/02/2018		(kg)	size (kg)		Limit	Limit	as %
3			Parameter	Ē	psi	Bias	Precision							
4 Cumulative %			م	4.16	2.06	0.0	0.02		0.06	1		₽ .4	4.21	0.82%
- I00 -		Į	9	14.46	3.85	0.0	0.04		0.43	7.3		14.02	14.92	117%
9	-		20	59.97	5.91	0.0	0.20		8.71	147.4		52.12	68.39	3.427
-			\$	95.11	6.57	0.00	0.50		176.35	2983.7		67.17	134.68	7.64%
8			6	115.15	6.85	0.00	0.50		420.21	7109.7		81.32	163.06	7.33%
- 08			35	147.60	7.21	FALSE	0.50		1260.87	21333.0		104.24	209.01	6.96%
0														
	-		mean	43.53	5.44									
- 90 -			sorting	:	1.46									-
			skewness	toward large	-0.50									
	 '■r 		(D84/D16)°	ب	2.56									
0	<			ATA A	Ì		0	4	las Care.	l				+
- 0+				NIN			Ĺ	aramet	Jeg Error	Ē	202			+
					Mass etained /									
	•			- E	ka			ы	0.05	4.16	0.05			
20.				724.0	00			φ	0.44	14.46	0.46			-
6				511.9	0.0			20	7.84	59.97	9.02			
••• 	_			362.0	0:0			84	27.94	95.11	39.57			
•••				256.0	0:0			6	33.83	115.15	47.91			
2		-		181.0	23.3			95	43.36	147.60	61.41			
3 -4 -3 -2 -1 0 1 2 3	4 5 6	8 9 10 11		128.0	36.8									
	-			90.5	97.8									
Grain size / ps	150			64.0	269.0				Precision	al Error - (Jpstream :	SubSurfa	Ge	
9				45.3	120.8			250.00						
7 % in class				32.0	80.5			norne v						
65 1 55 1				22.6	69.3									_
- 0+				16.0	49.9			200.00						
		29.9			292									
				5.7 5.7	28.2		104	- 150.00					-	56
30 20 1				40	234								-	
4 15 -		13.4 10.0		2.8	22.5		erei 2	100.00					-	-
5 10 1	5.6 7.7	9.0		2.0	19.8								1	
6 5 1 a a a a a a a a a a a 2 2 2 2 6 3.1	1 2.4 4.0	4.1 2.6 0.0 0.0	00 00	<2.0	0:0						9 50		-	
				8	8			20.00			-			
<1.0 15 25	5 33 4.5	5.5 6.5 7.5 8.5	5 9.5	00	0.0									
0	terio sino (nui			00	8			0.00	- - -	-	-	-		
	וומות אמוני המו			8	8			0	20	007	9	8	80	100
				8	8						Parameter			
2				0.0	8									
~				TOTAL	899.Z]-							H
		-				-		-				-		

10 Appendix

