Modelling basin-scale sediment dynamics in the Petit lac d'Annecy catchment, France

Thesis submitted in accordance with the requirements of the University of Liverpool for the degree of Doctor in Philosophy

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In memory of my four Grandparents

Hannah and Walter Fairclough

Marie and William Welsh

Modelling catchment processes in pre-Alpine France

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Changes in land management coupled with projected trends in climate could have significant impacts for the hydrology and sediment regimes within the pre-Alpine zone over the present century, particularly in terms of increased flooding and sedimentation. Previous attempts to discern the relative long-term contributions of climate and land use drivers on hydro-geomorphological from studies of palaeoarchives are hampered by the integrated and compounded nature of proxy signals. This research describes the testing, development and application of an established hydro-geomorphic numerical model (CAESAR) over different time scales (170-2000 years) to simulate at hourly time resolution, the changes in the hydrological and sediment regime of the Petit lac d'Annecy catchment in response to changes in system drivers. The cellular automata nature of CAESAR renders it an ideal tool for studying complex environments in a holistic manner, as it allows macro-scale emergent behaviour, such as the development of alluvial fans, from micro-scale physical process laws.

The hourly drivers of the model, precipitation, temperature and forest cover, have been derived a combination of instrumental data, documentary records and lake sediment records. The outputs of the model have been compared to field-based catchment geomorphology, empirical evidence and over longer-scales, against proxy records of detrital sediment in the catchment. Model developments included improving hillslope process representation which have been rigorously tested and compared with field evidence and known process rates. Additionally, a 'snow mass balance model' has been developed to imitate the effect of snow melt and snow store in the catchment.

Results from model runs over a few centuries show that in general the relationship between forest cover and total sediment discharge is nonlinear. Multiple-scenarios were modelled in order to quantify changes in the system. Forest cover scenarios suggest that flood peaks under forest cover above 60% (similar to current levels) may be too low to access overbank sediments in low, wide morphological settings. This is likely to result in channel incision occurring over decadal-centennial timescales accompanied by increased sediment discharge In narrow, well coupled sub-catchments, an increase in forest cover from ~20% to ~40% over a 2000 year timescale shows a 37% decrease in sediment discharge, whereas a further increase of 20% (i.e. 40% to 60%) shows a decrease of only 17%. The results tentatively suggest that there may be threshold levels of forest cover. Different climate scenarios show the effects of increasing precipitation on catchment geomorphology. For example, a 33% increase in precipitation over a 500 year timescale can drive gully

incision by up to 50% (8m deep to 12m deep), with a signficant increase in the extent of slope failure zones and up to 2m of fan head incision.

Two thousand year long model runs in five different morphological settings were simulated, the results suggest that intrinsic system behaviour such as storage-release, hillslope-channel coupling and supply-capacity relationships may well exert larger controls on sediment discharge patterns over this timescale than climate or land use drivers. The combined results from all five sub-catchments were compared to lake sediment proxy records and overall behaviour and temporal patterns compare well with the sediment accumulation rates and sediment proxies.

Hypothetical scenarios to investigate the geomorphic implications of a snow-free pre-alpine region over the last 2000 years show that there would be around 1.4 times more sediment discharge, with the annual hydrological regime radically altered with increased flooding throughout the year, particularly in winter months and a lack of a sustained discharge peak in the 'melt' months. This has implications for the projected environmental changes over the coming decades. The simulated effects of increased precipitation, reduced forest cover and snow-free conditions, in combination, point to increased amounts of coarse sediment discharge within the channels. Broad estimations show that a 20% reduction in forest cover or snow-free conditions can result in an additional 1m of sediment moving through the system and accumulating in the lake with potentially large impacts on flooding, in-channel fauna, benthic-dwelling lake fauna, aquatic macrophytes and water quality and water availability for storage and local power generation. Overall, CAESAR can be considered a robust tool for future predictive simulation modeling of hydro-geomorphic changes.

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AOGCM	Atmospheric-Oceanic General Circulation Models
RCM	Reduced Complexity Model
IPCC	Intergovernmental Panel on Climate Change
CAESAR	Cellular Automaton Evolutionary Slope and River
AD	Anno Domini
SRES	Special Report on Emission Scenarios
CAP	Common Agricultural Policy
Cal yr BP	Calendar years before present
LA 13	Lake Annecy 13 Sediment Core
DEM	Digital Elevation Model
ASCII	American Standard Code for Information Interchange
BP	Before Present
LGM	Last Glacial Maximum
MIS	Marine Isotope Stage
SMB	Snow Mass Balance

Chapter 1

Introduction and Aims

1.1 Research Context

Growing international demographic pressures on land (Messerli et al., 2000) combined with unprecedented changes in climate (IPCC, 2007) suggest an uncertain present and future for many global ecosystems. Many of the Earth's systems are large, inter-connected systems with many component parts and therefore we as researchers do not fully understand the intricacies of such systems, nor the effects that minor perturbations may have on these highly non-linear environments. Hence, these systems are decidedly unpredictable and this renders mitigation against potential hazards difficult. Mountain regions such as the European Alps are particularly sensitive to change due to rapid variations in climate, vegetation and hydrology over short horizontal distances (Whiteman, 2000) and this therefore makes them an ideal zone for impact assessment of the future changes that may occur (Beniston, 2003).

One system that is likely to be affected by changing climate and land use is the fluvial system. Changes in climate and/or land use can significantly alter hydrological regimes (Knox, 1993, 1995), which may lead to channel instability, increased flooding and changes in type and volume of sediment loads (Knighton, 1998). These changes have significant impacts on society, agriculture, tourism and economies at local, regional and in some cases global levels. Clarke et al., (1985) estimated that the downstream impacts of soil erosion alone have cost the United States $6100 million (at 1980 values). These impacts are a particular concern for mountain regions as 26% of the world's population live in sensitive mountain regions (Meybeck et al., 2001) and provide food for over 50% of the world's population (Beniston, 2003).

Climate projections, land use projections and impact studies e.g. projected runoff tend to operate at a continental or regional scale (e.g. IPCC, 2007). Whilst these scales are useful for providing general trajectories of change, extremes in precipitation and temperature that can be recorded at a local level are likely to obscured in the global and regional scale projections yet it is often the extreme changes that have the greatest impacts on environments.

Of particular interest at local scale is the geomorphic response to changes in future climate and land use, in particular how the varying incidence and magnitude of flood events will alter process rates, sediment delivery and storage and release cycles at catchment-scale. It is vital to understand how system behaviours may change, so that these zones and any potential hazards can be managed most effectively in the future. For this type of study 'catchment-scale' is ideal. A catchment or drainage basin is de-lineated by a watershed, which is usually an elevated ridge in the landscape and represents the boundary between two catchments. The two main sub-systems within a catchment are the fluvial system and the hillslope system. De Boer (2001) suggests that any understanding of how a drainage basin functions as a system should be analysed at the basin level rather than in simple component parts (e.g. considering the fluvial and hillslope systems holistically). Mulligan and Wainwright (2004:108) describe the drainage basin as "*a cog in the machine that is the hydrological cycle*". Within a catchment, there will be inputs (e.g. precipitation), stores (e.g. alluvial fans), transfers (e.g. water routing and sediment entrainment), outputs (e.g. geomorphological features), internal controls and external forcings all of which need to be considered (Schumm, 1977).

The main difficulty with this type of study is that the impacts on the system are not likely to be a result of either climate or land use; fundamentally it is the relationship between the two drivers that is essential. In addition to this the hydrological and sediment responses identified in palaeoarchives carry a mixed signal thus making it extremely difficult to determine the relative contributions of the impact of climate and land use drivers. Many researchers have attempted to ascertain the relative contributions of climate and land use through various methods (see chapter 2), yet quantification of the role of each system driver remains elusive. The extent to which changes in geomorphic regime is controlled by climate and land use continues to be one of the key questions in Holocene geomorphology (Chiverrell et al., 2007). Unravelling the contributions of each driver is fundamental for sustainable management both now and in the future.

1.2 Approach

Oldfield (2005) calls for the need to *"bridge contemporary and palaeoresearch, empirical and modelling approaches and biosphysical and socio-economic perspectives if a sustainable future can be achieved"*.

This research goes some way to addressing these issues by using and developing an established hydro-geomorphic cellular model (CAESAR: Coulthard, 1999, Van de Wiel et al., 2007) to assess changes in sediment transfer and budgets within a lake catchment in response to changes in system drivers (climate and land use). The modelled data are then rigorously tested and compared to short-term water discharge data, long-term lake sediment records and field evidence from catchment geomorphology to assess robustness of the model. CAESAR is then used for several hypothetical simulations to investigate the effects of changing temperatures, precipitation and land use.

1.3 Research Aims

The overriding aim of this research is to rigorously test and compare temporal and spatial output from a catchment-scale hydro-geomorphological model (CAESAR) to observed short-term instrumental data, long-term lake sediment records and catchment geomorphology in order to verify the model so that it can be used with future scenarios to identify the impact of future environmental change.

The study will aim to address a number of general and more specific questions such as:

- How well do short term instrumental water discharges compare to modelled water discharge?
- How well do modelled erosion and deposition patterns compare to empirical data and field observations of catchment geomorphology?
- To what extent can the relative contributions of climate and land use be quantified?
- How much of a role does catchment morphometry play in determining sediment delivery?
- What are the potential geomorphic impacts of a 'snow-free' Alps?

- By how much does a 20% reduction or 20% increase in forest cover affect sediment delivery?

- How well do behaviours within long term records of modelled sediment and lake sediment records compare?

- Can the model be used as a predictive tool to identify major future changes?

1.4 Structure of thesis

This thesis has eleven chapters, and following this introduction, the main part of this thesis focuses on applying and comparing sediment discharge data from a numerical model to proxies of environmental change from a lake sediment record in order to ensure the model behaviour is robust and can be used with confidence with future projections of land use and climate change to investigate local-scale hydro-geomorphic impacts. Each of the results chapters (6-10) are largely self contained with a context and brief method relevant to that chapter, results, discussion and conclusions within each.

Chapter 2 details landscape sensitivity to the main system drivers: climate and land use and goes on to explore previous attempts at ascertaining the relative contributions of each.

Chapter 3 reviews characteristics of complex systems and details the type of models needed to simulate complexity. It also provides a background to the evolution of cellular automata models, reduced complexity models and finally introduces CAESAR the numerical model used in this thesis.

Chapter 4 details the regional physical characteristics and describes the Petit lac d'Annecy catchment and history of environmental change from literature, historical documents and field observations, including a detailed description of each sub-catchment.

Chapter 5 describes the overall methodology for the research and provides detailed accounts of the methods used to create hourly input files for modelling from palaeorecords and historical documents. It also covers practical details on model setup and initial conditions.

Chapter 6 presents the first results from two sub-catchments for a 170 year period. This provided some insight to hydrological and sediment response of systems to drivers, and a major conclusion from this chapter suggests that the hillslope processes of CAESAR (e.g. Coulthard, 1999, Van de Wiel, 2007) are represented too simplistically to apply the model to a pre-Alpine environment where hillslope processes are prevalent. This chapter is from a published paper:

Welsh K.E., Dearing, J.A., Chiverrell, R.C., Lang, A. (2009) Testing a cellular modelling approach to simulating late-Holocene sediment and water transfer from catchment to lake in the French Alps since 1826. **Holocene**, 19(5)

Chapter 7 builds on the conclusions from chapter six and details the development of hillslope process representation within CAESAR model. In addition this chapter details how the modifications were then rigorously tested and the outputs are then compared to a set of results from the original CAESAR model. It also assesses how realistic the behaviour of the new hillslope processes is when compared to empirical data. Finally comparisons are made to sediment accumulation rates from the Petit lac d'Annecy to identify how successfully the new hillslope processes can predict sediment discharge.

Chapter 8 describes the application of CAESAR under two scenarios to investigate a hypothetical 'snow-free' pre-Alps for the last 2000 years in two sub-catchments. It then details the hydrological and sediment response under a typical precipitation regime and one of a hypothetical precipitation regime that does not permit any snow storage or melt thereby providing an insight into long term potential impacts of increasing temperatures and reduced snow cover at catchment scale.

Chapter 9 investigates the sensitivity of land use changes and quantifies sediment responses under different land use regimes over 2000 years for two sub-catchments.

Chapter 10 describes the application of CAESAR to all sub-catchments in the Petit lac d'Annecy and is the main results chapter that compares modelled sediment to proxies from lake sediment archives for a 2000 year period in to identify how well behaviours in modelled sediment discharge compare to behaviours identified in lake sediments.

Chapter 11 is the main discussion chapter that draws together the key findings of all the results chapters and discusses the wider implications. Finally it considers future research directions in light of the findings from this thesis.

Chapter 2

Landscape sensitivity to climate and land use: past and future approaches

2.1 Landscape sensitivity to changes in climate

Recent, rapid, global climate change is well documented (e.g. IPCC, 2007) and in recent years direct observations have been made which state that eleven out of twelve of the years between 1995 and 2006 were the warmest since the global surface temperature records began in 1850. The IPCC report suggests that this increase has a 90% likelihood of being a direct result of anthropogenic activity rather than natural climate change through variations in solar radiation or volcanic activity. The report suggests that the rise in temperatures is a result of increased greenhouse gas emissions (primarily carbon dioxide) and fossil fuel combustion since the industrial revolution (~AD 1850). Additionally changes in land use (e.g. deforestation) are a significant cause of climate change but have less pronounced effects. Although climate change has always occurred through multiple glacial-interglacial cycles, the major concern is the current rate and magnitude of change in temperatures and greenhouse gases. Over the last 650 000 years, evidence from ice core records suggest that present day carbon dioxide levels (IPCC, 2007) and methane (Spahni et al., 2005) exceed the natural range. These concerns are further compounded when considering the best estimate of future changes from fully coupled atmospheric-oceanic general circulation models (IPCC, 2007) for a range of emissions scenarios (SRES). For the various SRES scenarios, the models project that there is very little chance that the global warming will be under 1.5°C before AD 2100. The upper limit is likely to be around 3°C, yet the higher temperatures (~5.8°C) projected for some scenarios (A1F1) cannot be ignored.

With increasing temperatures, changes in precipitation are also likely to occur. Recent IPCC reports (2007) suggest that for European Alpine regions, 90% of the climate models concur that winter precipitation AD 2090-2099 will increase by 10-20% compared with the same period in AD 1980-1999. For the same period, summer precipitation will decrease by approximately 20%. The IPCC also conclude that extreme precipitation events will become more frequent. Increased precipitation combined with increasing temperatures (and therefore increased snow melt in mountain regions) may have major implications for fluvial systems such as increased flooding due to increased magnitudes of peak flows (e.g. Middelkoop et al., 2001; Milly et al., 2002) and greater frequency and magnitude of high discharge events (e.g. Hunt, 2002). Whilst the focus has remained largely on discharge, the changes in climate could have other catchment scale impacts that are "*currently and commonly overlooked*" (Lane et al., 2008:228). These may be alterations in behaviour of landsliding (e.g. Dietrich et al., 1995), hillslope behaviour, sediment discharge regime (e.g. Hooke, 2003a) and aggradation of sediment which ultimately enhances flood risk (e.g. Stover and Montgomery, 2001; Korup et al., 2004; Pinter and Heine, 2005). With this in mind, it becomes increasingly crucial to understand the role that long term climate and extreme events exert on catchment behaviour, geomorphology change, hydrology change and sediment regimes at catchment scale.

2.2 Landscape sensitivity to changes in land use

It has long been recognised that vegetation within a catchment plays an intrinsic role in moderating the hydrological system. The effects are difficult to examine directly as different land use types and different tree species have different levels of control over hydrology. In the early 20th Century, the first paired-catchment (i.e. comparing catchments of similar size and characteristics but with different land use) experiments took place. Hibbert (1967) built on this work and undertook a more comprehensive evaluation of 39 paired catchments and found that (1) water yield increases as landscapes become more open, (2) afforestation reduces water yield, (3) response to changes in land use are highly variable and unpredictable. From this Bosch and Hewlett (1982) updated these observations to include more results from experimental catchments and also long-term observations. They found that different tree types alter water yield e.g. by reducing coniferous and eucalyptus trees by 10% in a catchment, annual water yield increases by 50mm, similarly 25mm reduction occurs in deciduous hard woods and 10mm change in water yield occurs in a 10% grassland reduction (Bosch & Hewlett, 1982). It has also been suggested that at least 20% change in land use is needed before any change can be identified in water yield. However, Brown et al., (2005) contest this and suggest that these results may be due to the small size of experimental catchments and the length of the study undertaken by Bosch and Hewlett (1982). Increased water yield may increase capacity of a river to transport sediment, therefore altering the fluvial regime.

Land use changes may be natural or indicate anthropogenic impact on the catchment. For example, climate can alter vegetation composition (Rosenmeier, 2002) or human settlement could promote deforestation or alter the land use from grazing to intensive agriculture or vice versa. By reducing forest cover there is a reduction in transpiration and soil moisture storage, thereby increasing water yield (Hibbert, 1967, Bosch & Hewlett, 1982, Bruijnzeel, 1990, Hornbeck and Swank, 1992, Stendnick, 1996).

In addition to the control that land use exerts on hydrology, sediment regimes may also change under different land uses. However the two primary controls on soil erosion and sediment delivery are changes in climate and land use (Verstraeten,

2009) and it is extremely difficult to disentangle the relative role that climate and land use play on both sediment delivery and soil erosion due to the similar nature of responses seen in the landscape (see 2.3 and 2.4 for further discussion).

Land use can alter sediment regimes in two ways, firstly lower forest cover can increase the susceptibility of the landscape to erosion (Macklin and Lewin, 2003; Chiverrell et al., 2007) due to lower levels of root binding, increased rilling and less evapotranspiration. In addition to this, there may be increased sediment supply under low forest cover as water yield increases (Hibbert, 1967) thus augmenting the rate of slope erosion processes such as creep, soil erosion and slope failure. It is well documented throughout Europe that human impacts have likely played an important role in the sediment regime. For example Starkel et al., (1991) identified accelerated sedimentation in central Europe during Roman and Medieval times that caused sediment overload, increased braiding in rivers and thus increased flooding.

Present day land use in the Western Alps is dominated by dairy cattle farming (Macdonald et al., 2005) the scale of which has been greatly reduced when compared to previous centuries. More recently, there has been a high level of abandonment of agricultural and pasture land in these areas due to remoteness of location, steep, harsh conditions, increasing age of farmers and finally a move away from traditional 'land-based' jobs to work in urban environments (Walther, 1986; Campagne et al., 1990). Furthermore, changes in shepherding of cattle has led to overgrazing and increased soil erosion at lower altitudes due to localised concentration of cattle close to mid-altitude alpine huts rather than in the high alpine pasture (Macdonald et al., 2000). Abandonment may have long term positive effects with regards to re-growth of vegetation, as this may reduce impacts associated with deforestation as slopes become increasingly stable and the chance of flooding is reduced. However, short-term impacts are thought to increase the risk of natural hazard such as landslides and avalanches (Cernusca et al., 1996) due to a build up of biomass through neglect of mown or grazed alpine pasture through abandonment. The study by Macdonald et al. (2000) shows that negative impacts of abandonment are consistent across the Western Alps whereas other European regions demonstrated a more variable pattern. Current European Common Agricultural Policy (CAP) supports farmers to maintain an open landscape to mitigate against the negative impacts of abandonment (Macdonald et al., 2000). If these kinds of policy continue into the future and are coupled with projected increasing precipitation, there may be serious implications for the hydrological, fluvial and sediment regimes at catchment scale, which may ultimately lead to increased frequency and magnitude of flooding.

2.3 Previous approaches

A key question of Holocene geomorphology is the extent to which human impact and climate drive sudden changes in geomorphic regime (Chiverrell et al., 2007). Changes in the flood regime of a catchment are important on a local level, as increased flooding in certain parts of the catchment could cause damage to property, destroy agricultural land and have a negative effect on local societies and economies. Increased flood frequency and flood magnitude (through changes in climate or land use change) would augment the effects of the flood regime in the catchment and cause increased problems for the local inhabitants.

With increased flooding comes a change in the sediment regime as more sediment is purported to be removed from the hillslopes under larger floods. This can cause dams and bridges to silt up, thus leaving no place for increases in river discharge, leading to overbank deposition and flooding outside of the channel thus disrupting the connectivity of the fluvial system (Fryirs et al., 2007) and the sediment conveyer (Lang et al., 2003).

Early research into the system drivers attempted to identify a single driver of the system, yet this idea has evolved to look more realistically at the combinations and resulting interactions of climate and land use, and not to unsuccessfully search for a sole driver of the system. The emphasis has begun to be placed on the human-environment interactions. This is supported by complexity theory in non-linear systems that suggests that component parts of systems (e.g. human and climate or *fluvial system and hillslope system) should be viewed holistically* as a co-evolutionary and adaptive process (cf. Holling, 2001; Lenton et al., 2004). Messerli et al., (2000) suggests that this holistic approach should be taken, but better understanding of the interactions should be made within a 'human-dominated' system rather than a nature dominated system. Currently the complex relationships between climate, environment and humans are poorly understood (Dearing, 2005) and this is an area of research which requires a closer focus in order to provide answers for a sustainable future.

There have been several attempts to ascertain the relative contributions of the external drivers of the system; longer term climate, land use and storm events via various approaches (Harvey et al., 1981, Macklin, 1999, Dearing & Jones, 2003, Foster et al., 2003,Arnaud et al., 2005, Macklin et al., 2005 Chiverrell et al., 2007) yet it is difficult to attribute the effects the drivers have on the system as increases in lake sedimentation accumulation and alluvial sequences could be attributed to either increased flooding from storm events, woodland clearance, or from longer term climatic wet shifts.

Early studies showed how changes in climate correlated well with changes in sedimentation and erosion (Macklin et.al, 1992, Passmore, 1994). There was also good correspondence between alluviation and erosion periods during wetter/cooler phases evident in raised bogs (Barber et. al, 1994). However other studies (Chiverrell, et al., 2007; Dearing, et al., 2001) challenged these ideas and suggested that human impact may be a more dominant driver in controlling catchment-scale responses. Chiverrell et al., (2007) demonstrate the lack of gullying in areas of the Solway Firth pre-human impact (2500 cal yr BP), and conclude that local changes to hillslopes at that time from anthropogenic input must have triggered a response to upland instability. Foster et al. (2003) suggest that lake sediments from the Petit Lac d'Annecy catchment tend to indicate that human impact has conditioned the landscape, and extreme events have mobilised the increased sediment supply, however they concludes that this may vary over temporal and spatial scales.

Understanding how the system behaves can provide information about its internal properties, the likely trajectory it may take, and it's predictability (Dearing and Zolitschka, 1999). The main focus of this research is to approach the research questions within a modelling framework to examine what extent climate and land use control fluvial and sediment systems as this is a fundamental line of thought for geomorphological research.

2.4 Lake sediment archives as an indicator of past change

Lake sediment archives have long been used to reconstruct past changes in erosion and sediment delivery (Mackereth, 1966; Engstrom and Wright, 1984; Dearing, 1991; Dearing and Jones, 2003; Chiverrell, 2006). As human impacts on the landscape have increased during the late Holocene, the sediments preserved within lacustrine basins typically comprise a record of the erosion of catchment soils, especially in upland areas (Mackereth, 1966). Alpine lake sediments are characteristically influenced by climatically induced floods through snowmelt or intense summer thunderstorms (Foster et al., 2003), active hillslope processes such as mass movement and gullying on steep slopes (Foster et al., 2003) and avalanches which deliver sediment from highest slopes to lower altitudes rapidly. Geochemical and environmental magnetic analyses can assist the reconstruction of sediment-source linkages and the changes in the geomorphological regime of the catchment (e.g. van der Post et al., 1997; Dearing et al., 2001, 2008; Foster et al., 2003, 2008; Shen et al., 2007; Chiverrell et al., 2008). The linkages between catchments and lakes are reflected in the erosion, transmission and depositional processes that produce the eventual lacustrine sedimentary record. Both coupling and connectivity between geomorphological processes in the catchment and sediment delivery to the lake govern the extent to which externally driven events such as storms (floods) land-use change, and longer-term climate changes are reflected in the sediment record.

Previous research has seen geomorphologists and palaeolimnologists attempting to ascertain the relative contributions of these external drivers (eg, Harvey et al., 1981; Macklin, 1999; Foster et al., 2003; Arnaud et al., 2005; Chiverrell et al., 2007), but they have been hampered by the integrated nature of the meteorologically driven flood events, longer-term climate changes and land-use signals preserved in sedimentary and geomorphological archives. In addition to drivers that are extrinsic to the system, other intrinsic factors moderate the process-response to landscape change, meteorological events and long-term climate change. Variable sediment distribution, erodibility of the landscape, local storage and release of sediment (e.g. floodplains), and longer-term sediment sinks (e.g. higher-level lakes and alluvial fans), can all interrupt the sediment conveyor (Fryirs and Brierley, 2001; Lang et al., 2003; Fryirs et al., 2007) with the potential to significantly moderate any palaeoenvironmental signal recorded within lake sediments (Dearing and Jones, 2003; Chiverrell et al., 2008). Individually the sedimentary and landform archives of these sub-systems are difficult to interpret and particularly difficult to correlate with changes in land use, meteorological events and/or long-term climate changes. Therefore, providing that drivers and process-responses can be incorporated and isolated, the catchment-scale modelling framework presented here offers a more holistic approach to better explore the response of sediment flux between geomorphological sub-systems.

2.5 The need for modelling

Catchment-scale studies have long been used to identify the impacts of climate changes and land use changes, but they have been traditionally empirically based studies. Long-established approaches to research have relied on empirical data observed in the field and actual instrumental data to consequently advance understanding and inform future data collection (Wainwright and Mulligan, 2004). However, available instrumental data is temporally limited as typically recorded data only extends a few hundred years into the past. Environmental sciences in the last 30 years have relied on proxy records of change from environmental archives such as lake sediments, ice cores and speleothems to extend the record of environmental change much further into the past. Longer temporal records are always desired in order to identify long term trends in the past which may be obscured in short term records. Proxy records of environmental change are not without their own problems of inaccurate sampling and poorly-defined chronologies, but they provide a comprehensive estimate of data which are not otherwise available for the past (Dearing, 2006) . Modelling environmental systems has been increasingly popular in recent years due to the urgent need for extrapolation of data for future prediction. Modelling allows various scales to be studied and can act as a virtual laboratory by using physically based models as it is sometimes difficult to recreate certain environmental processes in laboratories (Wainwright and Mulligan, 2004). Numerical models are essentially an excellent way to test our understanding. If a hypothesis is stated, and the model produces different results to that which we expect, our understanding of the process may be limited, which forces us to think in another way, alter parameters and re-test our hypotheses. Models are also an inexpensive way of testing processes and are at present the only way in which we can identify changes and future impacts (Wainwright & Mulligan, 2004). Observed/recorded data will always be closer to the truth and must remain the most important in science, but modelling offers the best opportunity for prediction of future change, where empirical data are not available. A combined approach, using data-model comparisons is likely to be the best way forward to understanding future change.

In an applied context, traditional methods of modelling sediment production, transport and deposition have concentrated on sediment budget approaches (Trimble, 1983) through field studies (Lang et al., 2003; Rommens et al., 2006; Hoffman et al., 2007). Whilst sediment budgets describe inputs, transfers, stores and outputs of sediment in a catchment, they cannot capture geomorphological change or interactions between catchment-scale processes. In fact, there have been few attempts to model interactions between hillslopes and fluvial systems, largely because of the non-linear behaviour and complex nature of the process interactions. This research addresses these issues by using and developing an established hydro-geomorphological model (CAESAR: Coulthard, 1999; Coulthard et al., 2002, 2005; Van de Wiel et al., 2007) to assess sediment transfer within a lake catchment. Previous validation of CAESAR model output has been attempted by comparison with well-dated fluvial sequences, yet these are fragmentary, discontinuous and spatially variable. A potential advantage of a lake catchment system is the availability of lacustrine sediments to provide an integrated record of sediment flux over long timescales against which model sediment outputs may be compared. Although lake sedimentary archives do not always reflect solely allogenic processes, by carefully selecting proxies that reflect detrital soil-sediment signals from the sedimentary archive, comparisons can be made with modelled CAESAR sediment discharge records.

Chapter 3

Complex Systems and Numerical Modelling

3.1 Introduction

There is a growing need for predictive numerical models that can realistically handle and simulate complex environments such as fluvial systems at a catchment scale. Numerical models are required so that we can firstly test our understanding of environmental systems in the anticipation that we may be able to progress to prediction of future changes in these systems and utilise numerical models as tools for more sustainable management of systems. The rationale for selecting CAESAR model for his research is underpinned by theories surrounding complexity. This chapter describes the characteristics of a complex system, provides a background of the evolution of models that can handle complexity and describes the nature and mechanics of CAESAR model.

3.2 Modelling complexity in environmental systems

3.2.1 Introduction to complex systems

Complex systems are often defined by their characteristics rather than expressed as a formal definition largely because *"complexity is an oft-used word that lacks precise definition"* (Ben-Jacob and Levine, 2001:986). A dictionary definition of 'complex' states that systems *"consist(s) of interconnected/interwoven parts"* with a firm emphasis placed on the interactions between the component parts of systems (Bar Yam, 1997). Complexity describes system behaviour rather than system structure Emergent behaviour occurs within systems when component interactions behave in non-apparent ways (Coulthard and Van de Wiel, 2007) and cannot be predicted from the simple rules describing the interactions, but instead emerges from the interactions themselves (Phillips, 1999; Harrison, 2001; de Boer, 2001; Hooke, 2007, Dearing, 2008). An example of emergent behaviour is that of sediment dynamics (de Boer, 2001), or the development of a simulated landform through repetition of simple physical based rules. For example, an alluvial fan can emerge from interaction between erosion and deposition when a tributary joins the main valley. However, the processes of erosion and deposition alone cannot suggest that a fan will occur as the fan emerges from the interactions of the processes in the prior landscape (Coulthard and Van de Wiel, 2007). It is a common misconception that complex systems are themselves made up of component parts that are themselves complex. This is very rarely true and in fact complex emergent behaviours can evolve from simple component parts and rules.

Essentially it is useful to know the typical characteristics of complex systems in order to identify a tool with which to model them. Environmental systems are saturated with complex responses and are typically characterised by both external forcing to the system (e.g. solar activity and climate) and also by intrinsic system behaviour such as feedback mechanisms and system thresholds (Chorley and Kennedy, 1971), self-organisation (the development of ordered patterns expressed in the functional and morphological aspects of the system (de Boer, 2001)) and non-linear responses. Non-linear systems can be defined as systems where the outputs are disproportional to the inputs (Schumm, 1979; Bak, 1996; Phillips, 2003). There are many examples of non-linear behaviour in fluvial systems including meander migration (Hooke, 2003b), bedload pulses (Gomez and Phillips, 1999) and the overall response of drainage basins (Schumm, 2005).

Complex systems generally contain a large number of mutually interacting parts (Bar-Yam, 1997) which can lead to major differences in the overall fundamental behaviour of environmental systems over time (Dearing and Zolitschka, 1999). In order to model complex systems in a realistic manner, we require a tool that can capture all of these system behaviours.

3.2.2 Reductionist or Holistic approach?

Complexity theory suggests that studying component part *interactions* of systems is vital if we are to gain a better insight into non-linear, self-organising systems and it is therefore useful for geomorphologists and hydrologists to study process interaction holistically rather than studying individual component parts. Hawking (1988) believes that the environmental sciences are "*strongly rooted*" in "*reductionist science*" (Dearing 2005) whereby a system is broken down into its component parts and studied individually. Wainwright & Mulligan (2004) also note the reductionist approach to the study of complex environmental systems and identify areas where fluvial system components e.g. hillslope-floodplain, river-floodplain, hillslope-river may be considered as a pair, but rarely is the whole catchment system considered as one. Hawking (1988) believes that "*reductionism poses a risk for full understanding*" and that vital evidence about how the system works will be obscured, an idea that is applicable when studying complex systems and thus the "*strong roots*" in reductionism should be moved away from (but not ignored) in order to advance science and study complex systems efficiently. An ideal methodology for studying complex systems would be to firstly identify the component parts, secondly identify the nature of component parts (i.e. complex or simple), and finally and most importantly, study the interactions and feedbacks between the component parts and the resultant emergent behaviour. Reductionist approaches cannot gain a full understanding of behaviour, feedback and complex response, therefore to model complex systems, a holistic approach is preferable.

Hydrological scientists have long studied component parts of systems at a range of scales which have been fundamental to our understanding of how the individual structures work (e.g. hillslope hydrology, behaviour of rivers). However, in the last ~10 years, holistic approaches to research have become increasingly common (e.g. catchment scale rather than hillslope or reach scale research), which allows a much less restricted view of process interactions and overall system behaviour. By understanding overall system behaviours one can attempt to identify a system's likely trajectory of change, its predictability and the ways in which it behaves with other systems (Dearing and Zolitschka, 1999).

There is therefore an evident need for simulation models that allow complex and macro scale emergent phenomena to arise from micro-scale interactions "*with as few constraints as possible on spatial and temporal scales*" (Dearing 2005). Advances in pioneering numerical computer simulation models over the last 15 years (e.g. Murray and Paola, 1994; Tucker and Slingerland, 1994; Coulthard et al. 2002, 2005) can capture non-linearity and emergent behaviour and are constantly evolving to allow a better understanding of holistic system behaviour. Coulthard (1999) and Dearing (2005) believe that cellular automata (CA) models appear to satisfy many of the requirements necessary to study complex systems. Favis-Mortlock (2004) describes cellular automata and cellular models as the main tool used for modelling self-organising systems.

This research makes use of a reduced complexity cellular automata numerical model (CAESAR) written by Coulthard (1999) and developed by several other workers (e.g

Van de Wiel et al., 2007, Cox et al., 2005). Further details about model mechanics are given in section 3.3., however considering theoretical concepts of traditional cellular automata (CA) models and the suitability of a cellular approach that is favoured for this research are firstly considered.

3.2.3 Traditional Cellular Automata Models

CA models are widely used in many disciplines (e.g. biological, mathematical) and are becoming increasingly used in environmental sciences. CA models make use of a regular or irregular mesh of grid cells to represent a surface, and low-level process

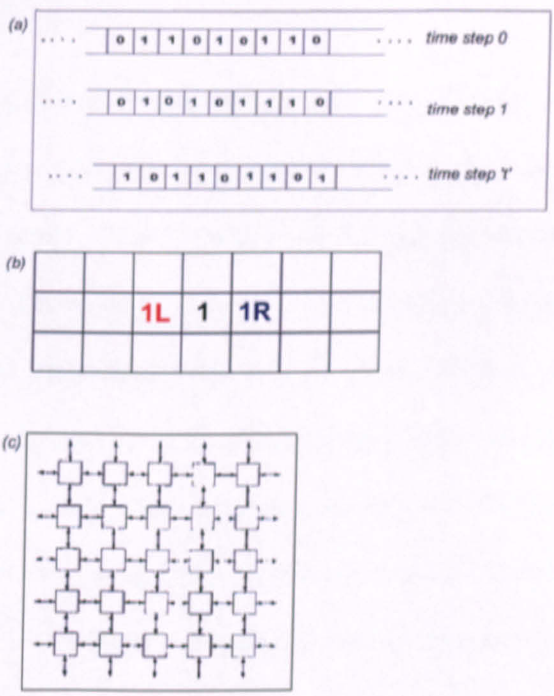

Figure 3.1 *(a) Evolution of a 1-D CA model (after Sarkar, 2000), (b) the value for cell '1' is determined by the value in cell '1L', cell '1R' and cell '1' itself. In 1-D CA models, all cells interact this way. They are not dependant on any other cell in the grid, (c) a 2-D CA model displaying the orthogonal neighbourhood (Sarkar, 2000)*

based rules or rules are applied to each cell. Wolfram (1984) identifies five key characteristics of cellular automata:

1. They consist of a discrete lattice of cells

2. They evolve in discrete time steps

3. Each cell takes on a finite set of possible states (Fig.3.1a)

4. The state of each cell evolves according to the same deterministic rules

5. The rules for cell evolution depend only on interactions with immediately neighbouring cells.

A 1-dimensional lattice is the simplest state a CA can exist in (Fig.3.1a). The 'time step' (or iteration) is essentially referring to continuous time which is discretised into cells, this is necessary if the model is to be computationally trackable (Favis-Mortlock, 2004). The value in each cell changes in every timestep according to the process rules and rules applied to the cell (Wolfram, 1984). In a 1-D CA model, the state of the cell is determined by the present state of itself, and the state of its left and right neighbours (Fig.3.1b). Von Neumann developed 2-D CA models in the 1960s, which proposed to "*bring rigor to complicated natural systems*" (Von Neumann, 1963) and took 1-D CA models to the next level of complexity by connecting each cell to its 4 orthogonal neighbours which is termed "cellular space" (Fig.3.1c).

3.2.4 Building on traditional CA models

Whilst strict cellular automata may suit some environmental studies, some researchers have moved forward from the fundamental concepts of traditional CA. One example of this is Cellular Automata Evolutionary Slope and River (CAESAR) model (Coulthard, 1999; Coulthard et al., 2002, 2005, 2007; Van de Wiel et al., 2007) which steps away from orthogonal neighbourhood rules between cells (von Neumann, 1963) and allows the cell to interact with a specified number of cells both orthogonal and diagonal (Van de Wiel et al., 2007). This moves beyond the rules of traditional CA by modifying the fundamental characteristics to incorporate more cells into the model calculation allowing more interaction between a greater numbers of cells. The value of water depth held within each cell within CAESAR is determined by the contributing cells upslope of any cell (section 3.3.1.3) and is therefore not determined solely by the next neighbour or orthogonal neighbour (Fig.3.1c) as per traditional cellular automata rules.

Earlier work termed this type of model a 'cellular model' rather than 'cellular automata' and there has been some discussion as to which terminology should be used. Favis-Mortlock, (2004) stated that terminology should be of little concern as long as the modifications to the complexity do not invalidate the assumption that the models can simulate non-linear emergence. Cellular automata /cellular model is referred to here interchangeably to describe models with fundamental roots in CA.

A fundamental commonality between complex environmental systems and CA models are the self-organising phenomena exhibited in each (Sarkar, 2000). Thus CA models appear to be well suited to simulating complex environments If CA self-organises, it gives rise to a larger scale response (e.g. larger than the small-scale interactions) that manifests as patterns on the computer grid cells (Wolfram, 2002; Wooton, 2001).

Cellular automata have proved a popular mechanism to attempt to model the non-linear behaviour exhibited in various environmental systems e.g. beach cusps

(Werner and Fink, 1993), stone stripes (Werner and Hallet,1993), braided river channels (Murray and Paola,1994) and sediment dynamics in fluvial systems (De Boer, 2001). These models have been used extensively particularly in fluvial geomorphology to study braided river systems (e.g. Murray and Paola; 1994, 2003; Thomas and Nicholas, 2002), bank erosion (e.g. Fonstad and Marcus, 2003), river meandering (e.g. Coulthard and Van De Wiel, 2006, 2007), sediment yield (e.g. De Boer, 2001), alluvial fan development (e.g. Coulthard et al., 2002; Nicholas and Quine, in press), and fluvially-driven catchment-scale landscape evolution (e.g. Coulthard et al., 1998; Luo, 2001; Crave and Davy, 2001; Coulthard and Macklin, 2003; Van de Wiel et al., 2007). Cellular automata have the ability to capture non-linearity, feedbacks and emergent behaviour (Dearing, 2005) thereby enabling complex responses to emerge from simple low-level rules. This provides a useful framework for studying complexity in the environment and for simulating catchment scale landscape evolution.

3.2.5 Landscape evolution models

Cellular automata principals have inspired the development of many of the models that are in operation across the research community under the bracket of 'landscape evolution models'. During the 1970's, computer models were first used to represent landscapes and network channels (Anhert 1976) which simulated the environment by routing water across a mesh and changing the elevation to represent fluvial and slope processes. As computer processing capacity developed, so did the models and the process representation within them (Tucker and Slingerland 1994; Braun and Sambridge 1997). There are various types of models, all based on different environmental interactions and processes. Different models are suitable for different purposes e.g. long-term tectonic modelling, short term fluvial processes and selection is dependent on the nature of the research.

A good review of landscape evolution models and their unsuitability for this type of research (largely due to temporal scale or process representation) is outlined in Coulthard (1999), Peeters (unpublished thesis, 2007) and Codilean et al. (2008). They summarise that the choice of model depends largely on the time scale, spatial scale and nature of the system that one wishes to model. There are many landscape evolution models in the academic community and Coulthard (2001) wrote a brief review of four landscape evolution models (other than CAESAR, Coulthard, 1999) which simulate landscapes, are grid based and can incorporate complexity, emergent behaviour and non-linearity. The models included in the review are SIBERIA (Evans and Wilgoose, 2000 & Wilgoose, 2005), GOLEM (Tucker and Slingerland 1994), CASCADE (Braun and Sambridge 1997), CHILD (Tucker 2000). For this research, none of these available models are suitable as they do not incorporate fine resolution flood-level modelling (e.g. hourly floods). Additionally, the length of time that the models operate at (10^5-10^6) and the coarse time-step at which they operate is often too long to capture the finer resolution (e.g. GOLEM, Tucker and Slingerland, 1994). Others operate over coarse spatial scales such as 1km (e.g. CASCADE, Braun and Sambridge 1997), again losing the finer detail required for local-scale modelling e.g. soil creep and fluvial sediment movement thus compromises between simulation length and simulation resolution must be made (e.g. Van de Wiel et al. 2007). Some models have better process representation than CAESAR e.g. SIBERIA and GOLEM contain better representations of slope processes including rock weathering than some of the other models and reinforces the idea that the selection of model is ultimately dependant on the temporal and spatial scales at which one wishes to model catchment processes.

The difficulty with generic landscape evolution modelling becomes apparent when looking at more complicated environments such as large river catchments. Influences on the catchment and drainage network include interactions between hydrology, fluvial erosion, slope processes, tectonic uplift, climate and lithology (Coulthard 2001). Large number of processes operating over a wide range of spatial and temporal scales renders realistic landscape evolution modelling difficult if trying to incorporate all of the processes acting at catchment scale over various spatial and temporal scales (Coulthard 2001).

3.2.6 Reduced Complexity Models

The term 'reduced complexity model' (RCM) has emerged in recent years, and refers to models that reduce the level of complexity of physics embedded in algorithms thereby simplifying the process rules, without compromising the ability of the model to simulate complex responses. Here the term 'reduced complexity model' refers to the developments within the environmental community, most of which have their roots in cellular automata. Different modellers (e.g. Werner, 1999; Paola, 2001; Murray, 2003; Nicholas, 2005) have different visions of the characteristics and principles of a RCM, and the classification appears largely dependent on the objectives of the modeller. One commonality with all of the models classified as 'reduced-complexity' is that they provide an alternative to the more traditional reductionist approach. Infrequently a model can be completely categorised into 'reductionist' or 'synthesist' (Paola, 2003) as they usually span a compromise between the two, which questions how useful the terminology of RCM is to a modeller. Nicholas et al. (2007) suggests that the term 'reduced complexity' is perhaps unnecessary as all models represent simplifications of reality. However, as many environmental systems have such high levels of complexity, at what point does a modeller decide that the level of complexity within the model is high enough to simulate an environment realistically? Murray (2007) suggests a 'best' level of complexity is assumed that will be *"simultaneously optimal for explaining and predicting the behaviors or patterns in question"*.

The type of models that are classified as 'reduced-complexity' in fluvial environments have a tendency to be those that are built on the fundamentals of original landscape evolution models (e.g. Kirkby, 1971 and Anhert, 1976), more complex landscape evolution models (e.g. Tucker and Slingerland, 1994) and more traditional cellular automata models (e.g. Murray and Paola, 1994), many of which may also be termed within the RCM bracket. Nicholas et al. (2007) suggests that some of the fluvial models of stream braiding (Murray and Paola, 1994; Thomas et al., 2002), meander migration (Sun et al., 1996; Lancaster and Bras, 2002), alluvial fan formation (Coulthard et al., 2002; Sun et al.,2002), rill development (Favis-Mortlock et al., 2000), and landscape evolution (Howard, 1994; Tucker and Slingerland, 1997; Coulthard et al., 1998) may all be termed reduced complexity.

Nicholas et al. (2007) suggest that '*reduced-complexity models have the potential to elucidate river behaviour over historic and Holocene timescales that are beyond the scope of alternative fluvial modelling approaches*'. Reduced complexity modelling essentially breaches the gap between large scale, long term landscape evolution models that lack the fine resolution necessary to study interactions between fluvial and hillslope systems at catchment scale and the highly detailed, yet time consuming computational fluid dynamic models (CFD) (Coulthard et al., 2007). The do however offer an alternative method to investigating hydraulics and fluvial sediment transport than laboratory flume experiments (Brasington and Richards, 2007). By reducing the level of complexity of many parameters to an acceptable level of process rules, emergent behaviour can still be demonstrated through cell interactions, thus raising the question of "what level of detail of process representation is required within numerical models?" Model complexity should also not be reduced too far so that the model has no correspondence with behaviours in reality (Murray, 2007).

Fundamentally, as with all modelling, the answer to the level of complexity required relies on the questions asked and also the spatial and temporal scale at which research is carried out. Murray (2007) outlines that development of geomorphic models lies partly by desire to answer questions that have a societal impact (e.g. impacts of climate change) but explanation of process and responses is seen as the goal of many geomorphological modellers (e.g. Church,1996; Kirkby, 1996; Richards, 1996). Geomorphology is often less concerned with exact timing of events, and more focused on the reasons why change takes place, processes occur and landforms changes. RCMs can therefore allow fundamental rules of physical processes to be reduced to an acceptably low level, and repeated many times to identify emergent behaviour from these basic level cell interactions.

Models can only simulate reality and all models have a degree of uncertainty and limitations associated with them. In short, all models are compromised representations of reality (Coulthard et al., 2000) e.g. CAESAR sacrifices the accuracy of water depth and flow in order to study the effects of erosion and deposition in more detail. Theoretically, the only limitation any model should have is the processing capacity (Coulthard et al., 2000) and it is computational power which is ultimately the greatest limitation to this type of research. Whilst this will inevitably improve over time through large scale GRID-based parallel processing systems linked to one large computer, Coulthard et al. (2007) argue that reduced complexity models may be superseded by the more detailed CFD models if simulation times of these models improve. Despite the complexity of environmental systems, simple inputs to models and constant refining of process representation can help to aid understanding, and test our knowledge.

3.3 Origins of CAESAR

The aims of this research are to test, validate and develop an extension of an existing cellular automata numerical model 'CAESAR' (Cellular Automata Evolutionary Slope and River) originally developed to model geomorphological development and coarse-grained sediment transport in upland river catchments in the British Isles (Coulthard, 1999). The overriding aim of this work was to investigate the response of the fluvial system to Holocene climate and land-use and simulate the impacts of flood events (Coulthard, 1999). Since then CAESAR has been developed and refined by Coulthard and a number of other developers (Coulthard et al, 2001; Coulthard et al. 2002, Coulthard et al, 2005, Cox et al., 2005, Coulthard et al. 2006; Coulthard and Van de Wiel, 2006; Van de Wiel et al., 2007, Gez Foster, Joe Wheaton) (see 3.3.1.5). The developments made to CAESAR are largely dependent upon each developer's research agenda. CAEAR has hybridised into a modular style of software to suit a multiplicity of purposes.

3.3.1 Model Structure

CAESAR uses a digital elevation model (DEM) to represent landscape surfaces such as drainage basins and fluvial reaches. The DEMs are comprised of a regularsized mesh of grid cells that holds spatially distributed information about initial conditions of a simulation, including elevation, water depth, roughness, grain size and vegetation cover. The resolution of the DEM can vary from centimetres to hundreds of meters, depending upon the user requirements, and this is often constrained by simulation time. For example, a reach-scale simulation with 1 year of input data and a fine resolution DEM may run quickly (minutes – less than 1 day) whereas a simulation of 2000 years with a DEM of 50m resolution may take several months (real-time) to complete. Therefore, by increasing the resolution of the cell size in a DEM, the number of cells is increased, which in turn augments the number of calculations to be made per time step, thereby exponentially increasing the simulation time.

CAESAR is based on the cellular automaton concept and applies low level process-based rules to each cell. The iterative and repeated nature of the rules allows complex behaviour to be produced and is thus ideal for simulating non-linear environments where positive and negative feedbacks and emergence can be captured. The process rules fall into the four groups: hydrological routing, hydraulic routing, slope process rules (mass movement, soil creep and soil erosion) and fluvial erosion and deposition (Fig.3.2).

Each process rule is translated into a simple numerical equivalent based on the equations that compose the rules. The low-level rules are applied to each cell and the values are re-calculated for each time step or 'iteration' within the run time of the model. The calculations per time step are not solely based on the equations in one single cell, but are also based on the value in the squares adjacent to cell *n*. Traditional cellular automata rely solely on the adjacent neighbouring cells however CAESAR is not strictly a cellular automata model but a modification of a cellular automata model due to the unique flow routing scanning algorithm that is built into it.

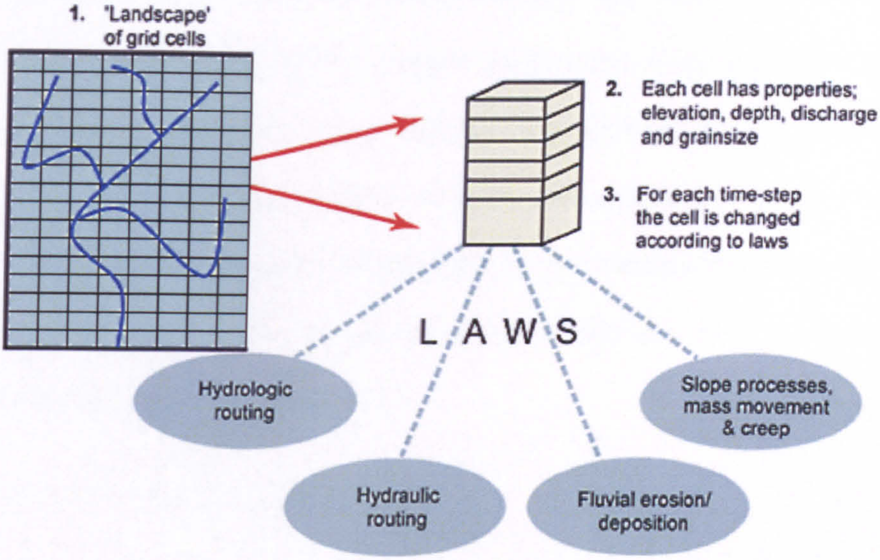

Figure 3.2 *Schematic diagram of the process laws operating in CAESAR (Coulthard, 1999)*

3.3.2 Spatially distributed data within CAESAR

Many of the parameters within CAESAR are spatially distributed including elevation (to represent topography), vegetation growth, grain size distribution, water depth, erosion and deposition, stratigraphy and sediment depth. Each cell within the model holds these data and uses the information to determine initial conditions. Precipitation and land use are the two drivers of CAESAR, both of which are not spatially distributed and are applied uniformly each iteration in every cell across the catchment. Fundamentally, this is a trade-off between model complexity and model run-time efficiency.

Temporal changes of land use within CAESAR are handled as a component of the overland flow routine within the hydrological model (see section 5.5). For the scope and purpose of this thesis, developing a spatially variable hydrological model is not possible as it would require a major modification of the implementation of the hydrological model, which is at the heart of CAESAR. This is a consideration for future development. This inability to handle land use and precipitation on a spatial basis will continue because of difficulties in implementing this spatial variability in the hydrological routing component of the model (i.e. cells downstream are a function of the wet cells upstream).

3.3.3 Flow routing scanning algorithm

The precipitation data (required to run CAESAR in catchment mode) are applied uniformly across the mesh of grid cells which ultimately drives the model. The water hits a cell and hydrologic and hydraulic routing rules are applied to the cell based on the water depth of each cell supplemented by input from the precipitation file. Water depth of each cell is held in a grid and used for the next iteration. J_mean is the parameter used to determine cell saturation/water and further details about this parameter can be found in section 5.5. Conventional DEM based hydrological models are forced to 'sort' the grid based landscape altitudinally which is a computationally expensive task (Coulthard, 1999) in terms of time and computer processing capacity. This is not such a problem in non-spatial hydrological models as altitudes need to be sorted only once during a model run. However as CAESAR erodes and deposits, the path of water from highest to lowest constantly changes (Coulthard, 1999). The model therefore needs to re-sort the altitudes following each iteration, which takes an unacceptably long time for conventional model algorithms. To address this Coulthard (1999) and Van de Wiel et al. (2007) developed a scanning algorithm, which enables model processing time to be spent more efficiently on areas of greatest activity e.g. the channel. Hillslopes are periodically checked (daily in model time) as less activity takes place here over smaller timesteps (e.g. minutes or hours) in the original version of CAESAR (version 5.9 and earlier). This version of the model is used in chapter 6 of this thesis, however, the representation of hillslope processes has been developed and is detailed fully in chapter 7.