1. Statistical analysis of bulk sediment data.

18	32		1			1	1	1	1																																		
:	recision as X		0.84%	129%	4.61%	7,18%	6,90%	6.68%						_														-		-	56	5	T	-1				000	-				
Upper	Precision P		54	2141	85 49	224.49	276.26	328.65																							+		-		1			UK	ana a				
Lower	Precision F		4.98	19.80	57.77	108.69	133.75	159.12																		ror										20		2	-	ter			
															So So			0.07	0.82	15.21	68.29	84.03	99.97			isional Er									+	•		υ	4	Farame			
Precision	sample cito (t.c.)	(Fy) and	14	116	282.3	7348.3	18640.2	60537.1							Ē			5.04	20.59	70.28	156.20	192.22	228.68			Prec											16	02	3				
Sample size	for zero bias	(64)	0.07	0.60	15.03	377.92	958,65	3113.36							leg Error			0.07	0.79	12.50	47.51	58.47	69.56														•		2				
	-														Paramet D			a	16	50	84	6	95				35.0.00	0000000	30.0.00	25.0.00	(200.00	() 12000	215	10 0.00	5		0.00					
Γ	01/02/2018	Precision	0.02	0.06	0.28	0.52	0.52	0.52																																			
Ī	Date	Biac	000	000	80	000	000	FALSE																																			
Γ		isi	233	4 36	613	7.29	7.59	7.84		5.93	1.56	-0.30	2.75			Mass	retained /	ğ	0.0	0:0	0.0	17.4	99.0	<u> 9</u> 9.6	144.1	186.4	112.4	89.1	74.0	42.5	30.0	17.1	24.3	20.6	19.8	16.0	0.0	0.0	0.0	0.0	0.0	0.0	992.2
		Ē	5.04	20.59	70.28	156.20	192.22	228.68		60.91		toward large	2		DATA			шш	724.0	511.9	362.0	256.0	181.0	128.0	30.5	64.0	45.3	32.0	22.6	16.0	10.0	8.0	5.7	4.0	2.8	2.0	<20 20	33	200	80	0.0	0.0	TOTAL
		Parameter	5	. ¥	202	3	8	32		mean	sorting	skewness	(D84/D16)°																								0.00	[30	Ŋ,				
-																																					1.8 0.0 0		0.8				
╞					•																			10										2 11 0	100 100			30	C)				
t		ŀ				<u> </u>																		8										8.8	11.3	-			ŋ				
-											•	_		-	<			_	_	_	_	_	-	6												22 27	Ç		0 1	ц.			
																				ĸ	-	_		4	-	IEd											2.5 1.7 3.0		3	Grain size / p:			
T																					1			3		Grain size											2.0 2.1		1	Ū			
	am Surfae																							0													0.0 0.0 1.6		0.1-2				
	hie Unstre		2	2																			-														0.0 0.0 0.0						_
	IPI E Fee		Cumulstine 6	100 -	201	,		- 08			4	- 00	-		40 -		,		20 -	•	,		9	4				% in class	5]		5 -		5 -		5		5 0.0 0.0						
	SAM							_	_	_										-	_							-	4	4	- -	ñ		2		H							



$\left(1 - \frac{1}{16416} \right) = \frac{1}{16416} \right) = \frac{1}{16416} \left(1 - \frac{1}$							Sample For zero	size Pre. bias sar	ision nole	Precisi	on Precis	Sion Pre	ecision
$ \left(\begin{array}{cccccccccccccccccccccccccccccccccccc$	PLE Feshie Downstream Sub-Surface				Date	01/02/2018	(kg)	size	(kg)	Limit	Limit	ŝ	