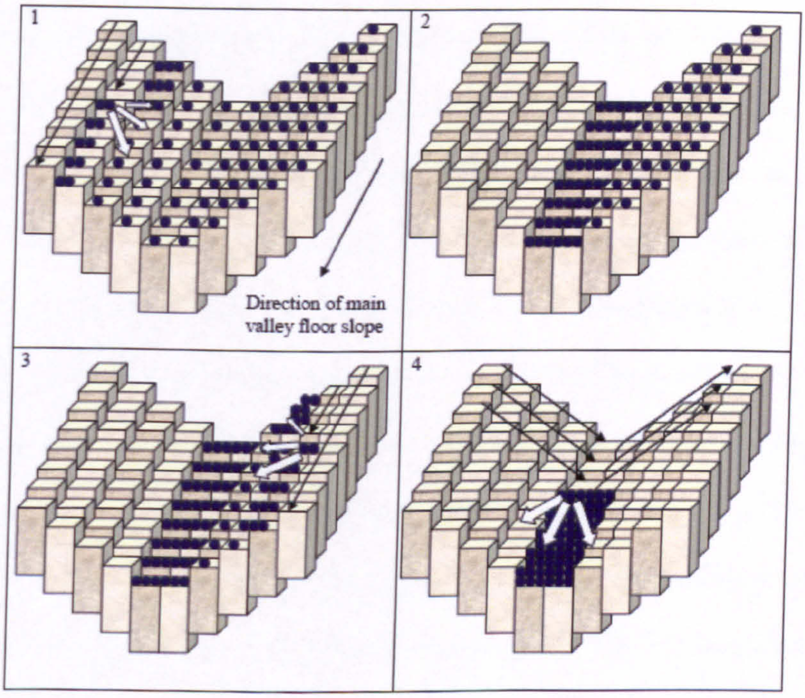

Figure 3.3 *Schematic diagram of flow routing algorithm within CAESAR (after Coulthard et. al 2002)*

The scanning algorithm is a modification (Fig.3.3) of the flow routing method used by Murray and Paola (1994) to simulate braided rivers, but allows calculation of flow in all directions (Coulthard et al., 2002, Van de Wiel et al., 2007). The initial scan moves from left to right across the grid representing the catchment (box 1); the routing pushes the water from each cell into the three lower cells to the right. When the base of the valley is reached, the scan continues upslope from left to right, but no flow is routed here (box 2). The second scan repeats the process but moves in the opposite direction from right to left (box 3), and consequently leaves all the water from the catchment in the base of the valley. The water is then routed and pushed out of the catchment during scan three (box 4) which scans from the top to the bottom. The process is repeated during scan four, from bottom to top (not shown here), the whole sequence is then repeated resulting in 8 scans, 2 in each direction which can thus accommodate high-resolution grids (Van de Wiel et al., 2007). During each scan, the maximum discharge is calculated and recorded for each cell, and taken as the discharge for that particular cell. Cells downstream are a function of upstream 'wet' cells. If the 11 downstream cells are higher than the contributing cell but the combined water depth and elevation of the contributing cell is higher, water is retained in the contributing cell up to the depth of the obstruction whilst the rest is routed on. Any water trapped in the topography allows this scanning method to simulate flow around channel geometries and complex bends. Coulthard et al., (2002) describes the use of this scanning routine in the modelling of meanders within a channel and explains how water may be routed around the first bend, but trapped on the second during the first iteration. During the second iteration, the trapped water is still within the cell but can be released in the next scan, and is replaced by water from upstream, which allows a continuous flow of water. This is similar to multiple-flow methods used by Moglen and Bras (1995) but is much quicker and can be applied to larger grids in excess of a million cells. Further details can be found in (Coulthard and Van de Wiel, 2006).

3.3.4 Process representation in CAESAR

The large number of processes that occur at catchment scale are represented efficiently within one model and can be applied generically to any catchment. Processes such as soil creep will be less important on fine resolution temporal scales (e.g. annual), however they could have large implications over the long-term evolution of the catchment (e.g. hundreds of years). Conversely, processes such as braiding of rivers require a higher level of detail and require many changes on very short time scales which are important to incorporate as changes make take place quickly over a few hours. Due to the scanning routine outlined in section 3.3.3, the model is able to concentrate most of its computational processing time on areas of activity within the catchment such as the channel, and it is programmed to periodically check other areas such as the hillslopes e.g. daily (model time). This also allows CAESAR to operate at sub-second time steps when geomorphic activity is at a maximum, and at a coarser resolution (e.g. hourly) when the system is relatively stable. Consequently, the run time of each simulation has been made more efficient by this scanning algorithm. Initially the model routes a water network through the catchment and identifies which cells have water in them, and subsequently creates a buffer zone around the identified 'wet' cells. For every iteration, fluvial erosion and deposition and local slope processes are calculated, and the whole catchment is then re-scanned every 100 iterations. If there is a large flood, and the discharge exceeds 50% of the set discharge used to calculate the buffer zone, the discharge is doubled and the buffer zone is recalculated (Coulthard, et al., 2002). This allows the dynamic expansion or contraction of channels during changes in environmental conditions within a catchment. During channel avulsions, the model automatically re-scans the mesh of grid cells, and determines a new channel area and buffer zone. Similarly if the number of cells with water in them is reduced, the CAESAR re-scans the grid. This ensures that the most active area of the catchment is focused on at all times. It also allows computational optimisation of time and allows the model to be more efficient e.g. for 98 or 99% of the computer's operating time, less than 10% of the cells are checked at any one time, yet they are periodically checked throughout the model run (Coulthard 1999). The scanning algorithm and computational processing optimisation of the time spent calculating cell values is at the heart of CAESAR and distinguishes it from other landscape evolution models by allowing long-term high resolution simulations on large grids in relatively short run-times.

3.3.5 Previous applications of CAESAR

Previous applications of CAESAR have included Holocene landscape evolution, long term evolution of alluvial fans (Coulthard et al. 2002), investigating the effect of climate and land use on the geomorphic development of fluvial systems (Coulthard et al., 2000) and investigating differential catchment response to climate and land use change (Coulthard et al., 2005). TRACER (a variation of CAESAR, Coulthard) has been used to predict the extent and dispersion of contamination of heavy metal pollution in a floodplain reach (Coulthard and Macklin, 2003). More recently, the flow routing algorithm within CAESAR has been refined (Van de Wiel et al., 2007 and Coulthard et al., 2007) to allow more realistic simulation of meanders in a cellular framework. At present, CAESAR is the only fluvial model to allow braiding and meandering processes to be incorporated within a single model. Previous validation of CAESAR model output has been attempted by comparison with well-dated fluvial sequences, yet these are fragmentary, discontinuous and spatially variable.

This research uses proxy records from lake sediment archives from the pre-Alpine zone to rigorously test and compare modelled sediment data to. In addition short-term water discharge data will be compared to instrumental discharge data, and modelled erosion and deposition will be compared to catchment geomorphology to give confidence that CAESAR can be considered a robust tool for modelling future behaviours in light of changing climate and land use.

Chapter 4

Study Site: Petit lac d'Annecy, Haute Savoie, France

4.1 Introduction

Mountain regions are particularly sensitive to changes in system drivers (Beniston, 2005) e.g. climate and land use as the impacts of each are amplified in these high gradient and high altitude zones. Steeper slopes are subject to accelerated erosion rates and are sensitive to changes in precipitation such as extreme rainfall events which are more common in high altitude environments. These areas are also sensitive to changes in land use and although the present day land use in European mountain regions on hillslopes is low intensity hill pasture and forest, this has not always been the case in the past (MacDonald et al, 2000). This chapter outlines the regional setting for this study and describes the physical characteristics of the Petit lac d'Annecy catchment and sub-catchments.

4.2 European Alps

4.2.1 Key characteristics

The Petit lac d'Annecy (Fig.4.1) is situated in the pre-Alpine foreland of the French region of the European Alps which extend over 800km in length across France, Italy, Switzerland, Germany, Slovenia, Lichtenstein and Austria. The Alps have a mean width of 200km and a maximum height of 4808m at the summit of Mont Blanc (Fig.4.1b) (Steininger and Weck-Hannemann, 2002). The Alps were formed as a result of orogeny during the early Cenozoic era, beginning in the early Eocene ~55mya and slowing during the Miocene. The collision of the African and European tectonic plates (Schmid et al., 2004) compressed the western part of the Tethys

Ocean causing continental-scale uplift and creation of the Alpine mountain ranges. The present landscape has been shaped by a series of glaciations over the last 2.4 m years.

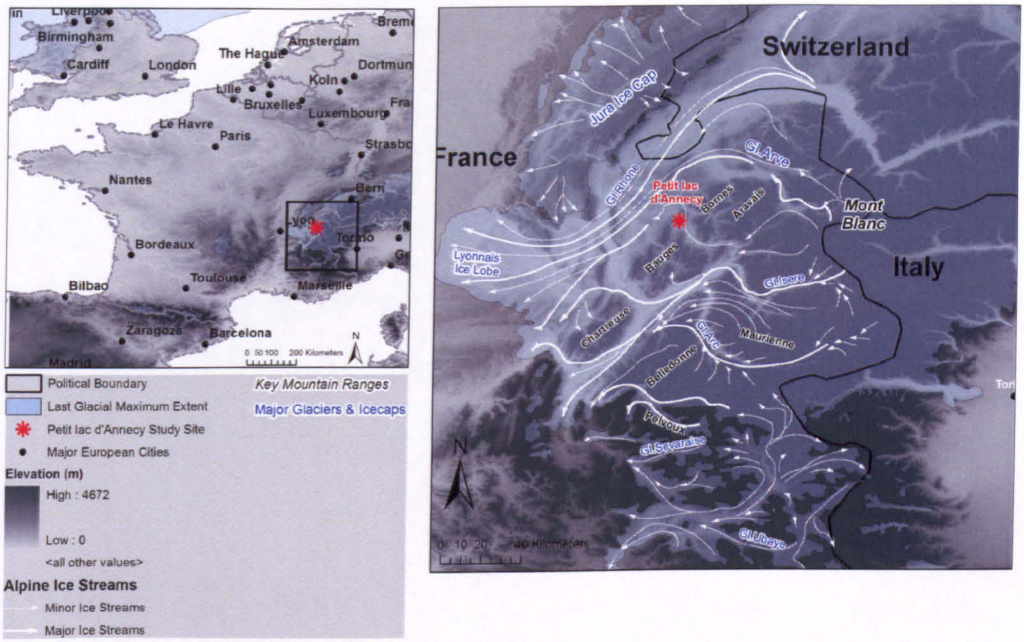

Figure 4.1*(a) Study site and surrounding countries. (b) Major ice flows around study site and surrounding areas during the last glacial maximum (LGM). Data source: Ehlers and Gibbard (2004)*

4.2.2 Typical Climate

The Alps comprise the boundary between two major climatic zones; the mid-latitude temperate and the Mediterranean climate zones (Steininger and Weck-Hannemann, 2002). Instrumental data have shown that weather extremes (e.g. precipitation, temperature, winds) are characteristic of Alpine environments, with weather systems modified by and in some cases triggered by the scale and geometry of the terrain (Cebon et al., 1998). Precipitation regimes are extremely variable, with differences at scales spanning the whole of the Alps to more local-scale variations within individual valleys and slopes. Rapid snowmelt, intense precipitation events and summer thunderstorms are common features of the climate system that impact significantly on geomorphic processes in these vulnerable areas (e.g. severe floods in Piedmont 1994, Brig 1993, Vaison-la-Romanie 1992; Steininger et al., 2002).

Temperatures are also very variable, typically decreasing 1°C with each 100 metre increase in altitude (Foster, 2001). Seasonal and permanent snow cover, permafrost and glaciers all characterise the Alpine landscape and are an integral part of the climate system (Cebon et al., 1999). Snow depth, cover and duration vary with altitude (Schuepp, 1980). Throughout the Holocene, glacier cover has varied and was more extensive 250 years ago largely a function of greater precipitation, but cover has declined in line with increasing temperatures since late 19th century (Schonwiese et al. 1993; Zemp et al., 2006) and declined more sharply in recent years.

4.2.3 Past Regional Glaciations

In the Alps during the ultimate cold stage of the Pleistocene (Marine Isotope Stage (MIS) 4-2), two major glacial advances are thought to have taken place; the first around 60 000 BP (MIS 4) and the second around 28-20 000 BP (MIS 2) also called the Last Glacial Maximum (LGM) (Buoncristiani & Campy, 2004). During the LGM, in addition to the glaciers extending from the Mont Blanc massif down the Rhône valley, other ice streams flowed through the Isère, Arc and Arve valleys, they coalesced and extended westwards towards Lyon (Fig.4.1b). It is thought that the Jura ice cap (Fig.4.1b) was a discrete entity that had little interaction with the adjacent larger Alpine icesheet during the LGM (Buoncristiani & Campy, 2004). Rapid global warming across the MIS 2/1 termination led to the wasting and decline of the Alpine and Jura icesheets. During the Holocene the Alps have and continue to sustain glaciers at higher altitudes, with more extensive valley glaciers currently descending from cirque source zones down to altitudes of ~1550 m (Aletsch glacier). Long term monitoring and palaeogeographical reconstructions show that the margins of Alpine glaciers have varied through the Holocene (Holzhauser et al., 2005). Beniston et al. (2003) suggest that the duration of snow cover could be reduced by 50% (2,000 m) and by 95% (below 1,000 m) if there is a 4°C shift in mean winter temperatures as projected for the European Alps (IPCC, 2007 A2 emissions scenario). The reductions that the IPCC suggest would not be compensated for by any predicted increase in winter precipitation. This would produce an upward shift of the glacial equilibrium line (Vincent, 2002; Oerlemans and Klok, 2004) and encourage substantial retreat with smaller glaciers disappearing and larger glaciers perhaps reducing in volume by between 30% and 70% by 2050 (Schneeberger et al., 2003; Paul et al., 2004). These changes would have large implications for river flows in glaciated and non-glaciated catchments. In addition the extent of permafrost will become further restricted rising by several hundred metres and these temperatures and melting permafrost will contribute to de-stabilisation of mountain slopes (Gruber et al., 2004).

4.3 The Lac d'Annecy catchment

4.3.1 Situation

The Lac d'Annecy lies at an altitude of 446m in the Haute-Savoie region of the Western pre-Alpine zone of the French Alps (Lat. 45° 48'N; Long. 6°8'E). The lake is composed of two sub-basins, the Grand lac ($20km^2$) and the Petit lac ($6.5km^2$),and receives drainage from a total catchment area of $251km^2$ with the majority of the land ($170km^2$) draining into the Petit lac at the southern end of the lake (Fig.4.3).

The Petit lac d'Annecy has a lake to catchment area ratio of 1:27. The catchment is mountainous (average elevation 1036m) and has summit massifs reaching ~2400m.

4.3.2 Geology

The Petit lac d'Annecy lies within a transverse tectonic valley situated between the calcareous limestone (Cretaceous to Jurassic) sub-Alpine Bauges and Bornes Massifs (Fig. 4.1a) (Nicoud & Manalt, 2001). Alpine orogeny and reverse and strike slip faulting created a regional pattern of ridges running in a SSW to NNE direction (Foster, 2001) which have been eroded and shaped by Quaternary glaciations to form sharp arête crests. The lake lies in a SE-NW fault-controlled depression cross cutting the main structural trend through the Bauges and Bornes massif (Nicoud & Manalt 2001). The Petit lac catchment is dominated by Jurassic-Miocene marls and Cretaceous limestones (Fig. 4.2a).

4.3.3 Catchment Evolution

Lac d'Annecy is one of many surviving peri-alpine lakes in the pre-Alps, a remnant over-deepened basin scoured by ice during the last glaciations (MIS 4-2, Würmian) (Nicoud & Manalt, 2001). Evidence of the penultimate glaciation (Rissian, MIS 6) is scarce although studies suggest that an ancient lake may have been present in the Eau Morte valley (Fig.4.3) to a level of 510m above sea-level (Monjuvent & Nicoud, 1987), much higher than the current level of 446m. Further evidence of last interglacial environments is present in lignite beds (Bourdier, 1962) exposed in bank sections of the Bornette river south west of the Petit lac (Fig.4.3).

The rock lithologies contained in glacial deposits of MIS 2 (Würmian) age in the Petit lac catchment include schist erratics from the inner Alps and local limestones and marls which reflect the ice dispersal pathways of the Alpine glacier icesheets (Dearing et al., 2001). During the last glacial maximum, the valley was over-deepened to 220m a.s.l by the combined efforts of icestreams from the Arve-Mont Blanc glacier and further deepened by the the Rhône, Val d'Arly and Bauges icestream that coalesced near Annecy (Fig.4.2b) (Nicoud & Manalt, 2001). It is estimated that the MIS 4-2 glacier reaching an estimated ice thickness altitude of 700m-800m (Deleau, 1974) and evidence of Würmian glacial drift has been identified at 900m-1000m within the Petit lac catchment (Benedetti-Crouzet & Meybeck, 1971; Benedetti-Crouzet, 1972).

During the retreat of the Mont Blanc glacier (c. 15000BP, Campy, 1992), the Lac d'Annecy was much larger than the present day, extending to southerly shoreline near Faverges (Fig.4.3) and it is thought to have had a surface area of 63km^2 and surface altitude 460m (currently 20km^2 and 446m) (Nicoud & Manalt, 2001). The start of endogenic calcite formation by 14200BP reflects the rise to dominance of biogenic activity (Lee et al., 1988) over detrital sediment supply in the Grand Lac and indicates that the catchment was ice-free by this time (Brauer and Casanova, 2001). As the climate warmed during the late glacial, the Younger Dryas stadial (c.12 500 BP) interrupted this climatic amelioration by a brief (c.1300 years) return to cooler temperatures, sparking the re-advance of glaciers to the margin of the lac d'Annecy drainage basin (Foster et al., 2003). Field evidence of high altitude cirque glaciation in summit regions of the St. Ruph sub-catchment (Fig.4.3) with extensive moraine ridges of up to ~20m deep. Other field observations show the Bornette sub-catchment has 15m high kame terraces which are likely to be remnant from the LGM.

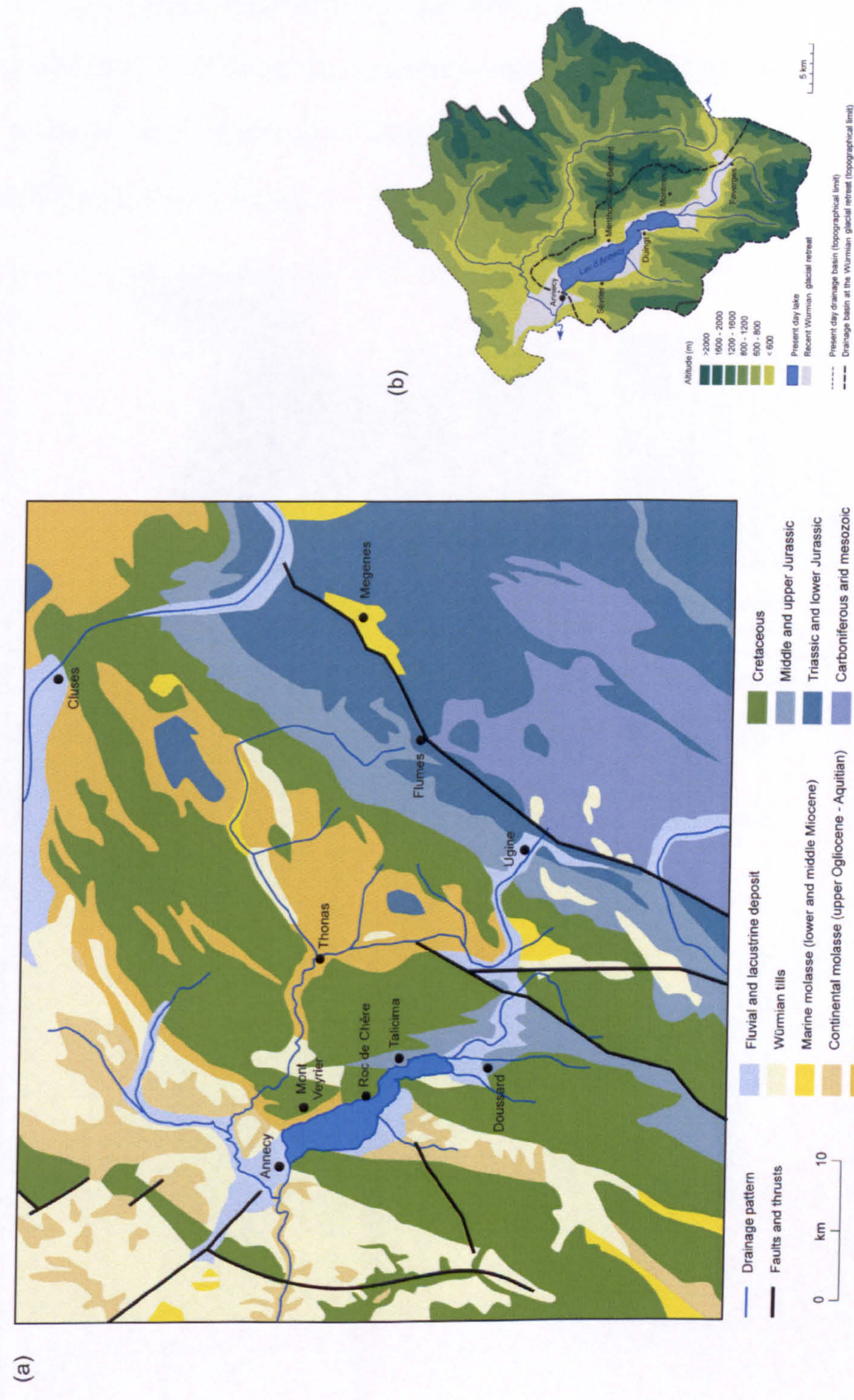

Figure 4.2 *(a) Lithostructural setting of the Annecy cross valley (after Nicoud and Manalt, 2001), (b)Recent and present day Lacs d'Annecy (after Manalt, 1998 and Nicoud & Manalt, 2001)*

During the retreat stages of the last glaciation a significant infilling of the valley took place (Nicoud & Manalt 2001), reducing both the physical area of the lake basin and the water level (446m). The configuration of the Lac d'Annecy catchment was altered significantly by an

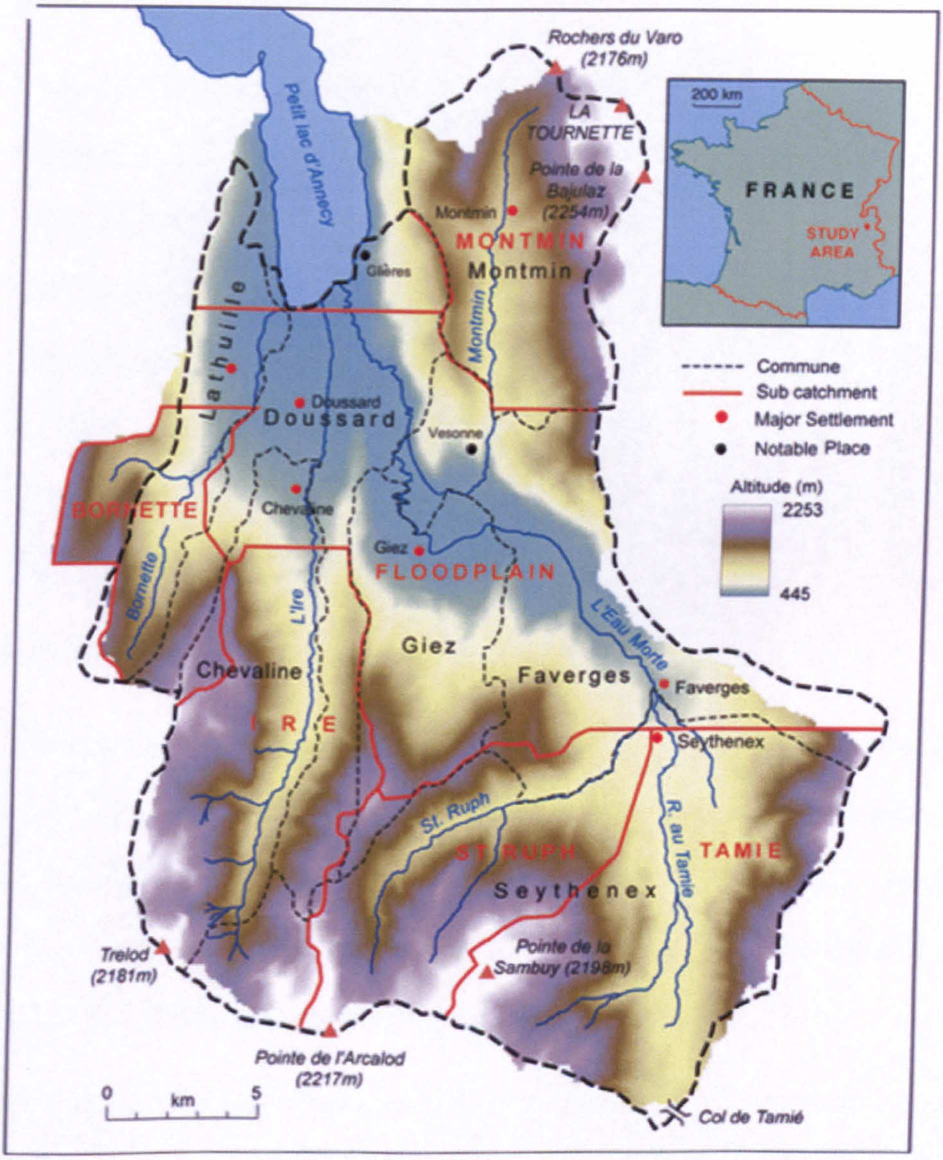

Figure 4.3 *Location map of Petit lac d'Annecy showing major inflows, lake basin, sub-catchment boundaries (for modelling purposes), commune boundaries, elevation, notable places, major settlements, major summits. Inset shows location of study area in France*

admittedly poorly understood river capture, where the Fier system was diverted across the low amplitude relief between Faverges and Saint Ferreol into the Chaise catchment ~ 6000-5000BP (David et al., 2001). Losing the Fier reduced the total catchment area draining into the lake from 440km^2 to the present day 251km^2 (Fig.4.2b).

Magny (2004) suggests that synchronous regional changes in lake levels can be indicative of climatic fluctuations such as changes in evaporation or precipitation. Archaeological evidence in the Petit lac catchment that suggests the lake water levels have been lower in the past, for example, Neolithic (~4000 BP) lake dwellings have been excavated in 3 metres of water below the present day water surface (Marguet et al., 1995). This suggests that previously, the lake has been at a lower level and is thought that this may be due to increased evaporation or decreased precipitation at this time (Nicoud & Manalt, 2001).