$\left(\begin{array}{c} 1 \\ 1 \\ 1 \\ 1 \\ 1 \\ 1 \\ 1 \\ 1 \\ 1 \\ 1 $		Paramete	en mm	psi	Bias	Precision							
$\left(\begin{array}{c} 1 \\ 0 \\ 0 \\ 0 \\ 0 \\ 0 \\ 0 \\ 0 \\ 0 \\ 0 \\$	Cumulative %	0	3.41	1.77	0.00	0.01		0.01	0.2	ri	33	3.43	$0.47 \times$
$\left(\begin{array}{cccccccccccccccccccccccccccccccccccc$		φ	9.00	3.17	0.00	0.03		0.08	1.8	œ	-84	9.16	0.80%
$ \left\{ \begin{array}{cccccccccccccccccccccccccccccccccccc$		20	37,90	5.24	0.00	0.14		2.50	54.6	те В	38	1.78	2.68%
$ \left\{ \begin{array}{cccccccccccccccccccccccccccccccccccc$	•	84	91.74	6.52	0.00	0.54		76.15	1663.0	3	02 13	3.55	8.317
$ \begin{bmatrix} 36 & 4873 & 783 & 000 & 064 \\ 1 & 7163 & 4873 & 783 & 000 & 064 \\ 1 & 7163 & 4873 & 783 & 784 & 000 & 064 \\ 1 & 7163 & 7163 & 7163 & 7163 & 7163 & 7163 \\ 1 & 7163 & 7163 & 7163 & 7163 & 7163 & 7163 \\ 1 & 7163 & 7164 & 7163 & 7164 & 7163 & 7164 & 7163 & 7164 \\ 1 & 7163 & 7164 & 7163 & 7164 & 7164 & 7164 & 7164 & 7164 \\ 1 & 7164 & 71$		6	113.15	6.82	0.00	0.54	~	04.23	1459.8	22	73	4.72	$7.94 \times$
$ \left\{ \begin{array}{c c c c c c c c c c c c c c c c c c c $	80 -	35	145.79	7.19	000	0.54		711.63 15	540.4	- 10	0.15 2.1	2.24	7.54%
$ \left(\begin{array}{c} 0 \\ 0 \\ 0 \\ 0 \\ 0 \\ 0 \\ 0 \\ 0 \\ 0 \\ 0 $													
$ \left(\begin{array}{c} 0 \\ 0 \\ 0 \\ 0 \\ 0 \\ 0 \\ 0 \\ 0 \\ 0 \\ 0 $		mean	3151	4,98									
$ \frac{1}{2} = 1$		sorting		166									
$ \begin{pmatrix} 0 \\ 0 \\ 0 \\ 0 \\ 0 \\ 0 \\ 0 \\ 0 \\ 0 \\ 0$	- 09	skewnes	s toward laro	-0.26									
$ \frac{1}{2} = \frac{1}{2} + 1$	•	[D84/D16	1.5	3.19									
$ \begin{array}{c} 1 \\ 1 \\ 2 \\ 2 \\ 2 \\ 2 \\ 2 \\ 2 \\ 2 \\ 2 \\$			2										
$ \begin{array}{c c c c c c c c c c c c c c c c c c c $			DATA			Paran	net Ned Fro	Ē	у С				
$\sum_{k=1}^{2} \left(\frac{e^{2i}}{1} + \frac{1}{2} + \frac{1}$	- 0+			Mass		-	0111 Foot		3			-	
$ \left\{ \begin{array}{cccccccccccccccccccccccccccccccccccc$				retained /									
$ \frac{2^{2}}{2^{2}} - \frac{2^{2}}{2^{2}} - \frac{1}{2^{2}} - \frac{1}{$			Ē	ka		2		0.02	3.41	0.02			
$ \left\{ \begin{array}{cccccccccccccccccccccccccccccccccccc$			724.0			9		0.16	9,00	0.16			
$ \frac{1}{2} + 1$			511.9	8		20		3.52	37.90	3.88			
$ \frac{1}{2} + \frac{1}{3} + \frac{1}{2} + \frac{1}{3} + \frac{1}{2} + 1$			362.0	00		8		28.72	91.74	41.81			
$ \frac{1}{2} - \frac{1}{2} - \frac{1}{2} - \frac{1}{2} + 1$			256.0	00		6		35.43	113.15	51.57			
$ \begin{array}{c ccccccccccccccccccccccccccccccccccc$		[181.0	20.4		36		45.64	145.79 (36.45			
Crain size / pi 015 734 Precisional Error % in class 63.0 0.042 0.042 0 63.0 0.042 0.042 0.042 0 730.0 63.0 0.042 0.03 0.03 0 730.0 63.2 0.033 0 0.03 0 0 730.0 52.1 0.03 0.03 0.03 0.03 0.03 0.03 0.04	4 -3 -2 -1 0 1 2 3 4 5 6 7 8 9 10	11	128.0	315									