4.3.4 Climate

The mean annual temperature in the catchment varies with altitude. At the southern end of the Petit lac, the settlement of Glières (443m), (Fig.4.3) has an mean annual temperature of 9.5°C, whereas Montmin (1020m) (Fig.4.3) experiences mean annual temperatures of 6°C (Benedetti-Crouzet, 1972). Mean annual temperatures at the summits (~2400m) are estimated to be 0°C (Benedetti-Crouzet, 1972).

Meteorological records (between 1826 and 1996) for Annecy (altitude ~450m) on the northern shore of the Grand lac show that it received an average annual rainfall of 1250mm with a minima and maxima of 648mm and 1871mm respectively (Foster, 2001). February is the driest month on average with October recorded as the wettest month. These values vary across the altitudinal range; for every 100m in altitude gained, there is an increase in precipitation of around 70mm, along with a temperature reduction of around 0.6°C (Benedetti-Crouzet, 1972).

There are no glaciers in the catchment, and the current climate cannot sustain any permanent snow cover (Oldfield& Appleby 1991). However, precipitation during the winter months falls primarily as snow (November-March) followed by several months of snow melt (April-May). Studies in the 1970s show that between 6th November and 24th April, an average of 2.7m of snow was received by Montmin (1050m) and an average of 0.53m between 4th December and 3rd March at floodplain altitude (446m) (Richard, 1973).

Both the snow regime and high intensity (20-50mm hr^{-1}) convective summer (June-August) rainfall events play an important role in the hydrology of the catchment (Foster, 2001) Overall typical annual discharge patterns (Figure.4.4) show low average discharges in January-February due to snow storage in the catchment. There is a steady increase in average discharge in March as warmer temperatures bring the onset of a snowmelt period. April and May have a sustained time of higher discharges due to snowmelt before reducing slightly in June. June-August have incidences of peak discharges which are likely to be from summer thunderstorms followed by an increase in discharge from September to December which indicates increased precipitation in Autumn and Winter.

The NNW-SSE configuration of the catchment focuses the movement of continental air masses from the north bringing dry air, whereas maritime air masses from the west bring rain, thus winds in the catchment tend to have a northerly or westerly component (Higgit, 1985). The humid maritime air masses occasionally arrive in

December, thus raising the temperature, and create a peak in flood discharge, due to early onset of snow melt (Higgit, 1985).

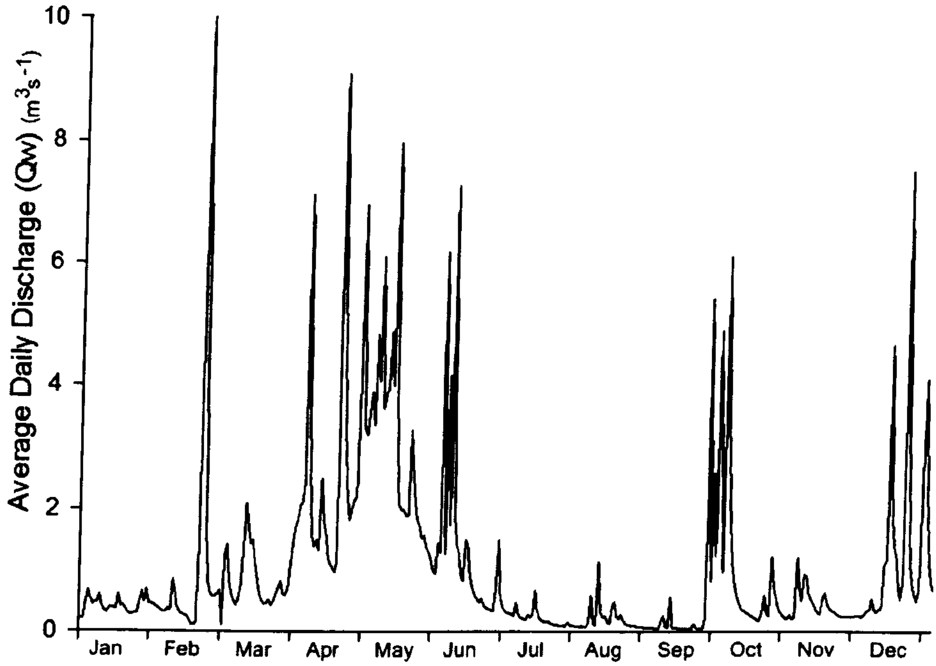

Figure 4.4 *Average daily discharge for the Ire sub-catchment, 1999 (Source: CLIMASILAC)*

4.3.5 Vegetation History

Understanding of the regional vegetation history since the most recent deglaciation, after 15,000 BP, is facilitated by pollen diagrams obtained for both basins of Lac d'Annecy. There have been three main pollen analyses undertaken for both the Grand lac and Petit lac; firstly David et al., (2001) analysed a 44m core from the Grand Lac which gives an account of the vegetation history from the last ~15000 years. Secondly Noël et al., (2001) analysed a 813cm core (LA13) from the Petit lac which spans approximately ~6000 years and finally Jones et al., (in prep.) have analysed the top part of core LA 13 in light of new data (Nomade, 2006) and improved chronological control. Some of general trends from Noël et al., (2001) are still valid but the regional vegetation description here and later magnetic proxies make use of the most recent chronology outlined in Jones et al., (in prep.).

During the early Holocene the landscape stabilised, and climatic amelioration and soil development enabled the establishment of an extensive deciduous forest cover (~10 000 BP - ~ 8000 BP). These forests were dominated by *Pinus* and *Quercus*, until ~ 8000 BP when *Quercus, Tilia* and *Abies* became dominant (David et al., 2001). These early Holocene environments were characterised by a full forest cover and well developed catchment soils which in turn encouraged a reduction in catchment erosion and low suspended sediment loads and delivery to the lake (Foster et al., 2003). A marked decrease in *Pinus* percentages from 80% to 50% are noted at ~9650 ^{14}C yr BP which coincides with the first Holocene climatic deterioration phase that was noted in the Alps (Bortenschlager, 1978).

During the *Quercus* maximum at 8150 ^{14}C yr BP, a maximum peak in *Tilia* and expansion of *Abies* occurred. *Abies* dominated before declining sharply at ~4500BP (Jones et al., in prep.) during which time *Fagus*, *Alnus* and *Quercus* increased to their maxima. The changes in forest composition at this time may signify the end of the Holocene thermal climatic maximum (Jones et al. in prep.) The dramatic decline in *Abies* combined with increasing non-arboreal pollen concentrations, particularly grasses (including cereals) and other pastoral/arable indicators at around 3000 BP, reflect the reduction of forest cover owing to human activity such as fire clearance during this period (Tinner et al., 1999). The first appearance of cereal grains and other agricultural indicators in the Petit lac record highlight the onset and expansion of anthropogenic settlement in the catchment. The later Holocene forests (~2300BP) show evidence that deforestation may have taken place due to a large sediment peak in organic matter in lake sediment (Jones et al. in prep.). This is co-incident with a decline in *Alnus*, suggesting a potential early phase of clearance on the flood plain and perhaps beyond. The introduction of *Juglans* around 2000BP is linked to Roman settlement in the region (Jones et al., in prep.; Higgit, 1985). Another series of sediment peaks in the lake sediment accumulation record occurr between 900 AD - 1000 AD and were coupled with a peak in both the NAP around 1000 AD and an increase in agricultural indicators such as *Juglans* and cereals which therefore suggest agricultural expansion at this time.

The declines in forest cover during the later Holocene concord with archaeological and historical evidence that outlines human settlement from the Neolithic period onwards (Privat, 1973; Jones et al., in prep.). During Iron Age and Roman times (120 BC to AD 260), the town of Casuaria (Faverges) was established and the valley was an important trade route through the Alps. Various nationalities controlled the region during medieval times, including Sardinians, the French, but also short occupations by both Germanic and Spanish-populations took place (Jones et al., in prep.). Monastic communities (both Benedictine (AD 879) and Cistercian (AD 1132)) were important for agricultural development in the uplands encouraging population growth punctuated by catastrophic declines driven by epidemics (the Black Death AD 1348) and climate (early 1300's and 1400's) (Crook et al., 2004). Population levels recovered after AD 1470 and continued growth led to agricultural and economic systems experiencing severe stress thus encouraging expansion of cultivation and grazing into higher altitude areas ~AD 1550. Climatic downturns of the Little Ice Age may have punctuated this expansion of agricultural land, but the area of cultivated land appears to have increased at the expense of forest due to peak population levels ~ AD 1850-1870. Thereafter, abandonment of hill pasture, the regeneration of woodlands and planting of new forests have led to a gradual increase in catchment forest cover (Crook et al. 2002; 2004).

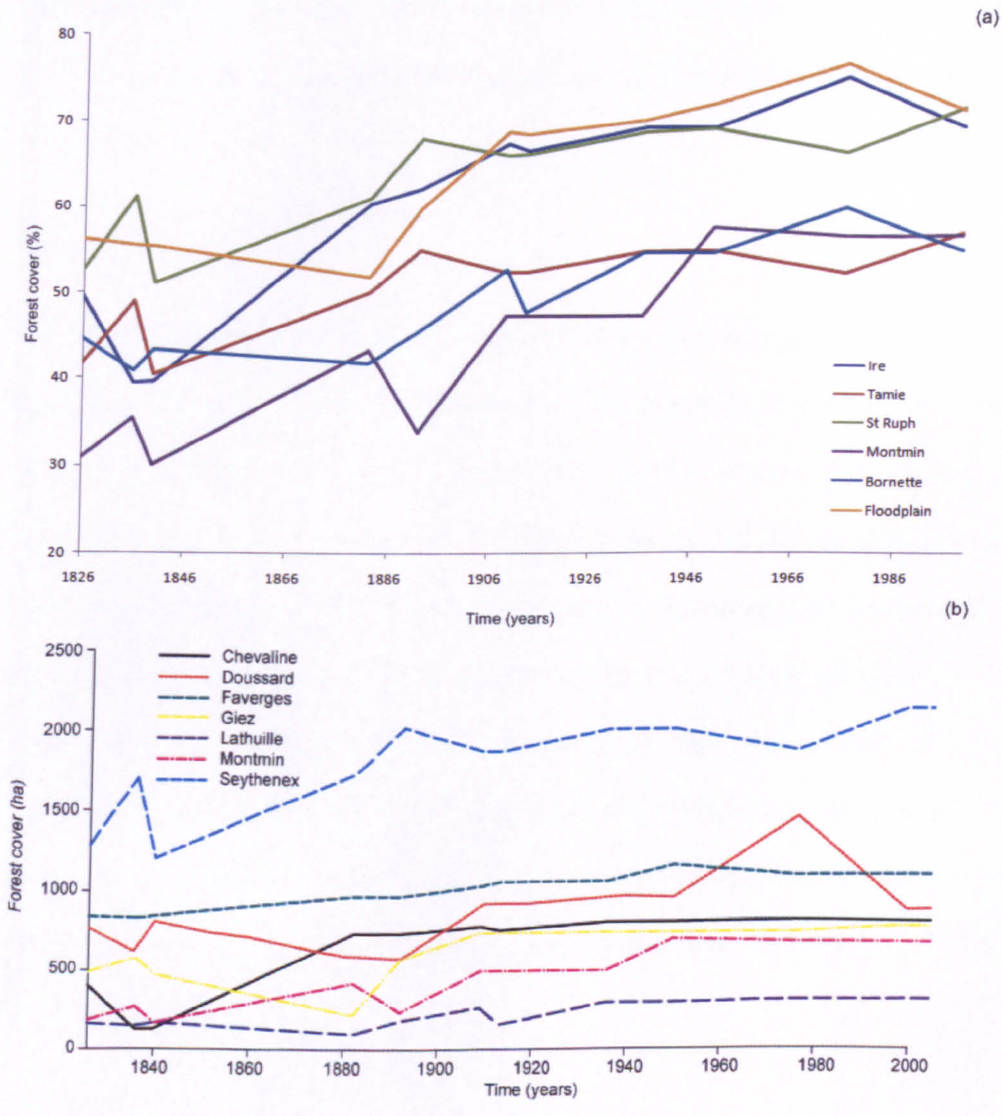

Figure 4.5 *(a) Calculated percentage forest cover per sub-catchment, see 5.5.5, (b)Forest inventory records for communes within the Petit lac d'Annecy (after Crook et al., 2002)*

Forest inventory data are available for each of the communes (Crook et al., 2002) from ~1730 AD with more complete records available from 1826 AD onwards (Fig

4.5b). These records are discussed in the more detailed descriptions of the individual sub-catchment sections (section 4.4). In addition to forest cover from historical records from communes, data for the sub-catchments delimited in this research have been calculated (Fig.4.5a) from the commune records by area-weighted mean (see 5.5.5 for further details) as the commune and sub-catchment boundaries are different (Fig.4.3).

4.3.6 Hydrology

The three major rivers, the Eau Morte, Ire and Bornette, drain the pre-Alpine terrain from the highest summits at La Tournette (2358m) (Fig.4.3), extending down to the extensive pre-lake flood plain or delta at 446.7m . The Eau Morte is the largest of the three tributaries draining an area of 130km^2 and is fed by several tributary headwaters, St Ruph, Tamie, Bar de Tamie and the Montmin, which converge to form the 'Eau Morte' when as it passes through the town of Faverges (Fig.4.3). For the purposes of modelling, the Petit lac catchment has been divided into sub-catchments based on drainage patterns that were determined using ESRI ArcHydro software (see section 4.4 and Chapter 5 for further details). It is therefore more practical to describe the sub-catchments in detail rather than the Petit lac d'Annecy catchment as a whole.

4.4 Description of the sub-catchments

A brief summary of the key characteristics is given in Table 4.1. The Petit lac d'Annecy catchment varies in gradient (Fig.4.6), from flat valley floors (<10°) in the Tamie and Eau Morte flood plain up to steep, almost vertical cliffs (~70°) in the Ire, Bornette and Montmin sub-catchments. Various stream networks lie within the Petit lac d'Annecy catchment and for clarity are named according to the IGN maps (Fig.4.7)

Table 1 Key characteristics of subcatchments with the Petit lac d'Annecy catchment

	Area (km^2)	Dominant geology	Hydrology	Modern day vegetation cover	Geomorphology
Bornette	10.8	Cretaceous marl and limestone. Scree slopes	Mountain torrent	~70% forested	Scree and debris covered slopes
Ire	21.6	Cretaceous marl and upper/middle Jurassic limestone. Scree slopes	Mountain torrent, constrained bedrock channel	~ 60% forested	Extensive hillslope colluvial deposits, active well coupled gully systems, avalanche chutes, debris cones, feeder
St Ruph	17.3	Cretaceous marl and limestone. Scree slopes	Mountain torrent	~ 60% forested	Extensive hillslope colluvial deposits, active gully systems, debris cones, feeder fans, debris flows, mass movements
Tamie	24.4	Cretaceous marl and limestone. Quaternary glacial till	Lowland river system with steep headwater tributary	~50% forested	Tributary alluvial fans, palaeochannels, hillslope colluvial deposits, avalanche chutes, active gully systems
Montmin	15.5	Cretaceous marl and limestone	Perched valley with deeply incised bedrock gorge	~50% forested	Active gully systems and soil creep processes, scree slopes
Eau Morte Floodplain	42.1	Quaternary alluvium, glacial till and sand and gravel. Largely Cretaceous marl with some limestone	Limited natural meandering reaches, dominated by twentieth-century canalised channels	~25% forested	Flat low lying alluvial plain, with palaeochannels, and extensive low-angle alluvial fans in the mountain front areas

***Table 4.1** Petit lac d'Annecy sub-catchments, selected characteristics*

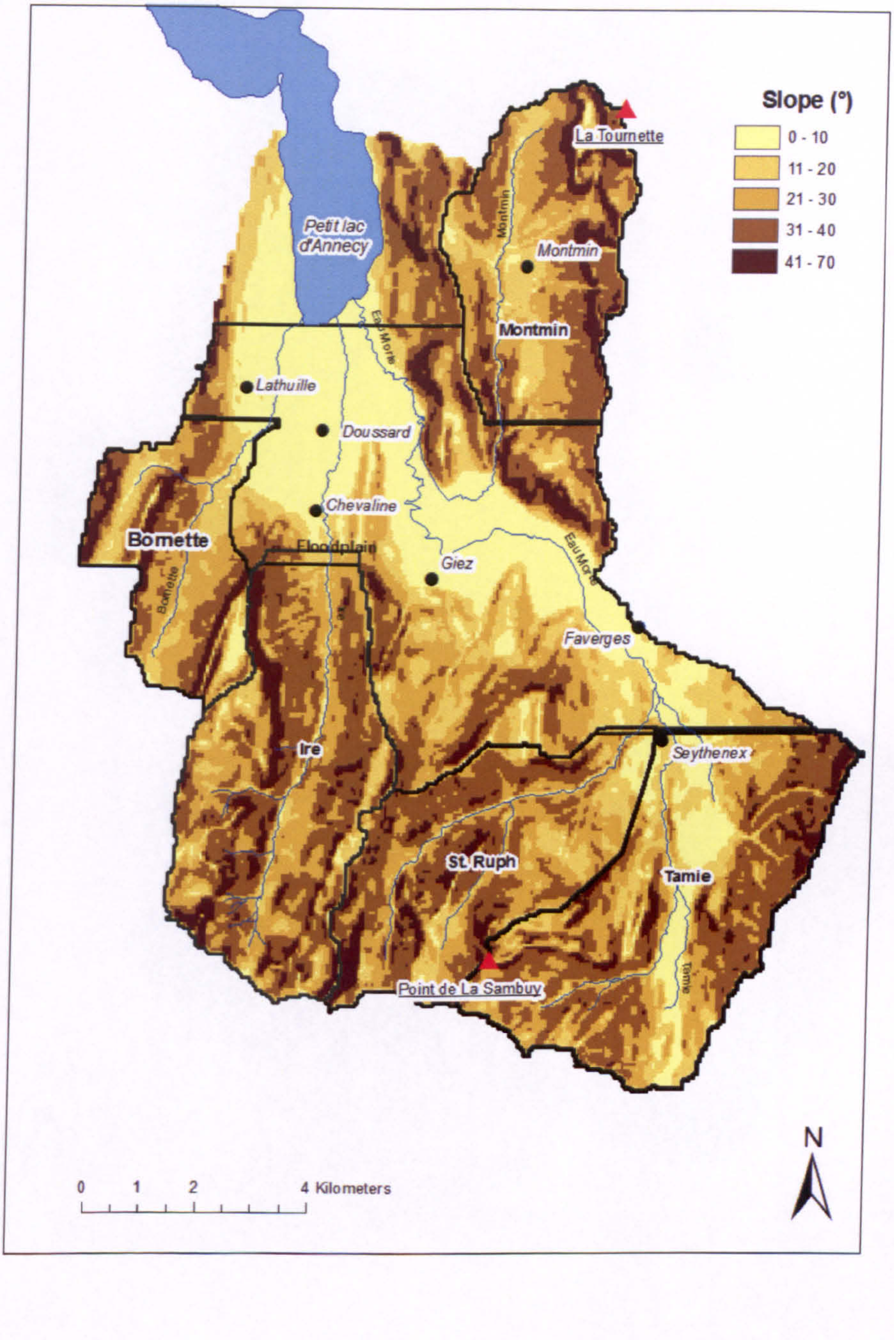

Figure 4.6 *Distribution of slope angle across the Petit lac d'Annecy catchment.*

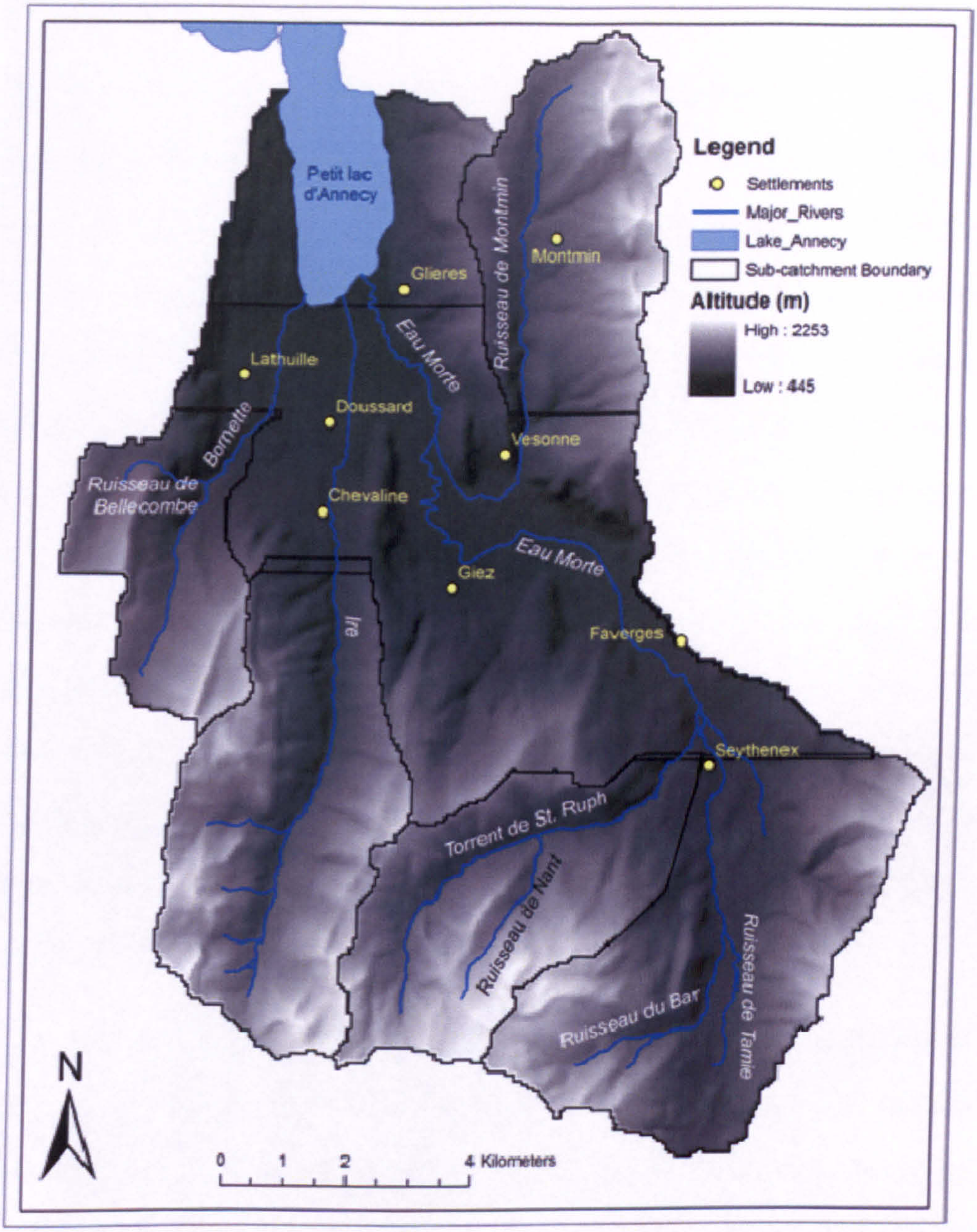

Figure 4.7 *Stream and river network within the Petit lac d'Annecy catchment*

4.4.1 Tamie

The two main streams in the Tamie (Fig.4.3), the larger Ruisseau de Tamie and the Ruisseau de Bar (Fig.4.7) drain a significant portion of the Petit lac d'Annecy catchment ($24.4m^2$). It is a diverse system, commencing in the steep sloped (30-50°) headwaters as mountain torrents (Ruisseau de Bar, Fig.4.7) along the watershed between the Tamie and St Ruph sub-catchments before flowing onto a flatter floodplain (< 15°) (Fig.4.3, Fig.4.8a) where the two streams converge with the lower reaches of the St. Ruph River at a confluence upstream of the town of Faverges and downstream of Seythenex (Fig.4.3). On entering the pre-lake flood plain at Faverges, the river becomes the Eau Morte. The bedrock geology of the sub-catchment is dominated by Cretaceous limestones, with less dominant Cretaceous marls also present (Fig.4.2a). Locally, bedrock is overlain by glacial diamicton both in situ and remobilised by hillslope processes, with thicker accumulations at lower altitudes, on shallower gradient slopes and in occasional higher level basins (Fig.4.3). There is some evidence of karstic drainage within this sub-catchment.

Exposed bedrock, debris slopes and snow avalanche chutes are prominent features of the upper slopes, and help to feed sediment to the lower slopes. Snow induced avalanches scars (Fig.4.8b) are common on the hillslopes. The lower reaches of the Tamie differ from the other mountain torrent dominated sub-catchments as there is greater potential for sediment storage on the wider floodplain. At the confluence of the Ruisseau de Tamie and Ruisseau de Bar (Fig.4.7) there is an extensive and active alluvial fan with several fan terraces, the surface of which displays well defined palaeochannels. Over the course of the Holocene, the Ruisseau de Tamie and Ruisseau de Bar appear to have varied in dominance in this alluvial fan complex with channels migrating laterally for the most part in an aggradational setting.

Chronologies are not available for these palaeochannels but the braided channel planform is suggestive of a sediment-rich environment, something which is also evident in the current channel.

Figure 4.8 *A. Views downstream from the Tamie floodplain upstream of the Ruisseau de Tamie and Ruisseau de Bar confluence, with the extensive former floodplain of the Ruisseau de Bar currently against the hillslope on the right of the photograph. Also shows the forested hillslope and pasture regularly mowed for winter silage. B. Avalanche chutes scar hillslopes.*

The glacial till, with both carbonate matrix and clast, varies in thickness across the montane valley floor (~20m), and encouraged the development of mature calcareous brown earth soils (Foster et al. 2003) and cultivation and pasture land uses. This sub-catchment in particular has been a preferred area for settlement and farming due to the wide (~2km wide), low altitude valley floor (600 m - 900 m) and ease of accessibility compared with the other sub-catchments (Fig. 4.3).

The commune of Seythenex lies within the boundaries of the Tamie sub-catchment, with the commune of Faverges lying at a lower altitude just outside of the boundaries of the Tamie sub-catchment, yet the trends seen in forest inventory for Faverges are likely to have had an influence on the Tamie sub-catchment. According to historical forest records, the woodland use for the commune of Faverges has remained relatively stable (Fig. 4.5b) although there have been some fluctuations in the Seythenex record such as a reduction in forest cover at AD 1840. Calculations of woodland use (see Chapter 5 and Fig.4.5a) from forest inventory data (Fig.4.5b) (Crook et al., 2002) and aerial photographs (Google Imagery, 2008) for each sub-catchment suggest that the forest cover has remained relatively stable since 1826, with a small decline around 1840 and an increase through the late 19th century, followed by a small decline in the early 20th Century. The steeper slopes are at present largely forested which has contributed to reduce the potential for soil creep and soil erosion, though both processes are likely to have been more dominant during periods of deforestation and/or under a wetter/climate.