Gram size / pat Gram size / pat Frecisional Error % 1 2 003 0			90.5	79.4									
$ \frac{1}{2} 1$	Grain size / psi		64.0	104.2					Precision	al Error			
$ \frac{1}{26} \text{ find task } 1$			45.3	109.3									
$ \begin{array}{ c c c c c c c c c c c c c c c c c c c$	% in class	-	32.0	109.3			52 0,00				-		
$ \begin{bmatrix} 160 & 632 \\ 0 & 221 \\$			22.6	86.7									F
$ \begin{bmatrix} 100 & 521 \\ 80 & 235 \\ 80 & 347 \\ 10 & 10 & 10 & 0 \end{bmatrix} = \begin{bmatrix} 100 & 521 \\ 80 & 235 \\ 10 & 10 & 235 \\ 10 & 23 & 237 \\ 10 & 23 & 237 \\ 10 & 23 & 237 \\ 10 & 23 & 237 \\ 10 & 23 & 237 \\ 10 & 23 & 237 \\ 10 & 23 & 237 \\ 10 & 23 & 237 \\ 20 & 0 & 0 & 0 \end{bmatrix} = \begin{bmatrix} 100 & 0 & 0 & 0 & 0 & 0 \\ 10 & 20 & 250 & 0 & 0 & 0 \\ 10 & 20 & 250 & 0 & 0 & 0 \\ 10 & 20 & 250 & 0 & 0 & 0 \\ 10 & 20 & 250 & 0 & 0 & 0 \\ 10 & 20 & 20 & 0 & 0 & 0 \\ 10 & 20 & 20 & 0 & 0 & 0 \\ 10 & 0 & 0 & 0 & 0 & 0 & 0 \\ 10 & 0 & 0 & 0 & 0 & 0 \\ 10 & 0 & 0 & 0 & 0 & 0 \\ 10 & 0 & 0 & 0 & 0 \\ 10 & 0 & 0 & 0 & 0 & 0 \\ 10 & 0 & 0 & 0 & 0 & 0 \\ 10 & 0 & 0 & 0 & 0 & 0 \\ 10 & 0 & 0 & 0 & 0 & 0 \\ 10 & 0 & 0 & 0 & 0 & 0 \\ 10 & 0 & 0 & 0 & 0 & 0 \\ 10 & 0 & 0 & 0 & 0 & 0 \\ 10 & 0 & 0 & 0 & 0 & 0 \\ 10 & 0 & 0 & 0 & 0 & 0 \\ 10 & 0 & 0 & 0 & 0 & 0 \\ 10 & 0 & 0 & 0 & 0 & 0 \\ 10 & 0 & 0 & 0 & 0 & 0 \\ 10 & 0 & 0 & 0 & 0 & 0 \\ 10 & 0 & 0 & 0 & 0 & 0 \\ 10 & 0 & 0 & 0 & 0 & 0 \\ 10 &$	- 0		16.0	69.2		-1	00.000						
$ \begin{array}{c ccccccccccccccccccccccccccccccccccc$	5		10.0	52.1									+
$ \begin{array}{c ccccccccccccccccccccccccccccccccccc$	0 -		8.0	23.5		ш.	15 0.00						•
$ \begin{array}{c ccccccccccccccccccccccccccccccccccc$	5.1		5.7	34.7		u) a						-	5
$ \begin{array}{c ccccccccccccccccccccccccccccccccccc$	0 1		4.0	28.3		IZIS	100.00					84	-
$ \begin{array}{c ccccccccccccccccccccccccccccccccccc$	5		2.8	27.7								_	
$ \begin{array}{c ccccccccccccccccccccccccccccccccccc$	0		20	25.0			00.02			1	+		
0 00 <t< td=""><td>3.1 3.5 4.3 2.9</td><td>9 2.5</td><td><2.0</td><td>8</td><td></td><td></td><td></td><td></td><td></td><td>20</td><td></td><td></td><td></td></t<>	3.1 3.5 4.3 2.9	9 2.5	<2.0	8						20			
Cline 15 25 33 45 55 65 75 85 95 00 00 00 00 00 00 00 00 00 00 00 00 00		0.0 0.0 0.0 0.0	0.0	00			000	910			_		
Grain size / psi 0.0 0.0 0.0 0.0 0.0 0.0 0.0 0.0 0.0 0.0 0.0 1 0.0 0.0 0.0 0.0 0.0 0.0 0.0 0.0 1 0.0 0.0 1 0.0 0.0 1 0.0 0.0	<10 15 25 33 45 55 65	7.5 8.5 9.5	0.0	8			0	0 20	30 40	50 60	R	6	90 10
Grain size / psi			0.0	0.0						arsmatar			
00 00 00 00 00 00 00 00 00 00 00 00 00	Grain size / psi		00	0:0					-	1 1 1 1 1 1 1 1 1 1 1 1 1 1 1 1 1 1 1			
00 00 TOTAL 8012			0.0	8			_		_			-	
TOTAL 8012			0:0	00									
			TOTAL	801.2									