4.4.2 Ire

The Ire River occupies a south to north aligned valley deeply incised into a pre-dominantly Upper/Middle Jurassic limestone bedrock, with some Cretaceous marl outcrops (Fig.4.2a), where steep valley sides rise some 1000-1400m to the mountain summits. From a perspective of human occupation, the Ire is relatively inaccessible, with little potential for settlement around the channel, though the headwater and flatter summit regions have been exploited as alpine summer pasture.

The river is a gravel bedded mountain torrent with a mixture of bedrock constrained reaches, but also with substantial accumulations of alluvial gravel on a narrow restricted floodplain supplied by numerous alluvial fans draining gully systems that fret the valley sides. The hillslopes are draped with a mixture of glacigenic and hillslope debris that has been redistributed into the central and lower reaches of the valley. Accumulations appear thicker and more extensive on the west facing slopes. The Ire occupies a reach which has incised (up to 100m) into these deposits probably during de-glaciation. Tributary gully systems fret the lower debris slopes adjacent to the main channel. They are incised ~10m-20m in depth, ~10m in width and extend 100m-200m in length up the steep sided drift-mantled slopes (40-70°, Fig.4.6).

Bedrock cliffs crop out at higher altitudes, and are more extensive on the debris-free western flank of the Ire. The Ire, at present, receives a substantial sediment load from the headwater and hillslopes. There is very little capacity for sediment storage except for in and adjacent to the channel. The capacity for storage is reduced in the restricted gorge reach before entering the wider expanse of the pre-lake floodplain (not included in the limits of the sub-catchment presented here, see Fig.4.3).

Chevaline and Doussard are the communes that lie within this sub-catchment are Chevaline and Doussard, and the Ire river forms the commune boundary (Fig.4.3). For the purposes of using the forest inventories and historical data to discern land use patterns (see Chapter 5), the Ire is assumed to reflect the average forest cover conditions of the two communes (Fig.4.5b), though Chevaline may be more reflective given the proximity and accessibility of the area for settlement and human use (Fig.4.3).

Figure 4.9 *High altitude pasture land is found within the Ire, on the SSE, W and SW flanks and is mainly used for grazing (2007)*

The forest inventory records (Fig.4.5b) for Chevaline and Doussard and the area-weighted forest cover (Fig.4.5a) for the Ire sub-catchment suggest that there was a decline in forest cover prior to 1840, with the lowest forest cover recorded from 1826 and 2005. The low forest cover was largely due to increased population and invasion of Austrian military during the Napoleonic wars which had a devastating effect on the woodland in the Ire and Tamie valleys (Crook et al., 2002). Much of the Ire is too steep and inaccessible for farming, except for zones of high altitude alpine summer pasture (Fig. 4.9). Large parts of the sub-catchment were cleared to sustain the timber trade, which was a lucrative market during the 19th century (Crook et al., 2002). Following the Treaty of Turin in 1860, and the implementation of long-term woodland management (Crook et al., 2002), an increase in forest cover took place before decreasing to approximately present day values in 1910 as a result of land fragmentation and the desire for pasture land (Crook et al., 2002). The modern day forest cover lines the steep hillslopes, nevertheless field observations suggest that there are active gullies some re-vegetated and recent land slips. Soil creep and soil erosion are likely to be prominent features of this system, more so in the absence of complete forest cover particularly on the steeper slopes. The channel and hillslopes appear well coupled throughout due to the extensive sediment-rich gully systems that are evident even in areas with ~60% forest cover. For the purpose of modelling, the DEM has been clipped above the flood plain (e.g.Fig.4.3). Therefore all the modeled results for this sub-catchment reflect a mountain torrent, often bedrock-confined system that has limited capacity for storage of sediment.

4.4.3 St. Ruph

The St Ruph sub-catchment (Fig.4.3) lies between the Tamie and Ire and occupies a south to north east aligned valley which is deeply incised into Cretaceous limestone and marls (Fig.4.2a). It includes one of the highest peaks in the catchment, Point de la Sambuy (~2200m) and the drainage boundary lies close to the settlement of Seythenex at the base of the drainage area. This catchment has a small ski station at ~ 900m operative during the winter months, and is one of the busiest tourist regions in the Petit lac catchment. The two main tributaries that comprise the St Ruph river are the Torrent de St Ruph and the Ruisseau de Nant (Fig.4.7). The tributaries are both gravel-bedded mountain torrents supplied with substantial amounts of alluvial gravel via the alluvial fans issuing from valley side gullies. The steep slopes (30-

40°, Fig.4.6) in the headwaters of the Ruisseau de Nant display extensive evidence for active debris flows and associated gullying contribute to the lower debris-mantled hillslope by 10-20 metres (Fig.4.10). Lichenometric analyses date some of these features to within the last 100 years (Foster et al., 2003), though others are clearly older.

Figure 4.10 *Field evidence of debris flows in the St Ruph sub-catchment (2007)*

These steep slopes are frequently grazed by cattle, and are largely un-forested above the tree line. Some small shrubs are evident, with a rapid re-growth of vegetation evident in the last 10 years (Dearing, 2007, observation). Soil creep terraces are evident on steeper slopes in un-forested areas of the catchment. Historical forest records suggest that Seythenex (Fig.4.5b) had a small decline in forest cover around

AD 1840 followed by a steady upward trend in increasing forest cover. The calculated percentage of forest cover for the St Ruph sub-catchment (Fig.4.5a) suggests that forest has increased from ~50% to 70% since 1826. The larger Torrent de St Ruph drains a lower altitude and mainly forested area of the sub-catchment with some alpine pasture at higher altitudes (Fig.4.9). Downstream, closer to the settlement of Seythenex, large (~20m long, ~8m deep) sediment-rich gully systems feed the sediment-choked St Ruph River (Fig.4.11). There is little opportunity for lateral migration of the channel as it is largely confined by the steep slopes.

***Figure 4.11** - Sediment rich St Ruph River near to Seythenex.*

4.4.4 Montmin

The sub-catchment of Montmin lies directly east of the Petit lac d'Annecy and includes one of the highest peaks in the catchment, La Tournette. Although elevated above the main Eau Morte valley, the sub-catchment has long been a popular area for agriculture due to the low gradients of the main valley(< 20°) and good quality soils of the central area (Fig.4.12). Two small tributaries feed the main river, the Ruisseau de Montmin, which flows through a deeply (>100m) incised bedrock gorge dominated by Cretaceous marls and limestone before exiting the sub-catchment and joining the Eau Morte river near to the small hamlet of Vesonne (Fig.4.3).

Figure 4.12 *Montmin from Col de l'Aup showing steep unforested slopes on the left of the photo, with small parts of alpine pasture before moving down into a forested belt of land. Finally, the lowest gradients are used for grazing. (Dearing, 2007)*

Contemporary land use is mixed with a large amount of grassland used for grazing and for silage and some degree of crop cultivation. The steep slopes on the eastern flank of the sub-catchment around La Tournette have extensive scree slopes and gullies. On the lower steeper slopes, much of the land is forested. Soil creep is a prominent process seen on many of the slopes. Due to the suitability for agriculture, farming and grazing in this sub-catchment, area-weighted calculations from historical records (Fig.4.5b) for this sub-catchment (100% dominated by the commune of Montmin) have shown that there was only around 30% forest cover in this sub-catchment at the start of the 19th century (Fig.4.5a) rising to a maximum of 50% at present day.

4.4.5 Bornette

The Bornette River is the smallest of the three inflows into the Petit lac d'Annecy. The bedrock geology of this catchment is dominated by Cretaceous marls and limestone. The sub-catchment consists of two small valleys that cut into a deep sediment fill which is partly valley side kame terrace (mixture of glacial diamict, glacifluvial and glacilacustrine deposits at 600-800m asl) and glacial diamicts from the last glaciation (Fig.4.13). The rounded and angular large clasts in some of these sections suggest that they may be of fluvio-glacial nature rather than glacial diamict. Forest dominates the contemporary land use in the Bornette, and is currently at the highest level since the historical forest inventory records began ~ 1826 (Crook et al., 2002). At high altitudes, above the forest, there are zones of alpine summer pasture which are used for grazing purposes. Past land use was largely dominated by trends in the commune of Lathuille (Fig.4.5b) and area weighted means show a small drop in forest cover in the early 19th to around 40% with an upward trend to ~55% forest cover in the present day (Fig.4.5a)

The fluvial regime is composed of two main channels, one of which is perched above the main channel in a smaller tributary. Steep bedrock cliffs dominate the catchment, and they play a structural role controlling the drainage pattern. There is some opportunity for floodplain storage here as parts of this sub-catchment are of gentle gradient (< 30°). The channels that flow through are relatively small and in contrast to the St Ruph and Ire systems there are few large tributary gully systems.

Figure 4.13 *3-4m Kame terrace in Bornette sub-catchment; some sections have a fill of up to 15m*

4.4.6 Eau Morte Floodplain

The Eau Morte floodplain reach (Fig.4.3) covers c.40km^2 of the catchment and has an average elevation of 705m above sea level. The three rivers that flow across the floodplain are the Eau Morte and the lower reaches of the Ire and the Bornette rivers.

Prior to the land drainage and river management practices associated with the extensive agriculture practiced on the pre-lake floodplain/delta draining into the Petit lac d'Annecy, the hydrological regime was dominated by the shallow gradient, low-lying meandering reaches of the Eau Morte river (Fig.4.14). Since late pre-history (Crook et al., 2002) the area has been a focus for intensive agriculture and settlement owing to the flat terrain and fertile soils. Above the floodplain, forest dominates the steep slopes (~20-30°).

In addition to the Eau Morte river, the lower reaches of the Ire and Bornette rivers pass through the floodplain, and have been canalised since the early 20th century (Crook et al.,2002). The river banks and lake shore line are flanked by riparian woodland, but other than this the flattest zones of the floodplain (< 20°) are largely unforested. The Eau Morte flood plain is a large sink of fine-grained sediment storage from the tributary streams that lose energy at these lower gradients (Foster, 2001).

According to forest inventory records from the various communes (Crook et al., 2002), forest cover dropped to its lowest levels in 1880 before increasing from 1900 onwards to the present day levels of forest cover. Riverine flooding characterises the recent history of the region, with large floods recorded particularly close to Faverges (e.g. 1734, Crook et al., 2002) (Fig.4.3). Faverges lies along the poorly defined, shallow gradient watershed between the Eau Morte valley and the easterly Ugine valley. Historical maps document a large channel avulsion in 1734 close to Faverges which was most probably channel migration on the low gradient watershed boundary. Flooding caused a great deal of damage, for which inhabitants of the settlements were compensated for loss of land (Crook et al., 2002). By the early 20th century, large sections of the Eau Morte were canalised and are now intensively managed to reduce flooding (Foster, 2001). River management has reduced the opportunity for natural channel migration and meandering, though some degree of overbank flooding does regularly supply sediment across the flood plain, with flood layers evident in river bank sections along the Eau Morte.

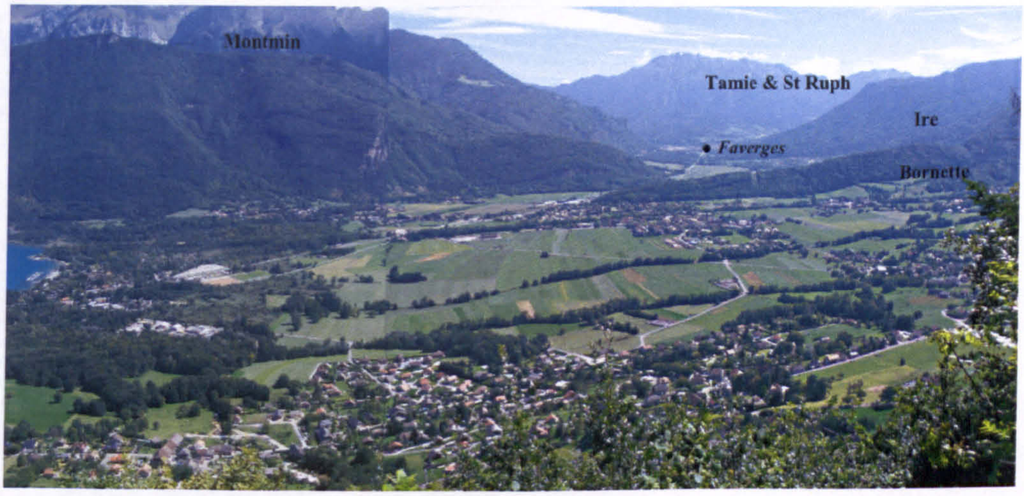

Figure 4.14 *Eau Morte flood plain. The Petit lac can be seen to the left of the photo, and the approximate location of the other sub-catchments are marked on the photograph. The town of Faverges can be seen in the distance (marked with black dot).*

Chapter 5

Methodological Framework

5.1 Introduction

The complex nature of fluvial systems at catchment scale and the difficulties in attributing changes in the system to human or climatic drivers highlights the need for a better understanding of these systems, particularly given the mounting pressures from uncertain climate, population and economic futures. This chapter describes the methodological framework of this research and outlines how numerical models such as CAESAR can be applied in conjunction with palaeoenvironmental records to attempt identification of system response to drivers of change. The general methodology is described, followed by detailed accounts of model setup, the palaeoenvironmental records used and the methodology for the creation of long term hourly model input files. Further chapter-specific methodological detail is provided in the five results chapters that follow which test particular aspects, hypotheses and model developments.

5.2 Linking the past to the present

Typically modellers of contemporary systems (e.g. fluvial systems) rely on empirical data in two ways. Firstly empirical data can be used as an input (e.g. precipitation) to drive a simulation and secondly another unconnected empirical record can be used for comparison with simulation output (e.g. river discharge) to assess the robustness of the model. For short term studies, these are acceptable approaches, but for longer studies, this method is limited because instrumental records are typically only available for around ~200 years. In order to extend studies further back into the past, to measure long term trends and identify past behaviours, proxy information from palaeoenvironmental records (e.g. lake sediments, ice cores and tree rings etc.) can be used to discern temperature, vegetation, precipitation, sediment erosion and land use histories all of which provide important information about the past and contribute to understanding of the sequence and drivers of past environmental change.

The difficulties in ascertaining the relative contributions of human and climate drivers in driving change in palaeoenvironmental records are long acknowledged (Dearing et al., 2006b). Attempting to explore these research questions in a modelling framework, allows individual forcings (in this case climate and land use (anthropogenic) change) to be isolated and compared to model outputs to simulate which combination of drivers that has the greatest impact. Palaeorecords are ideally placed to provide both long term forcings and the means of testing model robustness (Dearing, 2006b), but the variables used must remain independent of one another to avoid any circular reasoning.

In order to extend the focus of this thesis beyond the instrumental record, proxy evidence for land use and climate have been used to fill the gap that exists before the available documentary and instrumental records. For example, proxy records could be used to scale the hourly climate and land use time series required for input into long duration model simulations. To assess the robustness of CAESAR simulations, sediment erosion histories are required and these are available from lake sediment records through proxy records of change (e.g. environmental magnetism, geochemistry and grain size) which can be compared with simulated sediment discharges.

If the trends of proxy evidence in palaeoenvironmental and simulated time series are comparable, then CAESAR can perhaps be considered robust and subsequently used to predict the future responses of catchment processes to changes in climate and land use. Environmental changes throughout the next century are a subject of considerable interest to scientists, policymakers and lay audiences. Dearing (2006a) suggests that there is a need to place studies of contemporary lake systems into the longer term perspective through the combination of simulation models and palaeorecords. By combining complementary empirical data, palaeorecords and simulations (Dearing et al., 2006c), tools can be created that may allow us to better understand the past and subsequently look forward and perhaps mitigate against undesirable non-linear behaviour in systems as they respond to future changes (Fig.5.1).

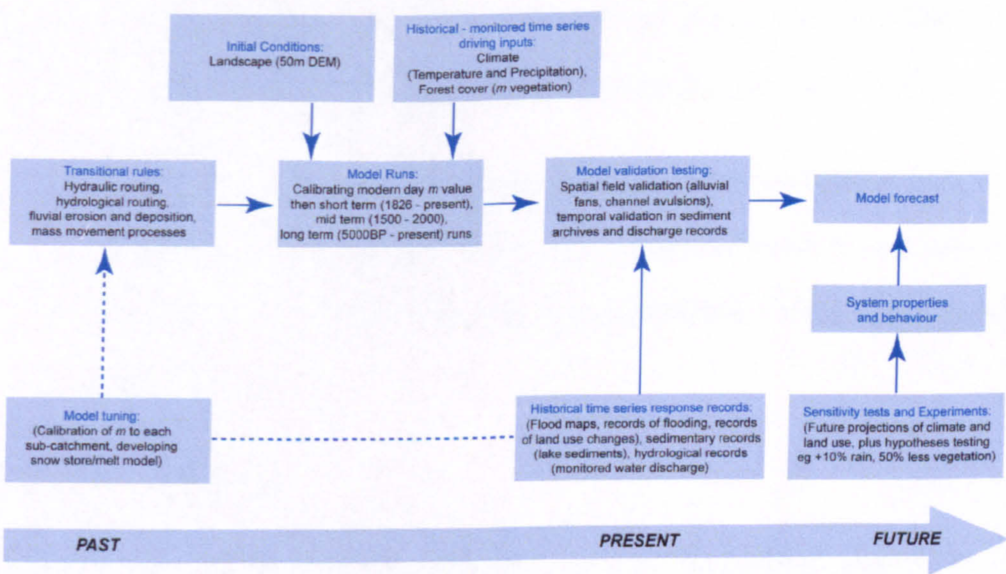

Figure 5.1 *After Dearing et al. (2006) application of methodology to the Petit lac d'Annecy catchment*

5.3 Model Input File Setup

5.3.1 Temporal resolution

The aim of this thesis is to enhance and assess the robustness of CAESAR model, the assessment of which is achieved by comparison of simulated and palaeodata. As with all modelling, an infinite number of possible hypotheses can be tested, therefore it is necessary to constrain the simulations so that they produce the best possible results to fulfil specific objectives. Various time scales have been used here to test CAESAR. These include:

- Short term instrumental data (10^0-10^2) which are used to assess robustness of the hydrological component of the model and to calibrate land use values (section 5.5).

- High resolution (monthly/seasonal) meteorological data that run over decades/centuries and are of similar resolution to projections for future climate (e.g. IPCC scenarios) to test theories of how extreme precipitation events and changes in land use affect the catchment (e.g. Foster et al., 2003).

- Low resolution (sub-decadal) proxy data (e.g. palaeoprecipitation from peat bog surface wetness records) that span century/millennia to test previous theories of how changes in land use and longer climatic change affect system response.

5.3.2 Data Constraints

Ideally, long term hourly records of climate, land use, water discharge and sediment discharge would be available to carry out these experiments, but in reality, records are often low resolution and fragmentary in nature. The data available for the Petit lac d'Annecy are constrained, in some cases by availability or proximity of data (Table 5.1) or otherwise by the resolution of data (Fig.5.2).

Record	Local	Regional	Continental
Precipitation data			
Hourly	10 years	-	-
Monthly	124 years*	174 years	-
Seasonal	-	500 years	-
Annual	-	-	-
Decadal/Sub-centenial	-	-	>2000 years
Temperature data			
Hourly	10 years	-	-
Monthly	174 years	-	-
Seasonal	-	500 years	-
Annual	-	-	-
Decadal/Sub-centenial	-	-	1973 years
Land Cover data			
Hourly	-	-	-
Monthly	-	-	-
Seasonal	-	-	-
Annual	-	-	-
Decadal/Sub-centenial	186 years	2000 years	-
Discharge data			
Hourly	2 years	-	-
Lake Sediment Records			
Decadal/Sub-centenial	~3000 years	-	-

* Incomplete record

Table 5.1 *Resolution, proximity and availability of data for the Petit lac d'Annecy*

There are complete records of hourly river discharge data available for the Ire for ~2 years, longer fragmentary records are available for around ~5 years with many missing data points and are therefore less suitable as they may omit many of the high discharge events. These high resolution dataset are useful for testing the precision of the simulated hydrological outputs from CAESAR and also for calibrating the contemporary land use values (*m* value, see section 5.5) used in simulations.

When considering the longer time scale, the availability and resolution of instrumental data is greatly reduced, and certainly not available at hourly increments. In order to identify longer trends in the water and sediment discharges, the recent past (1826-2005) is considered as there are limited hourly meteorological data available (1995-2005) and extended meteorological records available at monthly resolution for the whole catchment (1826-1995) (section 5.4.3). Historical forest inventory data (Crook et al., 2002) are also available for the individual communes in the Petit lac d'Annecy catchment and these provide reasonably accurate land use histories (1735-1999). For longer simulations that identify the effects of long-term climate (rather than short-term meteorology), gridded seasonal data (precipitation and temperature) are available for the region covering the period AD 1500-2000 (Luterbacher et al., 2004, Pauling et al., 2006). The period AD 0-1973 AD is the longest time scale simulated here, constrained by the availability of a continental scale synthesis of proxies that have reconstructed palaeo temperature data (Moberg et al., 2005), bog surface wetness data from peat sequences (Charman et al., unpublished) and proxy land use information from pollen-based reconstruction of vegetation history (Jones et al., in prep) which are increasingly poorly resolved (time-step) and difficult to find at longer time scales.

The inherent uncertainty implicit in proxy records imposes limitations on the modelled output in terms of chronology and precision. Essentially modelled output is constrained by the land use and climate time series each of which are underpinned by independently derived chronologies, each subject to limitations and error.

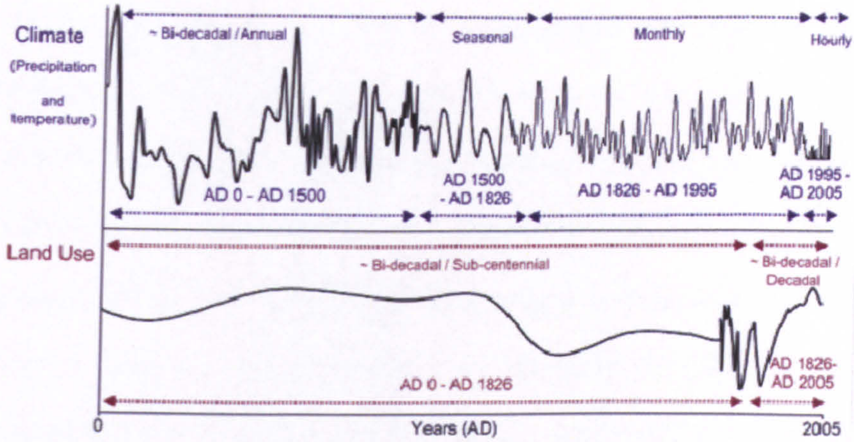

Figure 5.2 *Schematic diagram to indicate approximate resolution of available input data for Petit lac d'Annecy*

Temporal and spatial output from the various simulations completed using the aforementioned time series will be compared with sediment proxies from the Petit lac and the geomorphology of the catchment which are equally constrained in terms of chronology and resolution. However the pattern and sequence of change in the simulated output should broadly correspond to indications of change in the lake sediment record. It is difficult to improve on this approach, because there are few alternatives that exist given the constraint of using a model that requires hourly inputs, thus all the proxy records have been adjusted to hourly increments (See section 5.4).

5.3.3 Inputs and Outputs of CAESAR

The two drivers of CAESAR simulations are an hourly precipitation time series and a time series value that is representative of land use, both required as single column ASCII text files. Simulations are also conditioned by two gridded datasets, a digital elevation model (DEM) and a representation of catchment soil/sediment depth in the form of a second DEM at the approximate elevation of the bedrock surface; together these reflect the topography and sediment distribution within the catchment (Fig.5.3,

Fig.5.4). The outputs of CAESAR are an hourly time series of water discharge, sediment discharge for nine user-defined grain size fractions (output file = catchment.dat) and also total sediment discharge. In addition, CAESAR simulations can be configured to output gridded datasets at user defined intervals to allow spatial patterns within the catchment to be examined. These include a modified DEM (output file = elev.txt) at the end of the model run, a gridded dataset showing the spatial pattern of erosion and deposition i.e. the net difference in elevation (output file = elevdiff.txt), a gridded dataset of stratigraphy and grain size distribution (output file = grain.txt) and a gridded dataset of water depth within a cell (output file = waterdepth.txt). In addition to the input files, it is necessary to define a number of parameters e.g. soil creep rate, soil erosion rate, simulation length and minimum water discharge through a cell for each simulation. The input files, their creation and setup are described in more detail below.

5.4 Creation of model input files

5.4.1 Digital elevation model

For the Petit lac sub-catchments, a 50 m x 50 m grid cell DEM has been used (source: Institut Géographique National, France). The nature of cellular automata models demands high numbers of calculations per grid square, and therefore coarse resolution DEMs are often necessary for modelling longer time scales. This compromise results in the loss of the finer detail in terms of geomorphological change (e.g. Coulthard et al., 2005). There is essentially a trade off between run time and resolution, but a 50 m grid resolution is sufficient to model catchment sediment

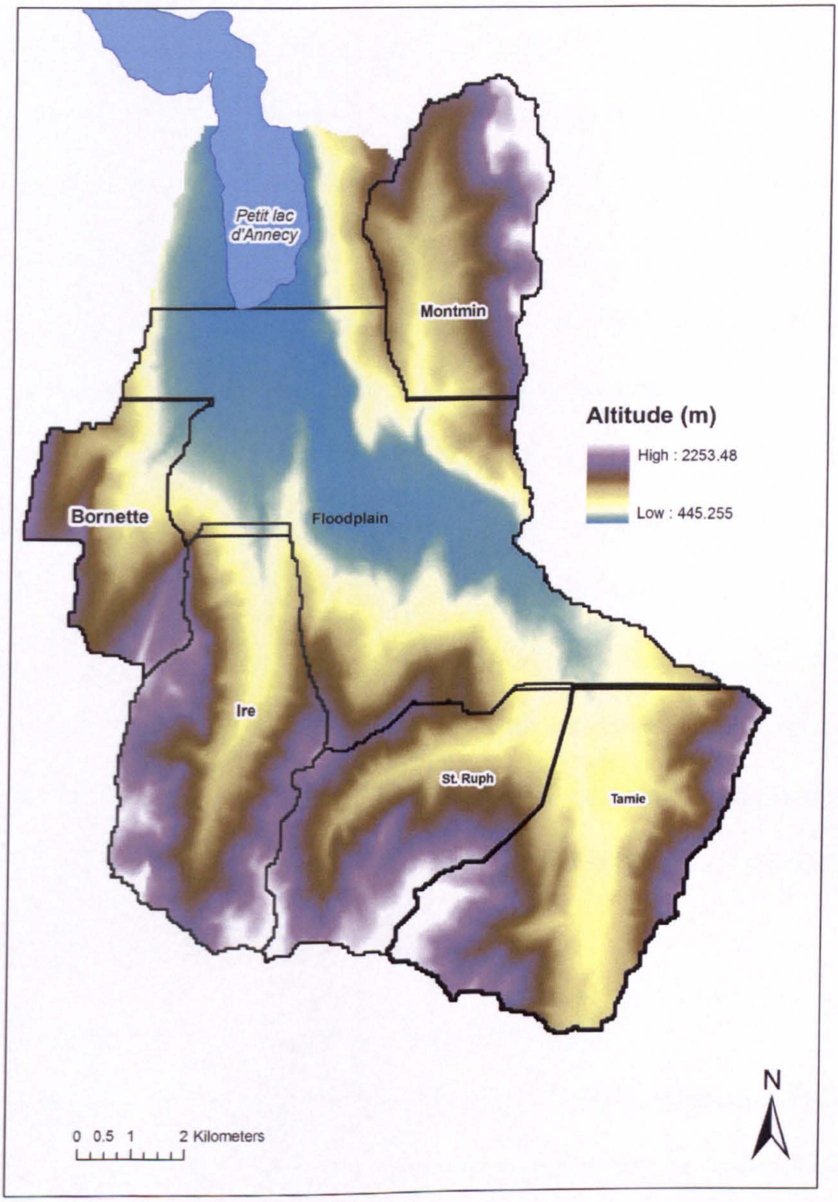

Figure 5.3 *Digital elevation model of the Petit lac d'Annecy. Black lines denote sub-catchment boundaries.*

output for comparison with the lake sediment records and this resolution was therefore used throughout this thesis.

The DEM was processed using ESRI ArcGIS Hydro extension tools (http://www.esri.com/) which are designed to assist with modelling water resources and surface water flows. The processing included the removal of sinks (hollows) in the DEM (Coulthard et al., 2002) and the use of flow routing algorithms to identify the channel position and distribution of sub-catchments. For CAESAR simulations the catchment has been sub-divided according to the hydrology to reduce model simulation time as there is a proportional relationship between run-time and the size of the DEM therefore smaller DEMs (less cells) result in shorter run time. In addition, simulating individual sub-catchments allows comparison of different basin geometries (Fig.5.3) and allows the variation of land use history between sub-catchments. At present this is the only method for varying land-use on a spatial basis in CAESAR simulations. At a finer resolution, the sub-catchment DEMs could be split further according to generalised patterns of land use to assess which slopes and which areas of particular sub-catchments are most vulnerable to instability and greatest geomorphological change.

5.4.2 Soil/Sediment depth model and grain size distribution

The distributions of unconsolidated sediment and soils are incorporated through the use of a second DEM, which imitates the elevation of the bedrock surface. Essentially the depth between the elevation surface and the bedrock surface reflects the amount of sediment that is readily available for erosion within CAESAR simulations. Depth to the bedrock was estimated for the whole catchment by field-based assessment and using an empirical relationship between sediment thickness recorded in the field, slope angle and elevation (Figure 5.4, Figure 7.1a,b). There is an additional spin up time of 60 years added to the start of each simulation; this allows initial conditions to reach equilibrium and enables the channel to establish a protective armour layer (c.f. Coulthard et al., 2002). The first 60 years of spin-up time are not included in the results.

Ranges of slope angles were identified through the DEM and each range was thereby assigned a soil/sediment depth. This process produced a spatially variable sediment depth gridded dataset ranging from 0.1 to 20 metres, with less soil/sediment on the summits and steeper cliffs and slopes. More soil/sediment was available at low gradients (Fig.5.4). All surfaces above 1400 m were assigned a soil/sediment depth of 1 meter to ensure the summit regions had a realistic depth of sediment after the run up period, during which time much of the sediment was removed from steep slopes, therefore bare rock zones would have minimal sediment depth. Areas above 1400m in the St Ruph and Bornette sub-catchments were assigned soil depths based on field evidence (c.10m), as the fill in these areas was considerably greater than 1m. Comparison with field data shows the estimated sediment/soil thickness to be broadly realistic.

CAESAR requires the user to input nine grain size fractions of typical sizes found in catchment streams, their fall velocity, their relative proportions and whether or not they should be treated as suspended sediment or bedload. Transport of bedload is distributed proportionally to local bedslope of the cells lower than cell *n* and deposition of bedload can be re-entrained each iteration. Suspended load is transported slightly differently as it is dependent on flow velocity and has the

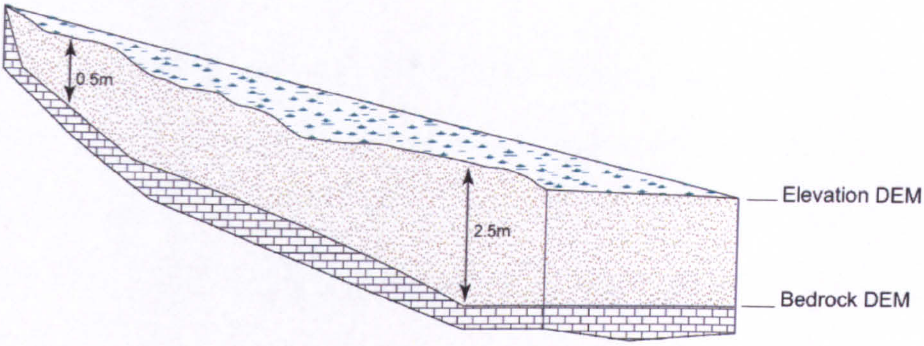

Figure 5.4 *Example of soil-sediment layer. This method allows steep slopes to have a shallower fill (e.g. 0.5 m) and lower gradients to have a deeper fill (e.g. 2.5 m).*

potential to move to cells where the bed elevation is less than the water depth in cell *n*. Deposition of suspended load is derived from fall velocities and concentrations for each grain size fraction. Sediment is stored in each cell in a series of vertical layers (*grain.txt*) (Van de Wiel et al., 2007) of which the active layer is affected by erosion (sediment moved, therefore erodes the active layer) and deposition (sediment added, contributes material to the active layer). All of these layers down to the bedrock (not erodible) are available for erosion.

Sediment distribution was subdivided into grain size fractions of: 63 µm, 250 µm, 500 µm, 1000 µm, 2000 µm, 4000 µm, 8000 µm, 32000 µm, 128000 µm. The finest fraction of clays and silts (<62.5 µm) were handled as suspended sediment and used for comparison with proxy data from the lake sediment record. This grain size fraction was selected for this research as CAESAR model does not include cohesion of clays at present. Distribution of grain size is uniform at the start of a simulation and during the run up period the distribution of grain size across the catchment will reach equilibrium (Coulthard et al., 2002). The grain size data were taken from Foster (2003) who analysed sediment from the Petit lac d'Annecy catchment using a Coulter machine.

5.4.3 Precipitation

The precipitation input file is an ASCII text file which consists of a single column of precipitation expressed in millimetres per hour for the duration of the model run. This file is representative of past hourly precipitation for the Petit lac catchment and was applied to each cell within the DEM homogeneously, regardless of location or altitude.

5.4.3.1 Precipitation series for the period AD 1826-2005

Hourly precipitation data are available for several sites close to the Petit lac d'Annecy catchment (Table 5.2). In terms of altitudinally comparable sites, Meythet (~28km from the Petit lac, Fig.5.5) appears most similar to the Petit lac d'Annecy, however the quality of data and number of missing data points created difficulties. Hence, the more complete Albertville (Fig.5.5) record has been used: its proximity, comparable altitude and similar precipitation pattern which hopefully captures similar localised precipitation events to the Petit lac catchment suggests that it is the most appropriate time-series to use. It is preferable to the La Féclaz record which is significantly wetter (Table 5.2) due to altitudinal differences.

The Albertville hourly series (1991-2005) and a linearly regressed Annecy-Geneva modelled monthly series (1826-2000) (Nicholson & Thompson, 2000) comprise the most suitable rainfall records close to the Petit Lac d'Annecy catchment, particularly as the Annecy record has 30 years of data missing between the two World Wars. Longer hourly precipitation series were created by repeating the 5 year (1995-1999) high resolution (Albertville) precipitation record back to 1826 and then adjusting the

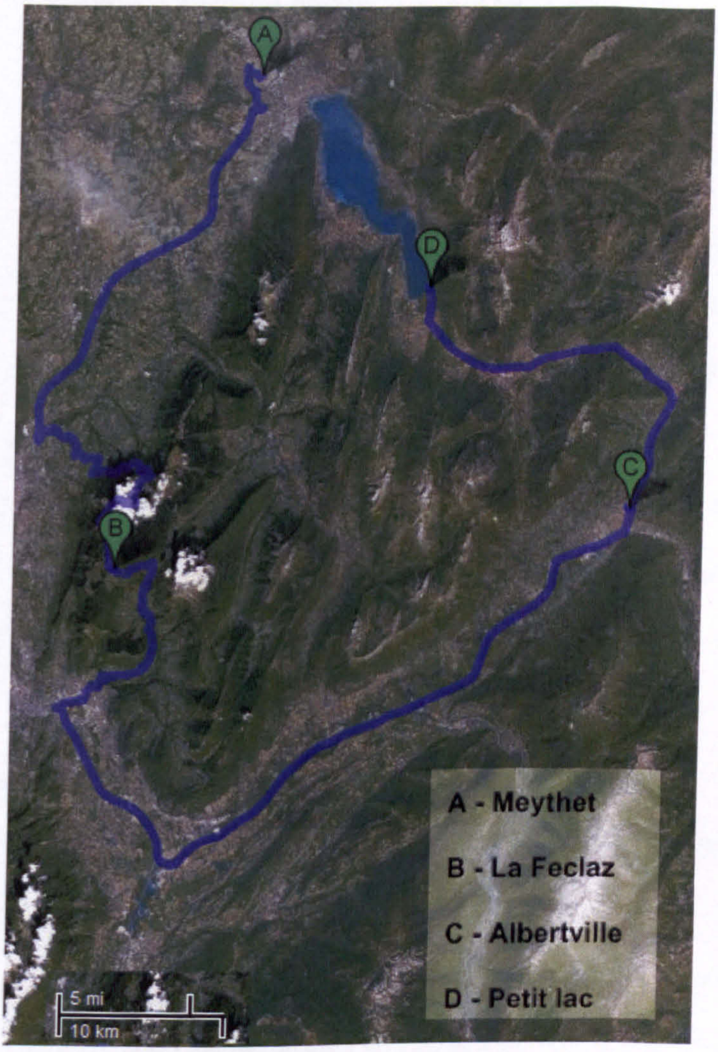

Figure 5.5 *Location of available hourly precipitation data close to the Petit lac d'Annecy catchment (e.g. D) © Google Images 2009*

Location	Distance from Petit lac (Km)	Altitude (m asl)
Meythet	~20	465
Albertville	~28	352
La Féclaz	~46	~1200

Table 5.2 *– Proximity of data sources to the Petit lac d'Annecy catchment*

averages according to the integrated Annecy-Geneva monthly precipitation series (Nicholson and Thompson 2000) on a cyclical basis. Whilst monthly totals will reflect the climate history for the last 180 years, the frequency and relative magnitude of stochastic events (rainfall) will repeat on cycles of five years which is a limitation of this approach. Some consideration was given to using weather generators to produce an arbitrary hourly precipitation series, but upon reflection, this method may not have replicated the local hourly rainfall pattern for the pre-Alpine region. Weather generators often produce daily data that have a lower inter-annual variability and they also require a large amount of observed data to be sufficiently parameterised (Semenov et al., 1998).

5.4.3.2 Precipitation series for the period AD 1500 to AD 2000

Gridded meteorological data are available for the Petit lac d'Annecy catchment from the NOAA palaeoclimatology data programme (http://www.ncdc.noaa.gov/oa/ncdc.html). Pauling et al. (2006) created a gridded seasonal precipitation reconstruction for European land areas (30°W to 40°E/30-71°N; given on a 0.5° x 0.5° resolved grid) using long instrumental precipitation series, precipitation indices derived from documentary evidence and proxy records from environments that are sensitive to precipitation signals (e.g. tree-ring chronologies, ice cores, corals, speleothem). These data are available seasonally (DJF, MAM, JJA, SON) between AD 1500 and 2000 and provide a meso-resolution data set beyond the historical records available in the Petit lac d'Annecy catchment. A similar methodology to section 5.4.3.1 was used to create an hourly precipitation record from the seasonal gridded data for the period AD 1500 – 2000 (Pauling et al.2006). Again the 5 year hourly series (1995-1999) was used, repeated for 500 years and scaled against the seasonal (DJF, MAM, JJA, SON) totals for each year. After calibrating the records for the period of 1826-2000, during which both monthly data from the Petit lac d'Annecy catchment (average 1246 mm $^{yr-1}$) and seasonal data from the 500 year Pauling precipitation record (average 1363 mm $^{yr-1}$) are available, the Pauling data has been adjusted to the known averages of the Petit lac d'Annecy catchment.

5.4.3.3 Precipitation series for the period AD 0 – 1500

To extend the hourly precipitation record further back into the Holocene (approximately 2000 years), a proxy for precipitation is needed from palaeorecords. There are many proxy records of assumed palaeoprecipitation for this region, including nearby Jura mountain bog water level studies (Mitchell et al., 2001), Magny's (2004) lake water level data from the Northern Alps, the Upper Rhone hydrological index (Bravard et al., 1992) and peat bog water levels from central European site (Charman et al., unpublished).

Here, a bog surface wetness record from a lowland raised site in Germany (Charman et al. unpublished data) has been used to infer changes in wetness and is thought to represent precipitation changes with minimal human influence (with the exception of AD 1800-2000). The record is taken from one of the sites accessed during the EU-funded ACROTELM project and the record covers the period between AD 2000 to 1252 BC though only the latter part of the record is used here (0AD-1500AD). The upper part of the available bog surface wetness record (1500AD – 1800AD) is compared to the data available from Pauling et al., (2006) and precipitation patterns in both appear to be regionally coherent, giving confidence that the record is suitable for modelling precipitation in the Western Alps. The data from the most recent part of the core (AD 1800 -2005) were discarded owing to human intervention and drainage which affected the recent surface wetness history (AD 1800 -2005) and would therefore be unlikely to reflect typical wetness patterns. In addition higher resolution data (seasonal; Pauling et al., 2006) were already available for AD 1500-2000 and were joined to the record derived from the bog surface wetness data to give a complete 2000-year long set of data. Using a similar methodology to section 5.4.3.1 an hourly precipitation record for the period AD 0 – AD 1500 was created by adjusting the 5 year hourly series (1995-1999) against the water table data (Charman et al., unpublished) and scaling the values to known averages for the Petit lac d'Annecy catchment. The series is regarded as broadly reflective of the bi-decadal pattern of changes in precipitation over 1500 years. This method is similar to that outlined in Coulthard et al., 2002.

Of the other possible archives available, many were deemed unsuitable for various reasons. Mitchell et al., (2001) have extracted bog surface wetness data from a peat bog in the neighbouring Jura mountain range. These data could have provided a fairly local signal of surface wetness, however unlike the bog surface wetness record from Germany that has been used in this research (Charman et al., unpublished), there are mixed signals in the peat record with land use playing a significant control over bog wetness, in addition to which the site had been affected by migration of the Orba river and many of the changes in wetness have been explained by either land use change or local mire microtopography. Therefore, given that this research requires a palaeorecord that reflects likely changes in precipitation, this record was deemed unsuitable.

Magny et al., (2004) produced a synthesis of Holocene lake levels using 26 lakes from the Jura Mountains, Northern French pre-Alps and the Swiss Plateau. The data identify phases of higher and lower lake levels that have been interpreted in terms of changes in climate (Magny et al., 2004.). However lake level data are constrained by a number of issues; firstly lake water levels reflect both precipitation and temperature (evaporation) in essence effective precipitation, and separation of these contributing factors is problematical. Secondly, the lakes included in the database are not exclusively closed basins, and the distinction between lake level data from open and closed basins are not explicitly defined. Finally, Lac d'Annecy and other local lakes are components of this database, and the reasoning could be regarded as circular if these data were used to drive CAESAR simulations.

5.4.4 Temperature

Temperature data are not a required input for the conventional CAESAR model simulations (e.g. Coulthard et al., 2002, Van de Wiel et al., 2007), but were used in model development to account for snowfall and water storage during winter months. Thus a programme (see 5.5.2) has been developed to moderate precipitation using the annual pattern of temperature to simulate snow storage, and consequently, hourly temperature data were required. Hourly temperature data are available for Meythet (Table 5.2) from 1995-2006. The same 5 year period as the hourly precipitation data (1995-1999) has been used as a template to produce a longer hourly temperature time series in a similar manner to the hourly precipitation records (section 5.4.3). Monthly average temperatures (1995-1999) for both Meythet and the Petit lac d'Annecy catchment have been used to scale the Meythet hourly series to reflect conditions in the Petit lac d'Annecy catchment.

5.4.4.1 Temperature series AD 1826-2005

A longer hourly temperature series was created by repeating the 5 year (1995-1999) high resolution (Meythet) temperature record back to AD 1826 and adjusting it by monthly averages for the Petit lac d'Annecy catchment on a cyclical basis. This again produces repeating 5 year event cyclicity but reflects the broad temporal pattern of changing temperature for the period.

5.4.4.2 Temperature AD 1500-2000

For the period before AD 1826, gridded meteorological data are again available from the NOAA palaeoclimatology data programme (ref. http://www.ncdc.noaa.gov/oa/ncdc.html). Luterbacher et al. (2004) have used various proxy records of temperature (e.g. ice cores, tree rings, speleothems, varved sediments, and subsurface temperature profiles obtained from borehole measurements) on multi-proxy networks which amalgamate natural proxy indicators with climate information obtained from early instrumental and documentary evidence to provide a gridded precipitation series. Temperature data were extracted for the same co-ordinates as the precipitation data (section 5.4.3.2). The Luterbacher data from 1826AD-2000AD exhibit similar trends to the Petit lac d'Annecy record available for the same time period. Similarly to previous methods, the 5 year hourly Petit lac d'Annecy temperature series was repeated and scaled to seasonal averages from the Luterbacher data.

5.4.4.3 Temperature 0AD-2000AD

To allow a 2000 year model run to take place, proxy temperature data are necessary to obtain a long climate series. Moberg et al. (2005) reconstructed millennial-scale climate variability by combining tree-ring data and other low resolution proxy data using a wavelet transformation technique. This dataset offers a best estimate of Northern Hemisphere temperature for 2000 years at annual resolution. As with previous methods, the 5 year hourly Meythet temperature data was repeated and scaled to annual averages from the Moberg record. A comparison for AD1500-AD2000 between trends in the Moberg and Luterbacher records were analysed and similar patterns found.

5.5 Land use, the *m* parameter

In CAESAR, land use changes are represented by the *m* parameter, which operates within the hydrological sub-routine, and has been adapted here (version 5.5 onwards) to allow *m* to be varied on an hourly basis paralleling the climate series. This parameter is a modification of the vegetation parameter used in TOPMODEL (Beven, 1997). Essentially, *m* values are derived to mimic the impact of catchment vegetation cover on the recession limb of the flood hydrograph. A high *m* value such as 0.013 slows the rate of decline of the recession limb of the hydrograph (Figure 5.6), and reduces the transmissivity within the soil, therefore imitating the impact that a dense vegetation cover would have on the catchment. A lower *m* value such as 0.005 increases transmissivity through soil as there is a quicker decline in soil moisture deficit. This imitates the impact of a more sparse vegetation cover. *M* values account for all the factors associated with water movement and water storage with regards to vegetation. In CAESAR *m* values are applied uniformly across the catchment.

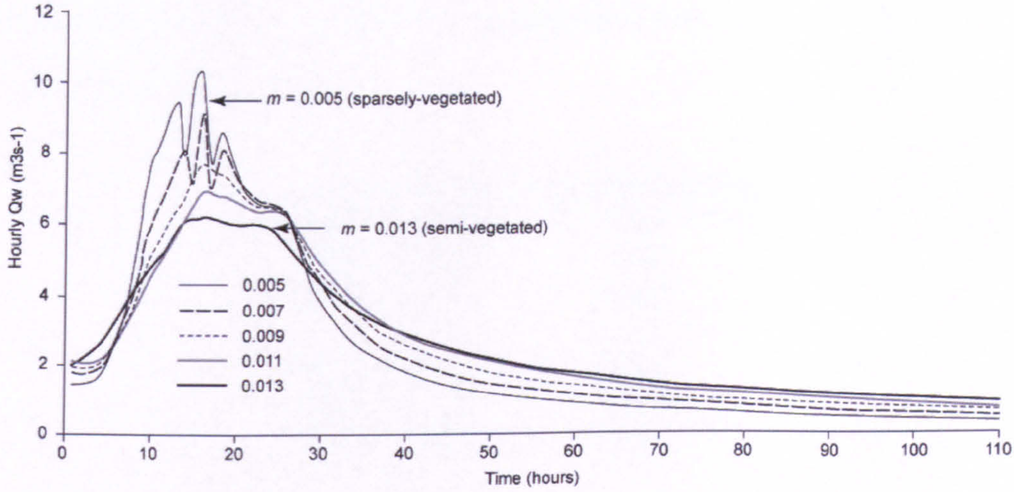

***Figure 5.6** Modelled discharge data under different m conditions for the Ire. The most sparsely vegetated landscape represented by the lowest m values (e.g. 0.005) has a flashy hydrograph with peak discharges of c. 10 m3s-1. Under identical conditions, with a more densely vegetated landscape represented by a higher m value (e.g. 0.013), the flood hydrograph is less flashy with peak discharges of c. 6 m3s-1.*

5.5.1 Preliminary tests to calibrate m value to contemporary land use

The inputs needed to run CAESAR are simple and require little parameterisation or empirical data when compared to other hydrological models (e.g. Mike-SHE: Abbot et al., 1986, Refsgaard, 1997). However, given the role the *m* value plays in governing the shape of the flood hydrograph it is important to calibrate modelled discharge against recorded discharge in order to obtain an appropriate *m* value for the present day land use. A series of CAESAR simulations were completed for the Ire catchment using a precipitation time series for 1999 but varying the *m* value across the range 0.002-0.02 (Coulthard pers. comm.; Beven, 1997). The shape and magnitude of flood hydrographs along with flood frequency distributions in the modelled water discharge for 1999 were then compared to instrumental discharge data for 1999 from the Ire river. As sediment tends to move under the largest flood magnitudes, both the r^2 values of *all* recorded flows *and* of recorded flows over 2 m3s-1 have been calculated.

m value	All flows	Flows over 2 cumecs
0.007	0.988	0.866
0.008	0.952	0.923
0.009	0.922	0.924
0.01	0.886	0.96
0.014	0.74	0.952
0.015	0.71	0.879
0.016	0.684	0.904

Table 5.3 *r^2 values for all simulated flows and simulated flows above 2 m3s-1 for a range of simulations with the same inputs and varying m values (land use).*

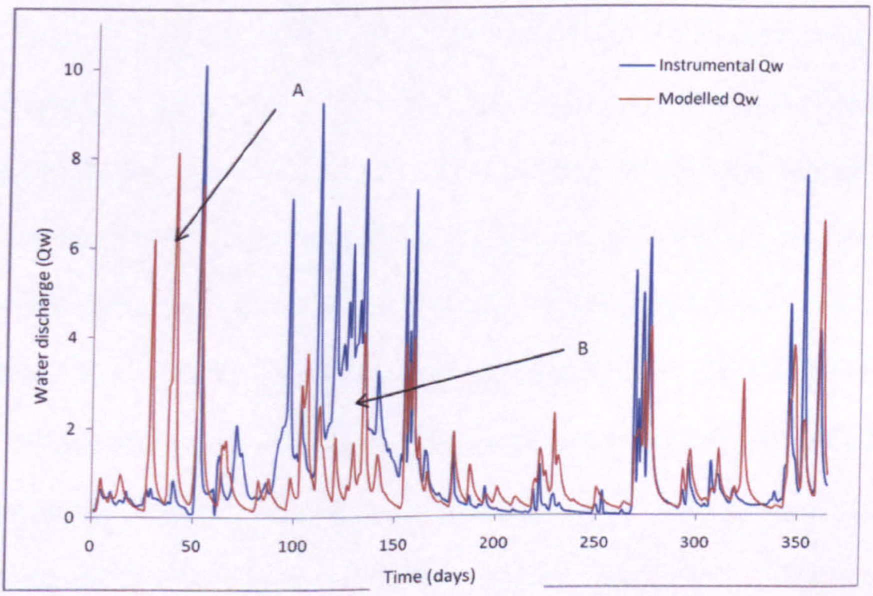

Figure 5.7 *Modelled discharge for 1999 from a simulation in the Ire sub-catchment compared with instrumental discharge data for 1999 from the Ire sub-catchment. A. shows over-prediction of discharge in winter months due to lack of snow accumulation and B. shows under-prediction in snow melt months due to lack of snow release.*

Correlation and analysis of variation (r^2) (Table 5.3), suggests that the *m* value should be 0.010 for present day conditions. However, the time series for daily instrumental and modelled discharges (m=0.010) for 1999 are far from comparable (Fig. 5.7). The lack of correlation becomes apparent because the simulations do not incorporate snow accumulation/melt, and so the lack of snow storage creates exaggerated runoff peaks during winter months (Fig. 5.7) and under-predicts runoff in the spring (Fig.5.7) due to the lack of snow melt waters. Clearly given the importance of snow in the annual hydrological cycle of the Petit lac d'Annecy catchment, the creation of snow melt sub-model was critical.

5.5.2 Incorporating snow accumulation/melt

Snow storage and melt have been incorporated by creating a simple snow mass-balance model (SMB) to moderate the precipitation series into a snow-adjusted series using available temperature data. The programme uses hourly precipitation and temperature data, and employs user defined values to set a freeze threshold, freeze percentage, melt threshold and melt rate. The annual snow storage cycle produced was similar to empirical data recorded from the Alps (e.g. Beniston et al., 2003), with winter storage (November – March) and a mid-March snow melt pulse. Limitations to this approach stem from discrepancies between the Albertville time series and actual events in the Petit lac catchment. Given the intention to apply the same approach to longer time series, these small discrepancies are not deemed major limitations. More critically the precipitation series is applied on a uniform basis across the DEM grid regardless of variations in altitude. In future research this may be revisited, because there is scope for a snow sub-model that incorporates the impacts of altitude (lapse rates) using the DEM, however the necessary programming was beyond the scope of this thesis.

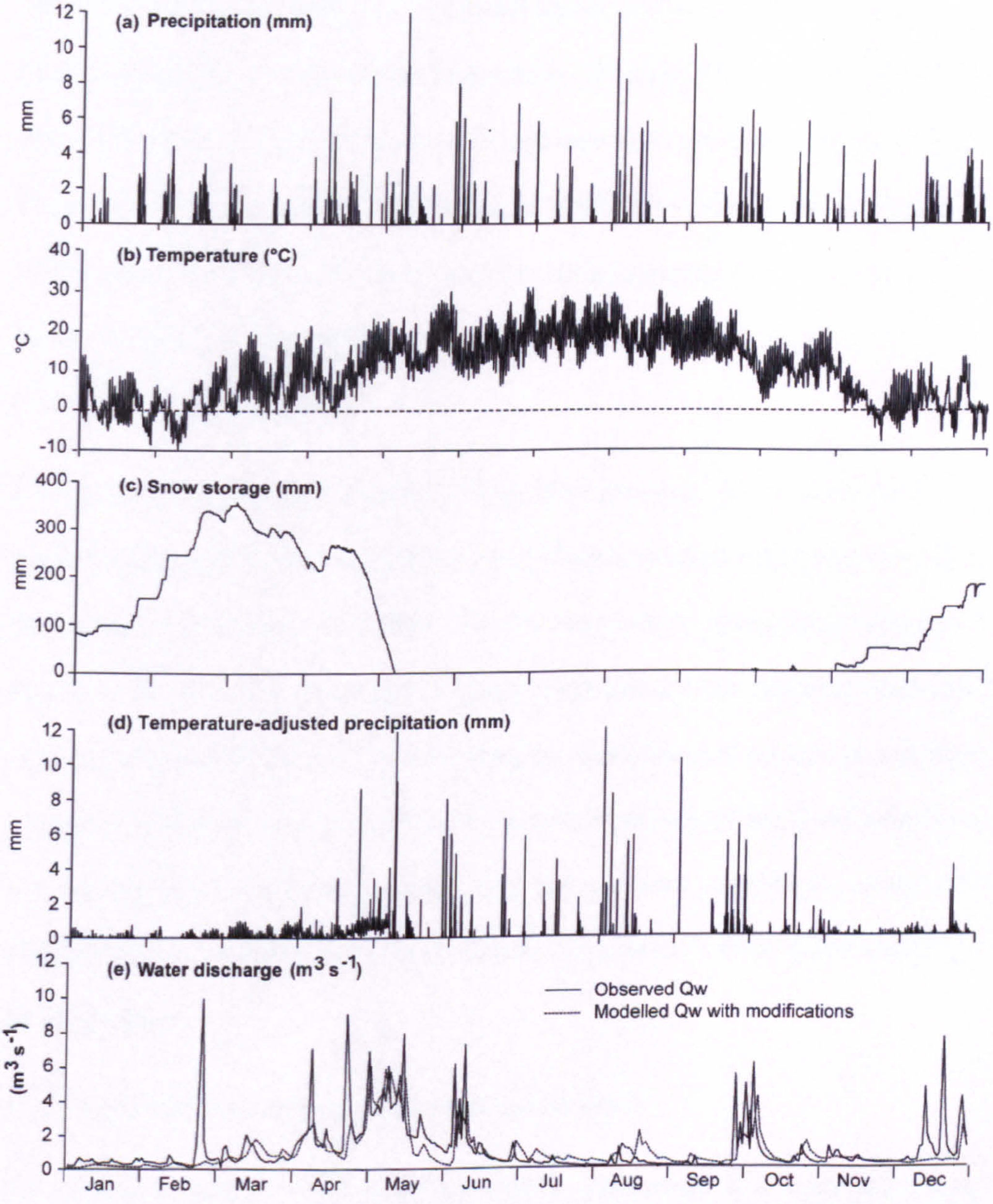

Figure 5.8 *Example of the precipitation pre-processing from SMB model for 1999. (a) Original precipitation data, (b) Temperature data, (c) Snow storage taking place during processing, (d) Temperature modified precipitation data, (e) Modelled daily average discharge data (snow-adjusted precipitation) compared with observed average daily instrumental discharge from Ire (1999).*

The 'snow-melt' programme can operate for any uniform interval time-step, here hourly, and uses two time series: precipitation (Fig.5.8a) and temperature (Fig.5.8b).

There are four user defined values; a freeze threshold (°C), freeze percentage (% hr-1), melt threshold (°C) and melt rate (mm hr-1). The output is a modified rainfall file (Fig.5.8d) adjusted using temperature to represent seasonal snow storage and melt. This modified rainfall series has been used in CAESAR to drive simulations in place of the original precipitation file. In addition, the programme produces a measure of snow storage (Fig.5.8c), which is used to confirm that snow is being stored during appropriate times of the year.

Correlation and analysis of variance (r^2) for SMB modelled and instrumental (1999) water discharges (Fig.5.8e) suggest the processing has worked well with the r^2 values improving for all simulations (Table 5.5). A comparison of timing and magnitude of floods in for the modelled (with original precipitation) and recorded discharges showed poor matches (Fig.5.8), but by using the snow-adjusted data the two datasets are more comparable in terms of the timing and magnitude of the flows (Fig.5.8e). The annual pattern includes a snow melt pulse around mid-March, along with substantial snow storage in the winter months (Dec-Feb), both of which are driven by temperature.

5.5.3 Testing the snow mass balance (SMB) programme

The freeze threshold, melt threshold, freeze percentage and melt rate were determined in part by a trial and error approach to achieve the best comparison with recorded water discharge. The key features that were essential to be capture in the simulations using the adjusted precipitation series were:

- storage of precipitation in winter months
- a sustained peak of increased discharge in snow melt months
- a strong correlation between recorded and modelled discharges

To assess the performance of the SMB programme a series of thresholds and rates were tested (Table 5.4) with a selection of these results presented and compared with

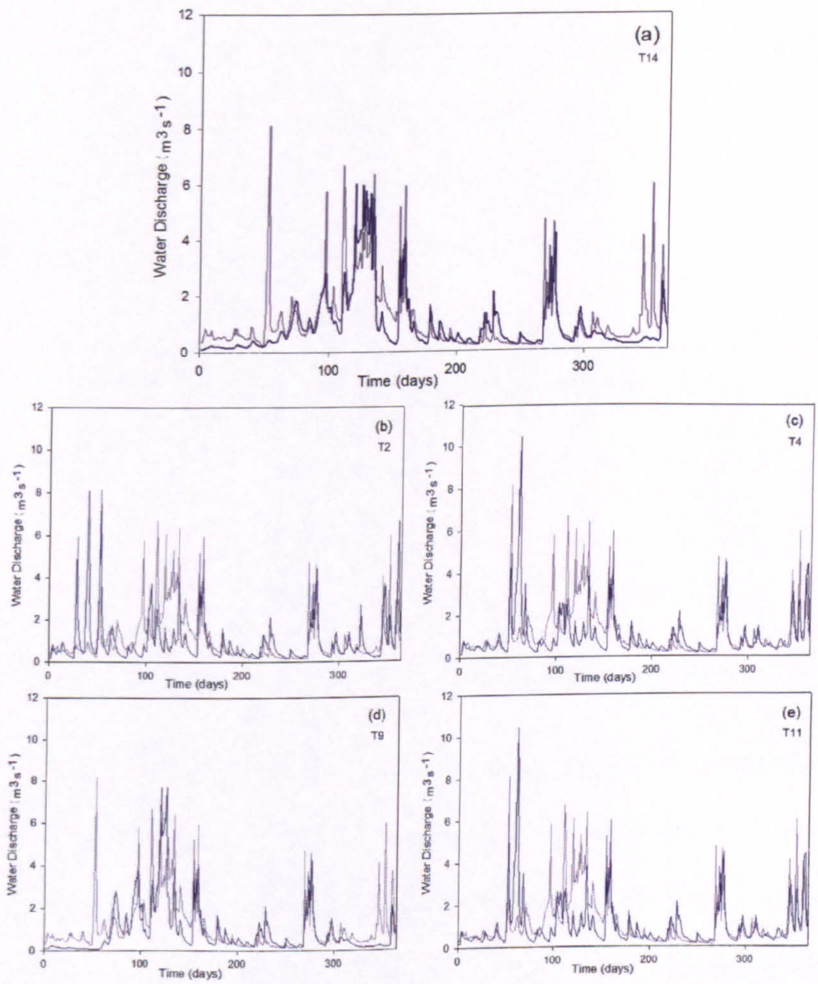

Figure 5.9 *(refer to Table 5.4 for values) One year of modelled water discharge(blue) from the Ire sub-catchment for (a) Test 14, (b) Test 2, (c) Test 4, (d) Test 9, (e) Test 11 compared to one year (1999) of observed data for the Ire sub-catchment (black)*

recorded discharge data for the Ire (1999). The aim was to determine thresholds and rates that are reflective of the annual trend in discharge for 1999 before applying the SMB programme at longer time scales. The actual values used for temperature are less important as they are reflective of temperatures at low altitudes (e.g. at Meythet c.465m asl) rather than reflecting the Petit lac d'Annecy catchment averages. Rather than adjust the temperature series to reflect this, here higher melt and store thresholds were used and then calibrated against recorded water discharges.

T2 (Fig. 5.9b) shows that if the melt threshold is set too low, then increased release of 'snow' will occur too early e.g. January rather than during the snowmelt season which is typically March-April. By increasing the melt threshold in T4, there is a delayed onset of melt. In addition, for T2 (Fig. 5.9b), T4 (Fig. 5.9c), T9 (Fig. 5.9d)

Test No.	Freeze Threshold	Melt Threshold	Freeze Percentage	Melt Rate
2	7	4	75	0.65
4	8	5	75	0.15
9	9	6	95	0.10
11	10	6	80	0.15
14	9	6	85	0.07

Table 5.4 *Selection of the values for the tests of the SMB programme*

and T11 (Fig. 5.9e) the melt rate was too high, and much of the stored volume of 'snow' was be released quickly and later months were deficient of 'snow'. T9 (Fig. 5.9d) shows that if freeze percentages were too high e.g. 95%, the early months of the year exhibited insufficient discharge as too much 'snow' was stored. The amount of 'snow' stored in T9 (Fig. 5.9d) suggests that a relatively high percentage needs to be stored in order to sustain the melt discharge peak in March-April. T11 (Fig 5.9e) shows that if freeze threshold was set too high then no or less precipitation was stored, resulting in flashy discharge peaks throughout winter months. T14 (Fig.5.9a) has produced a good match between the observed and modelled discharge data, replicating the reduced discharges through to day ~60 and producing a sustained increase discharge from day ~70 to ~170. Further refinements of this process took place when calibrating for the influence of the land use parameter *'m'* on the flood hydrograph. The values used for test 14 are regarded as broadly representative of the annual pattern of snow accumulation and snow melt and were applied to the longer records.

5.5.4 Revised calibration of the land-use parameter

A range of *m* values were tested using the processed snow-adjusted precipitation series for 1999. The flood frequency distributions of recorded water discharges and the simulated water discharges were compared to assess the best match for contemporary land use, *'m'* value (Fig.5.10). Discrepancies between the two discharge data sets were inevitable as the precipitation series data used in the simulations was derived from Albertville and may include local precipitation events that did not occur in the Ire catchment. The *m* value that returned the closest match (r^2 value) for all flows and peak flows (above 2 m3s-1) was 0.011 (Table 5.5), and on this basis for the Ire *m* values of 0.011 are regarded here as best representing the influence that the contemporary land use has on the flood hydrograph.

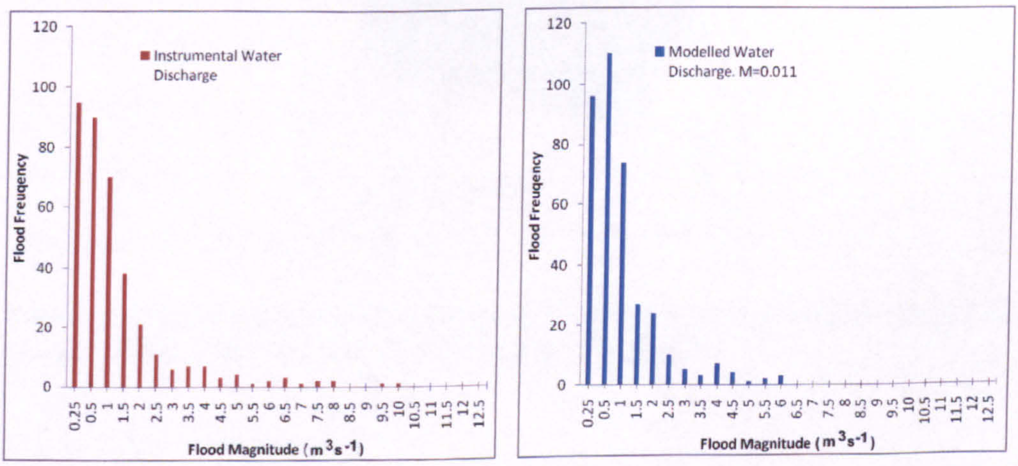

Figure 5.10 *Flood frequency distribution of instrumental and modelled water discharge.*

m value	All flows	Flows over 2 cumecs
0.006	0.95	0.91
0.007	0.972	0.889
0.008	0.982	0.899
0.009	0.982	0.903
0.01	0.969	0.891
0.011	0.981	0.922
0.012	0.978	0.911
0.013	0.97	0.866
0.014	0.957	0.829
0.015	0.961	0.82
0.016	0.951	0.847
0.017	0.947	0.89

***Table 5.5** - r^2 values for all simulated flows and simulated flows above 2 m3s-1 for a range of simulations with the same inputs and varying m values (land use) using temperature-adjusted precipitation input data.*

5.5.5 Land use 1826-2005

***Figure 5.11** Land use remotely sensed from Google imagery aerial photographs (© Google, 2008). Black lines denote sub-catchment boundaries.*

It is useful to know what the contemporary *m* value equates to in terms of present day land use so that values can be scaled and so that the record can be extended back into the past. Aerial photography (©Google Imagery, 2008) suggests an *m* value of 0.011 equates to a contemporary forest cover for the Ire (Fig. 5.11 and Table 5.6), of approximately 70%. The other sub-catchments were then scaled according to contemporary land use and the relationship between *m* and forest cover (Fig.5.12). For the past, there are relatively few aerial photographs available, which suggests that other sources of data are required for a longer history of changes in land use.

Sub-catchment	Open Land %	Woodland %
Ire	29.79	70.21
Tamie	28.06	71.94
St Ruph	20.68	79.32
Montmin	46.32	53.68
Bornette	22.86	77.14
Floodplain	36.60	63.40
Total Catchment	*31.87*	*68.13*

Table 5.6 *Percentage of open land and woodland in each sub-catchment*

Forest cover inventories are available for several communes that lie within the Petit lac d'Annecy catchment (Crook et al., 2002). To correct for differences in boundaries between communes and sub-catchments, the percentage of each commune lying within a sub-catchment boundary was calculated and used to calculate an average forest cover which was weighted by commune area for each sub-catchment. Further discrepancies occur owing to differences between the forest inventories (Crook et al., 2002) and the forest cover discerned from aerial photographs. The forest inventory records for 1999 (Table 5.7) show significantly less forest cover for 1999 than shown in the aerial photographs and observations made in recent field visits (1999-2007).

Sub-catchment	1735	1811	1824	1836	1840	1882	1892	1909	1913	1936	1950	1976	1999	Aerial Photos
Ire	51.763	36.028	30.979	19.056	19.171	39.839	41.657	47.104	46.241	49.012	49.012	54.813	49.125	69.589
Tamie	30.797	24.981	21.799	30.435	21.799	31.293	36.347	33.831	33.855	36.215	36.314	33.896	38.563	57.260
St Ruph	39.933	31.974	28.662	38.733	28.662	38.657	45.736	43.779	43.868	46.540	46.899	44.154	49.455	71.890
Montmin	26.197	18.266	8.631	14.040	8.631	21.857	12.334	26.003	26.000	26.000	36.335	35.326	35.547	57.001
Bornette	35.555	29.722	17.838	13.423	15.910	14.277	18.099	25.330	20.274	27.195	27.195	32.598	27.642	55.090
Floodplain	49.018	36.286	37.939	37.127	36.980	33.408	41.382	50.543	50.328	51.866	53.849	58.424	53.163	71.542

***Table 5.7** Percentage of forest cover per sub-catchment by area weighted mean from forest inventory records (Crook et al., 2002).*

The forest inventories are records of declared land use and therefore records may have been omitted or land may not have been declared by land owners (Crook et al., 2002). Consequently the forest inventory records have been adjusted to the forest cover that was remotely sensed from aerial photographs (e.g. Fig.5.11). The revised percentages of forest cover for all sub-catchments for all time periods are shown in table 5.8.

Sub-catchment	1735	1811	1824	1836	1840	1882	1892	1909	1913	1936	1950	1976	1999	Aerial Photos
Ire	72.227	56.492	51.443	39.520	39.635	60.303	62.121	67.568	66.705	69.476	69.476	75.277	69.589	69.589
Tamie	49.493	43.678	40.495	49.132	40.495	49.990	55.044	52.528	52.551	54.911	55.011	52.593	57.260	57.260
St Ruph	62.369	54.410	51.097	61.168	51.097	61.093	68.171	66.214	66.304	68.976	69.335	66.590	71.890	71.890
Montmin	47.651	39.720	30.085	35.493	30.085	43.311	33.788	47.457	47.454	47.454	57.789	56.779	57.001	57.001
Bornette	63.003	57.170	45.285	40.870	43.358	41.725	45.547	52.777	47.722	54.642	54.642	60.046	55.090	55.090
Floodplain	67.397	54.666	56.319	55.507	55.360	51.788	59.762	68.922	68.707	70.246	72.228	76.804	71.542	71.542

***Table 5.8** Revised percentage of forest cover for each sub-catchment areas after calibration with present day (2008) aerial photographs.*

To convert percentages of forest cover into an *m* value, the *m* values were scaled on a linear basis against percentage of forest cover (Fig.5.12). Clearly it is possible to postulate a wide variety of linear and non-linear models to summarise the relationship between forest-cover and the *m* parameter, but in the absence of empirical data to substantiate the nature of this relationship for past land use a linear relationship is used here (Figure 5.12). This approach is similar to that applied in previous simulations using CAESAR (e.g. Coulthard et al., 2002).

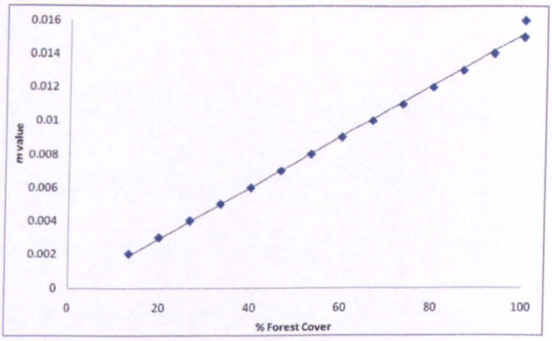

***Figure 5.12** – m value relationship with percentage of forest cover*

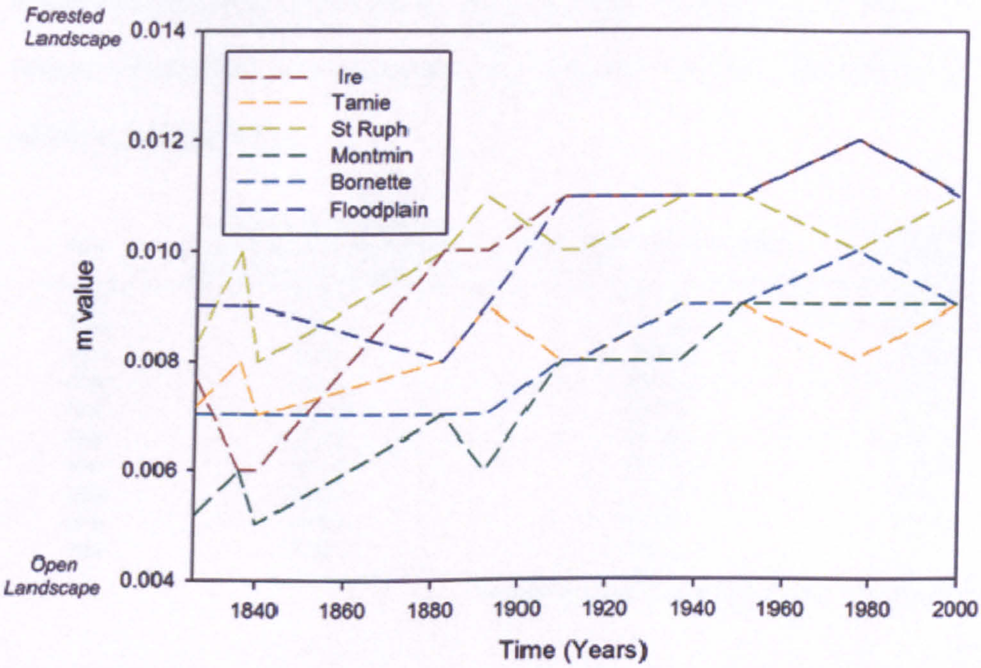

***Figure 5.13** m values used for each sub-catchment 1826-1999, based on data from Google imagery and historical records (Crook et al., 2002).*

Thus percentages of forest cover (Table 5.8) were assigned an *m* value, which enabled the historical forest inventory records to inform a land use history for the recent past (1826-1999) for each sub-catchment (Fig.5.13).

5.5.6 Land use 0AD to 1826AD

The historical records only extend back as far as 1826, thus for the longer simulations, pollen data in particular the ratio between arboreal and non-arboreal pollen have been used as a proxy for relative forest cover (e.g. Coulthard et al., 2002). Several pollen records have been produced for the Grand lac d'Annecy (e.g. David et al., 2001, Noël et al., 2001), and here the well-dated sequence for the Petit Lac d'Annecy (Jones et al., in press) is used with this pollen record. The percentage of average catchment forest cover from the periods available from the historical records (1826-1999) was calculated and compared with the same periods in the pollen data (Table 5.9).

Year	% Average catchment forest cover from historical records adjusted to aerial photographs	% Averaged Arboreal Pollen	% to adjust AP record
1999	68.134	92.806	24.672
1976	69.606	87.110	17.504
1950	67.650	86.865	19.215
1936	65.731	69.328	3.597
1913	63.434	64.173	0.740
1909	64.042	79.325	15.283
1892	58.701	77.720	19.019
1882	54.767	75.182	20.415
1840	48.111	75.311	27.200
1836	51.462	76.567	25.105
		Average % to apply to 2000 year record	**17.275**

Table 5.9 *Differences between % forest cover from adjusted historical records and average arboreal pollen.*

Based on the comparison (Table 5.9), the % arboreal pollen record has been scaled downwards by 17.3 % so that it is more comparable to the percentage of forest cover calculations based on forest inventory data (Fig.5.14a). The pollen record has then been converted into a series of *m* values (Fig.5.14b), and this time series was used to drive the longer CAESAR simulations runs for all the sub-catchments.

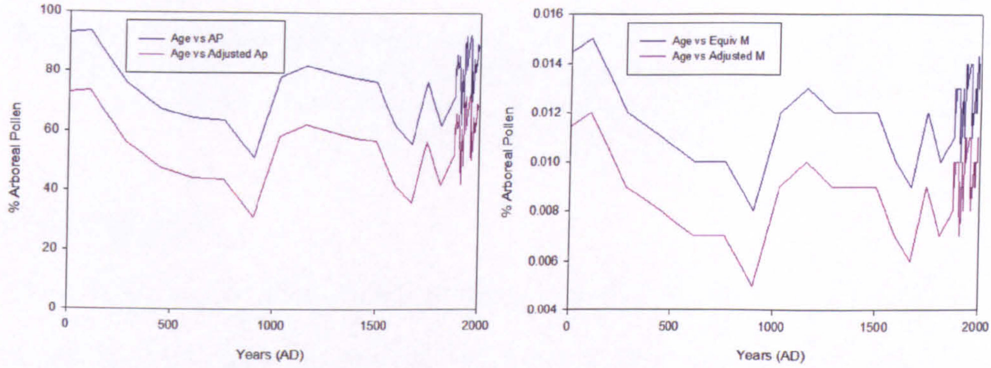

Figure 5.14 *(a)Arboreal pollen record and adjusted arboreal pollen record (b) m value land use record based on arboreal pollen adjusted to aerial photographs*

Chapter 6

Testing a cellular modelling approach to simulating late-Holocene sediment and water transfer from catchment to lake in the French pre-Alps from 1826.

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(2009) Holocene, 19(5), 785-798

6.1 Introduction

The application and testing of any numerical model in a rigorous manner necessitates comparison or validation against real observations. These issues have been addressed here by developing and using an established hydro-geomorphological model (CAESAR: Coulthard 1999, Coulthard et al., 2002, Coulthard et al., 2005, Van de Wiel et al., 2007) to simulate sediment transfer within a lake catchment. A time series of simulated sediment and water discharge for the last 180 years is presented here and driven by hourly time series of meteorology (instrumental) and land-use (documentary). In the past validation of CAESAR model output has been attempted by comparison with well dated geomorphological sequences (Coulthard et al., 2005), but these are inherently fragmentary, discontinuous, and spatially variable. The potential advantage of using lake sediment records is the availability of lacustrine sediments that can provide an integrated record of sediment flux from the catchment over long timescales against which simulated sediment delivery could be compared. Although lake sedimentary archives do not always reflect solely allogenic processes, by carefully selecting proxies that reflect detrital soil-sediment in the lake sediment, comparisons can be made with the sediment delivery simulated using CAESAR. Thus the objectives are to present a first attempt at cross-validating the sediment flux to a lake produced by this catchment-scale hydro-geomorphological model (CAESAR) and to explore the relative contributions of long term climate, meteorological events and land use in forcing or conditioning of this system.

6.2 Model Setup and Initial Conditions

Here the methodology and model setup is described only briefly, with more detail given in chapter 5. These experiments focus on two contrasting styles of sub-catchment; the Ire, a bedrock-confined mountain torrent, and the Tamie, also a steep-sided system but with a wider floodplain. Altitude and terrain are the main controls upon land use (Foster et al., 2003), with sub-alpine grasslands on lower gradients above 1500m in part maintained as summer pasture. Below ~1500 m these grasslands become extensive coniferous forests, and the steep valley sides are covered by deciduous and coniferous forest. On the flatter lower-lying terrain, the forests have been cleared and fine-grained alluvial soils support intensive cultivation.

These simulations make use of a 50 m x 50 m grid cell DEM (Institut Géographique National, France) which presents sufficient resolution to model catchment sediment output for comparison with the lake sediment records. Clearly these simulations do not include the Eau Morte floodplain (Fig.6.1), where there has been extensive deposition and storage of both coarse and fine-grained sediments during the Holocene. Sediment provenance analyses comparing catchment soil and sediment with the lake sediments show a strong connectivity between hillslope zones and the lake basin (Dearing et al., 1999; Foster et al., 2003). The principal implication is that the simulations of the Ire and Tamie should over predict absolute sediment flux, but the temporal pattern of sediment supply should be unaffected. The distribution of unconsolidated sediment and soils was incorporated through the use of a second DEM, which reflected the elevation of the bedrock surface (see 5.4.2). Comparison with the field data shows these estimated sediment/soil thicknesses after the model spin-up period to be broadly realistic.

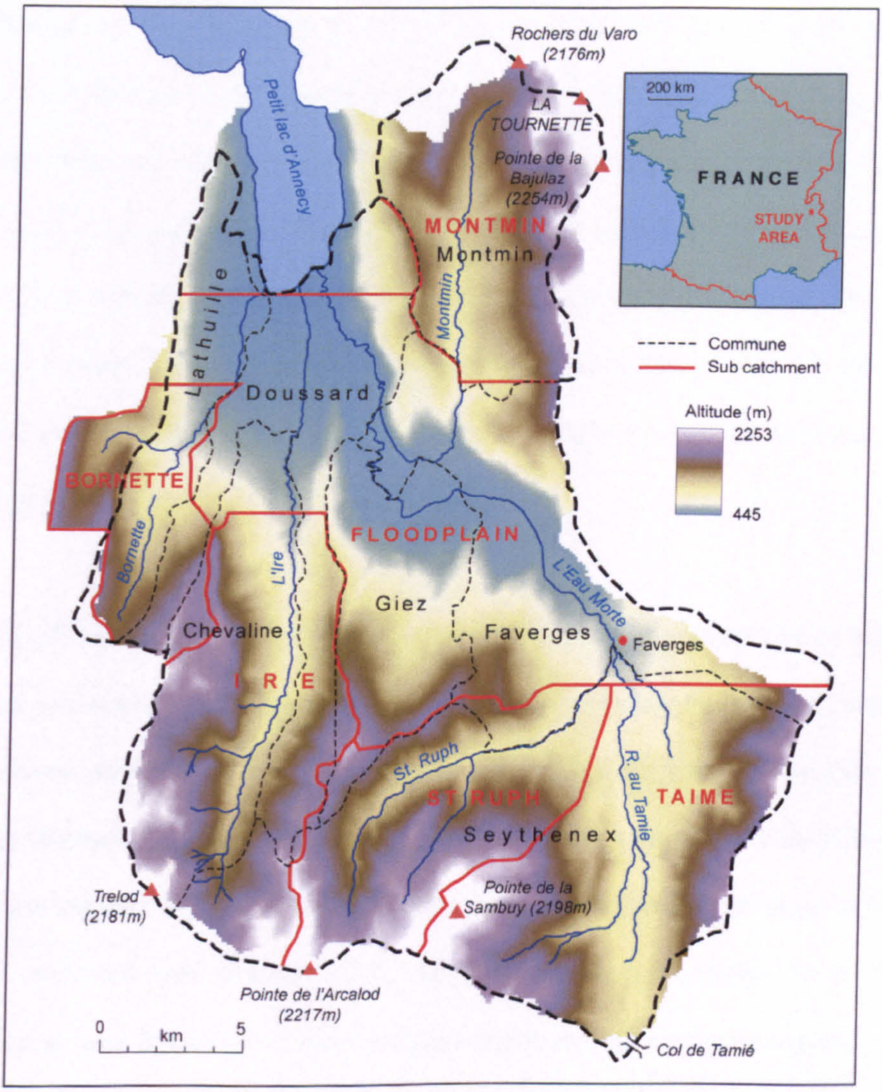

Figure 6.1 *Petit lac d'Annecy: main inflows, lake basin, sub-catchment boundaries, commune boundaries, elevation, major summits. Inset shows location of study area in France*

An hourly precipitation series was produced (see 5.4.3.1) using the hourly data recorded between 1991 and 2005 for Albertville (c. 28 km away) as a 5 year template and adjusted to monthly precipitation totals from an integrated Annecy and Geneva precipitation record (Nicholson and Thompson 2000). Snow storage and melt have been incorporated by using the SMB programme (see 5.5.2) to modulate the precipitation into a snow-adjusted series using temperature data available from Meythet (Fig. 5.3). The annual snow storage cycle produced was similar to empirical data recorded from the Alps, with winter storage (November – March) and a mid-March snow melt pulse. Thus the adjusted precipitation series incorporates changes in both thermal- and hydro-climate since AD 1826. Limitations to this approach stem from discrepancies between the Albertville time series and actual events in the Petit lac catchment. The adjusted precipitation series was applied uniformly across the DEM grid as currently it is not possible in CAESAR to vary meteorological inputs on a spatial basis.

CAESAR has been adapted to allow *m* to vary on an hourly basis thereby following the meteorological input series, and this is applied uniformly to every cell across the catchment. As the DEM has been split into sub-catchments, land use histories can be varied between sub-catchment, which at present is the only viable method by which land-use can be varied on a spatial basis. An area-weighted mean forest cover has been calculated (see section 5.5.5) using the commune coverage in each sub-catchment, and these forest cover estimates were calibrated to contemporary forest cover discerned from aerial photographs (©2008 Google - Imagery). Nonetheless, these data (Fig.6.2a) were used to provide a history of the land use for each sub-catchment in the CAESAR model runs and were calibrated against instrumental water discharge data (section 5.5).

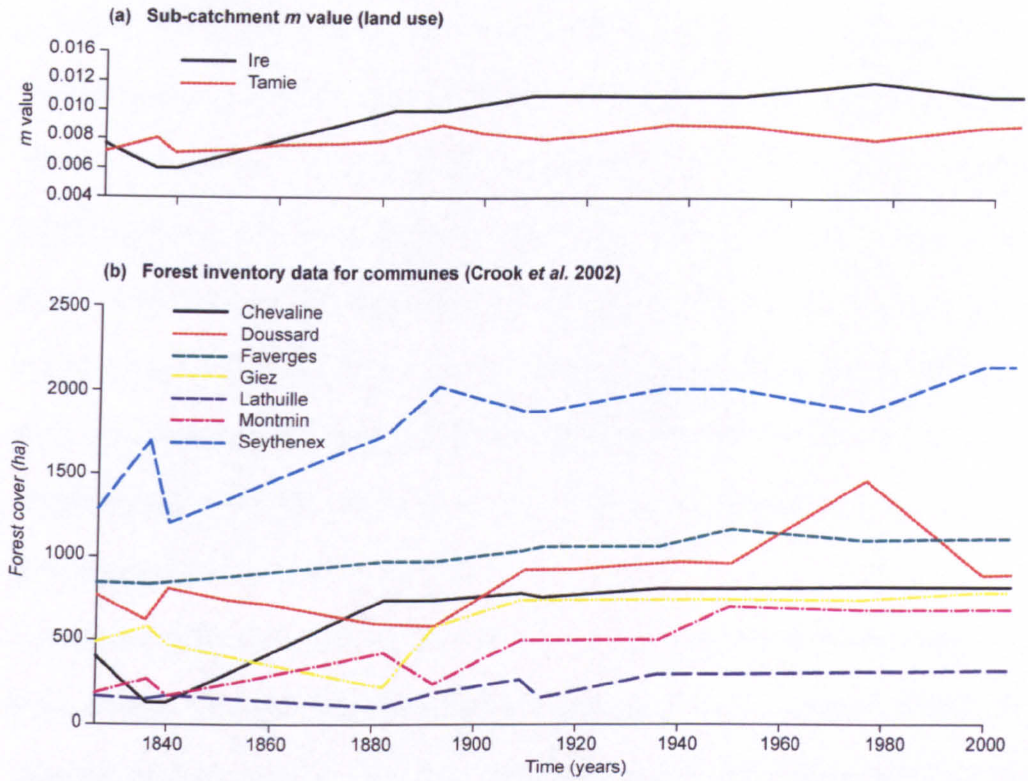

Figure 6.2 *(a) m values for Ire and Tamie sub-catchments (AD 1826-1999) used to drive CAESAR based on data from (b) historical commune forest inventory records (Crook et al., 2002).*

6.3 Methodology

6.3.1 Lake Sediment Record

In 1995, an 814 cm Kullenberg core (LA13) was taken from the flat central basin of the Petit lac in 55 m of water. Previously published lake sediment accumulation rates in this core were based on tuning a magnetic proxy of erosion in the lake records to the meteorological time series for the past few centuries (Foster et al., 2003). Here, an alternative chronology for LA13, which avoids the linkage to meteorological data and circularity of reasoning, integrates ages obtained from absolute dating techniques (^{137}Cs, ^{210}Pb and ^{14}C) and marker chronological horizons identified in magnetic and palynological proxy data (Jones et al., in prep.). The period from the early 19th century to the present (1995) is represented within the upper 1 m of sediment. Mean sediment ages from a cubic polynomial age-depth model provide temporal control for the sediment proxy data. Dry mass sediment accumulation rates calculated from the ^{137}Cs and ^{210}Pb depth-age model assuming constant rate of supply have also been used (Appleby and Oldfield, 1978). These age-depth models are an interpretation of the chronology, and include inherent error and uncertainty. In comparisons between the lake sediment record and modelled sediment discharge, there are no expectations of a 'perfect' temporal match, but a broad similarity in the overall pattern might be expected around a possible uncertainty of ±10 years for these timescales.

A proxy for sediment transport to the lake is provided by paramagnetic susceptibility (χ_{para}). This is the high field (800–1000 mT) susceptibility measured on a Molspin vibrating sample magnetometer (e.g. Snowball et al., 1999), expressed in mass specific units (10^{-8} m^3 kg^{-3}). Unlike measurements of low field susceptibility (χLF), χ_{para} records all Fe-containing minerals in a sample. Fe-containing paramagnetic clay minerals are ubiquitous in the soils and glacial sediments of the Petit lac and, quantitatively, more important than ferrimagnetic minerals. The use of χ_{para} as a proxy for detrital sediment transport also avoids the complications of non-catchment authigenic minerals, like bacterial magnetosomes and iron sulphides dominating the sediment magnetic record (Dearing 1999; Dearing et al., 2001).

6.3.2 Model Setup

Two model simulations were undertaken for the period 1826 to 2005 for the Tamie and Ire sub-catchments. The simulations were driven using an hourly 'snow mass balance'-adjusted rainfall series derived by modulating hourly precipitation by temperature (as described in 5.5.2). The simulations used hourly *m* value series that reflected the land use history of each sub-catchment (Fig.6.2b). Both simulations included a 60 year run up period (not shown), 1765-1825, to allow the model to stabilise (section 5.6). The outputs from CAESAR simulations are aggregated to produced hourly water (Fig.6.3c) and sediment discharge (m^3 s^{-1}) time series from flows leaving the right-hand side of a DEM. Although the model outputs hourly data, during flood events, the calculations are resolved at a much finer time step and aggregated to hourly outputs. Sediment distribution was subdivided into nine grain size fractions (boundaries: 62.5 μm, 250 μm, 500 μm, 1000 μm, 2000 μm, 4000 μm, 8000 μm, 32000 μm, 128000 μm) the finest of which <62.5 μm, coarse silt, was handled as suspended sediment and used for comparison with the lake sediment record (Fig.6.3d-h). The finest grain size fraction modelled in CAESAR was used for comparison against the lake sediment record as this is the dominant grain size fraction in lake sediments. Generally, the nine grain size fractions used in CAESAR follow a similar pattern of temporal sediment discharge, with all the largest events captured within all nine grain size fractions, but amplified in the finest grain sizes. Changes in surface elevation were recorded as a series of DEMs at a pre-determined interval (decadal) and reveal the spatial and temporal pattern of erosion and deposition.

6.4 Results and interpretation

6.4.1 Overall patterns

Figure 6.3 shows the water discharge (Q_w) (Fig.6.3c) and the modelled sediment discharges for the finest grain size (62.5 μm) for the Ire and Tamie which have been totalled on both annual and a 5-year basis for comparability (Fig.6d-h) with likely temporal precision of the lake sediments. The overall trend of Q_{ss} in both catchments shows high values in the periods 1840-1860, 1920-1930 and 1955-1970, though

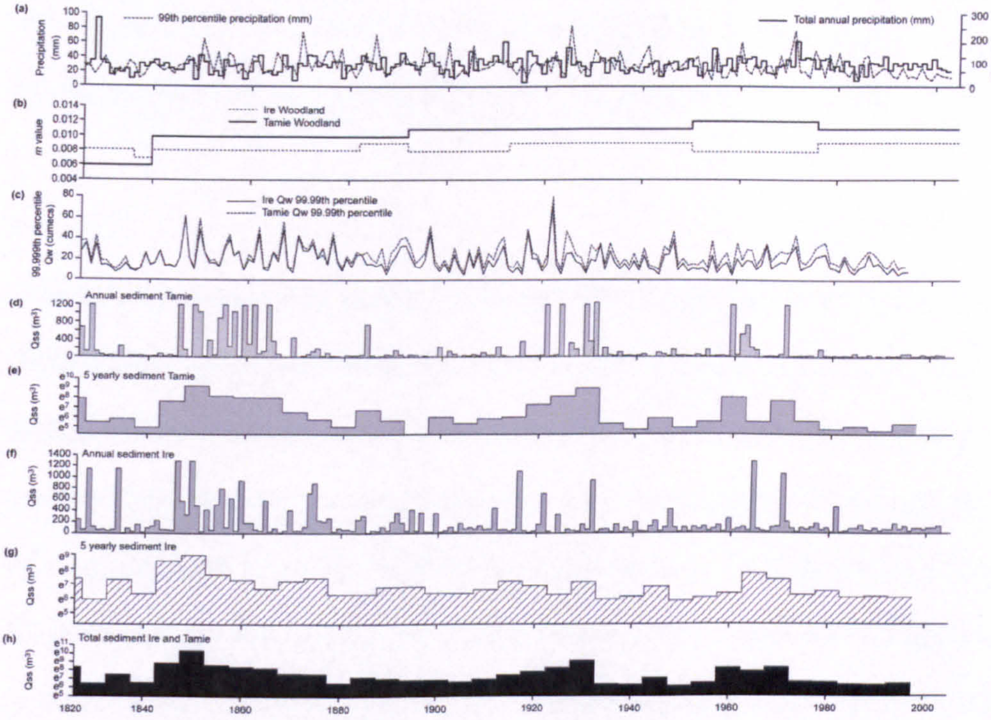

Figure 6.3 *(a)Total annual precipitation and 99th percentile of precipitation, (b) m value to represent forest cover in Ire and Tamie, (c)Ire and Tamie 99.99th percentile of water discharge, (d) Annual suspended sediment discharge, Tamie, (e) 5 year totals of suspended sediment discharge, Tamie, (f) Annual suspended sediment discharge, Ire, (g) 5 year totals of suspended sediment discharge, Ire, (h) Summed 5 year totals of suspended sediment discharge Ire and Tamie (combined)*

values for the period 1920-1930 in the Ire are proportionately lower. The sediment peaks in the Tamie (Fig. 6.3d-e) appear more pronounced, yet the level of 'background' sediment discharge in the Tamie is lower than in the Ire (Fig.6.3f-g), which may be exaggerating the appearance of the peaks. The Ire has greater annual sediment discharge than the Tamie, exemplified by the period between 1880 and 1910 when the Tamie has extremely low amounts of sediment discharge. In both sub-catchments, the largest floods e.g.1850, 1865, 1870, 1885, 1922, 1933, and 1971 were produced by the largest precipitation/snowmelt events. However, the magnitude of flood events was not solely a function of precipitation input as forest cover (*m* value) clearly amplifies (during deforestation) or reduces (during afforestation) the hydrological response to precipitation/snowmelt events of broadly equivalent magnitude, for example 1850 and 1940 respectively (Fig.6.3c).

6.4.2 Ire

The total suspended sediment (Q_{ss}) discharge from the Ire remained relatively high throughout the 180 years, which suggests the system was well-coupled and regularly flushed with materials from valley-side and montane source areas. Q_{ss} was relatively low from 1826-42, with a sustained peak from 1842-62 and then declining gradually to lows in 1862-68 (Fig.6.3f). Further peaks in Q_{ss} occurred 1868-78, 1888-97, 1912-32, 1942-47 and 1962-77 (Fig.6.3f). Almost all of these coincided with a single year or series of years with substantial floods. Considerably large floods (e.g. 1848 and 1850) produced a greater sediment response, and this also appeared to leave a legacy of sediment in the channel which was available for transport in the years after. The 1850 event rendered the system particularly sediment-rich, with even relatively small subsequent floods producing large sediment peaks in the period 1850-65. Similarly the flood event at 1971 had an initial peak of sediment, followed by another large peak at 1982, which was a response to a much smaller flood event reflecting that the channel remained sediment-rich following the 1971 flood.

The steep valley sides and restricted floodplain of the Ire provide relatively little accommodation space for sediment storage. Figure 6.4a-c shows that the channel was mainly eroding throughout the time period 1836-1866, but there was localised temporary sediment storage in and around the channel (within the constraints of a 50x50m DEM).

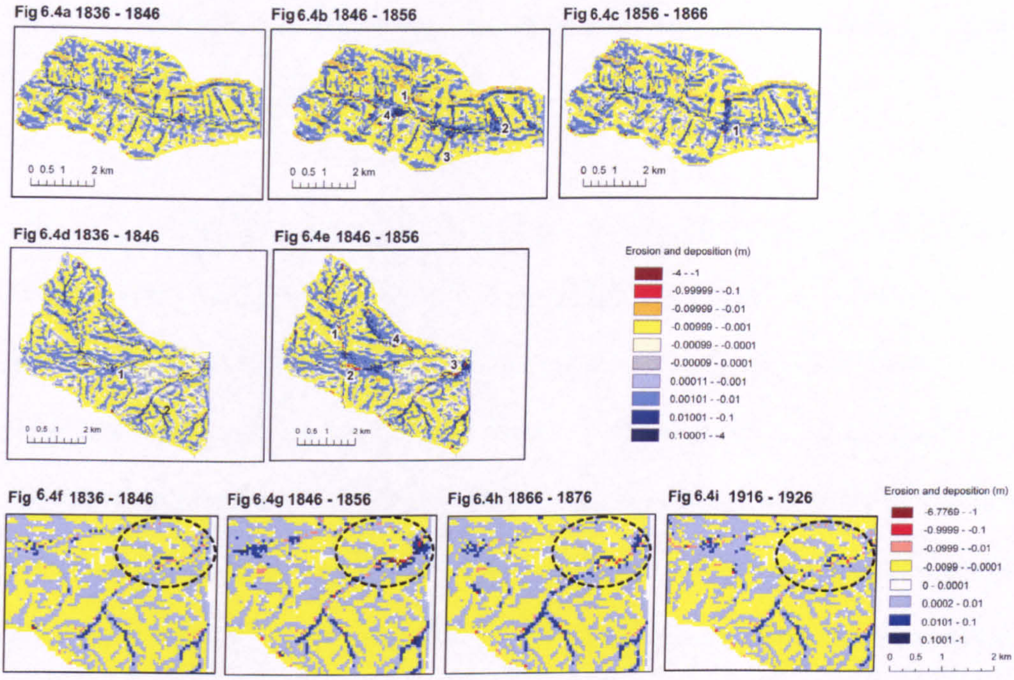

Figure 6.4 *Erosion and deposition in sub-catchments at various time windows (a) 1836-1846 Ire, (b) 1846-1856 Ire, (c) 1856-1866 Ire, (d) 1836-1846, Tamie, (e)1846-1856, Tamie, (f) 1836-1846 tributary junction, Tamie, (g)1846-1856, tributary junction, Tamie, (h) 1866-1876 tributary junction, Tamie, (i) 1916-1926 tributary junction, Tamie*

Fig.6.4 shows the spatial pattern of erosion and deposition at decadal time-steps through the substantial peak in Q_{ss} at 1850-64 (Fig.6.3). Slope processes, for example soil creep and soil erosion, occur gradually and at a uniform rate (approx. 7 mm/year for soil creep and 5 mm/year for soil erosion) throughout the modelled period only varying with slope angle, and are visible as the minor amounts of erosion and deposition of sediment across the DEM (Fig.6.4a-c). Soil erosion rate calculations in CAESAR are similar to the slope length term within the USLE, and imitate a slope wash term (Coulthard, pers. comm.) In the comparative lull of Q_{ss} between 1836 and 1846 the rates of erosion and deposition associated with channel processes were an order of magnitude greater than on the hillslopes (Fig.6.4a). Prior to the 1848 and 1850 flood-induced peaks in Q_{ss} the system experienced incision (c.10cm erosion) in the main channel and tributary systems, with localised temporary storage in zones along the main channel (<1 m deposition).

During the 1848-1850 floods (Fig.6.3c) there was greater erosion in the hillslope tributary systems (Area 1: Fig.6.4b), with deposition and growth of tributary junction alluvial fans (Area 1:Fig.6.4b.1) particularly at broader sections of the valley floor. There was also localised storage of materials in and adjacent to main channel and tributary systems (Area 3: Fig.6.4b). Supply of sediments at this time exceeded the transport capacity of the system creating local depo-centers (Area 4: Fig.6.4b). In the years 1850-1864, the river channels and more substantial tributaries experienced flows and remained sediment-rich (Area 1: Fig.6.4c), which may explain the higher background sediment discharge outside of the flood-induced sediment peaks (Fig.6.3f).

Assessing the impact of forest cover on sediment and water discharge is difficult owing to the relatively minor changes in forest cover since 1826. However, the falls in forest cover (*m* value) 1892-1915 and 1953-75 both coincide with minor increases in Q_{ss} and coincide with a time when peak discharges were low. The slightly flashier nature of the flood hydrograph under lower *m* values was sufficient to mobilise sediments. These impacts are clearer for the 1892 to 1915 phase than 1953 to 1975 when the system was perhaps starved of sediment due to low flood discharges from 1930 to 1953. This further highlights the importance of conditioning and sediment history in controlling the response of the fluvial system to flood events.

6.4.3 *Tamie*

The total annual suspended sediment discharge (Q_{ss}) was relatively low and somewhat discontinuous throughout the 180 year record perhaps indicating that this system had a better capacity to store sediment and that under low flows the channel was poorly coupled to sediment source zones. Q_{ss} was low during 1826-1845 (Fig. 6.3d-e) followed by large peaks which were sustained during 1845-1866 before a subsequent decline in Q_{ss} during 1866-1920. Three other clear peaks in Q_{ss} occurred during 1920-1932, 1960-1965 and 1967-1972, all of which coincided with large flood peaks demonstrating the strong link between floods and Q_{ss}. The larger floods, as with the Ire, produced the largest sediment responses e.g. 1850, 1885, 1924, 1925 and 1934. The peaks in 1848 and 1850 may have rendered the channel sediment-rich for a similar period of time as seen in the Ire (c. 14 years) because there were large peaks in sediment discharge until c. 1864 (Fig.6.3a). The broad valley floor of the Tamie is less restricted than in the Ire and provides a greater capacity for sediment storage, but areas, particularly the valley floor, are not mobilised by the channel on an annual basis creating a greater potential for sediment storage. These circumstances enabled the valley floor to accrete, evidence of which is seen in local valley floor depocenters and particularly in the tributary alluvial fans (Fig.6.4d.1).

Prior to the large flood-induced Q_{ss} peaks in 1848 and 1850 (Fig.6.3f), channel erosion was 10 cm-90 cm in the main channel and localised tributaries (Fig.6.4d.2). A similar amount of material was deposited in depocenters in and around the channel, particularly on the tributary alluvial fans (Fig.6.4d.2). The period 1846 to 1856 saw incision in localised zones of the main channels and large tributaries (Fig.6.4e.1) of up to 1m; this included up to 1m of fan-head incision on the alluvial fans (Fig.6.4e.2) and tributary junctions (Fig.6.4e.3). The large floods at 1848 and 1850 enabled the incision of a small tributary valley (Fig.6.4e.4) where erosion and deposition during the period 1846-1856 (c. 1-10 cm) increased by an order of magnitude compared to the erosion and deposition in this zone from 1836 to 1846 (c.1 mm-1 cm). As with the Ire, the forest cover did not vary dramatically enough for there to be significant responses in the hydrological and sediment regime. Flood magnitudes were much more closely related to precipitation and snow melt inputs, and this appears the primary driver of sediment flux in both sub-catchments.

6.4.4 Differential response in the sub-catchments

Many factors appear to have influenced the response of these two sub-catchments, effectively rendering them systems that function largely on an independent basis. Whilst the precipitation inputs remain the same, and the land use histories similar, the geomorphology of these sub-catchments are different and essentially this has governed the hydrology and the response of the sediment regime. The Ire is a generally well-coupled fluvial and hillslope system, whereas the Tamie has a much wider floodplain with zones where long-term (decadal) storage of sediment is prevalent. Within the Ire opportunities for sediment storage are limited to zones in and around the channel, thus high flows regularly flush and remobilise sediment through the system. This renders the system sensitive to discharge events. The Tamie is more complex and has greater potential for the temporary storage of sediment outside of the channel. During high magnitude flows, sediments were mobilised, transported and then deposited in temporary depocenters in the hillslope tributary systems, on alluvial fans and on the wider floodplain away from the active channel

(Fig.6.4e.2). In short, there was greater interruption to the 'sediment conveyor' in the Tamie than the Ire.

The confluence of several streams in downstream reaches of the Tamie (Area 3: Fig.6.4e) is particularly interesting, because many of the small sediment pulses (Figure 6.3d) e.g. 1835 (Fig.6.4f), 1858 (Fig.6.4g), 1885 (Fig.6.4h), 1922 (Fig.6.4i), 1924 (Fig.6.4i) appear to be controlled by erosion at the tributary junction (Area 3: Fig.6.4e). Since there was relatively little erosion elsewhere in the catchment this could explain the source of the increased sediment discharge. Both systems have exhibited storage in the lower and shallower gradient reaches of tributaries (Fig. Area 3: Fig.6.4b, Area 1: Fig.6.4c, Area 2: Fig.6.4d, Area 5: Fig.6.4e), which highlights their role in storage of sediment and interruption of the sediment conveyor.

As overall forest cover reduced (Fig. 6.5h) between 1835 and 1885, changes in the *m* parameter have increased flood magnitudes (Fig. 6.5f), which in turn have produced greater peaks in sediment discharge (Fig.6.5c). Conversely, as forest cover increased (Fig.6.5h) and flood peaks reduced in magnitude (Fig.6.5f), the sediment transmission appears to decline (Fig.6.5c). These patterns reflect the conditioning of the flood hydrograph by forest cover, and this indirectly affects sediment transmission. The response of individual catchments to a sequence of meteorological events, climate and land use is in part governed by historical changes, but the study shows that there was significant variation between adjacent systems or reaches governed by catchment morphometry and sediment distribution (Coulthard et al., 2005). Thus in order to understanding the flux of sediment within a large basin there is a clear need to model all feeder sub-systems on an individual basis.

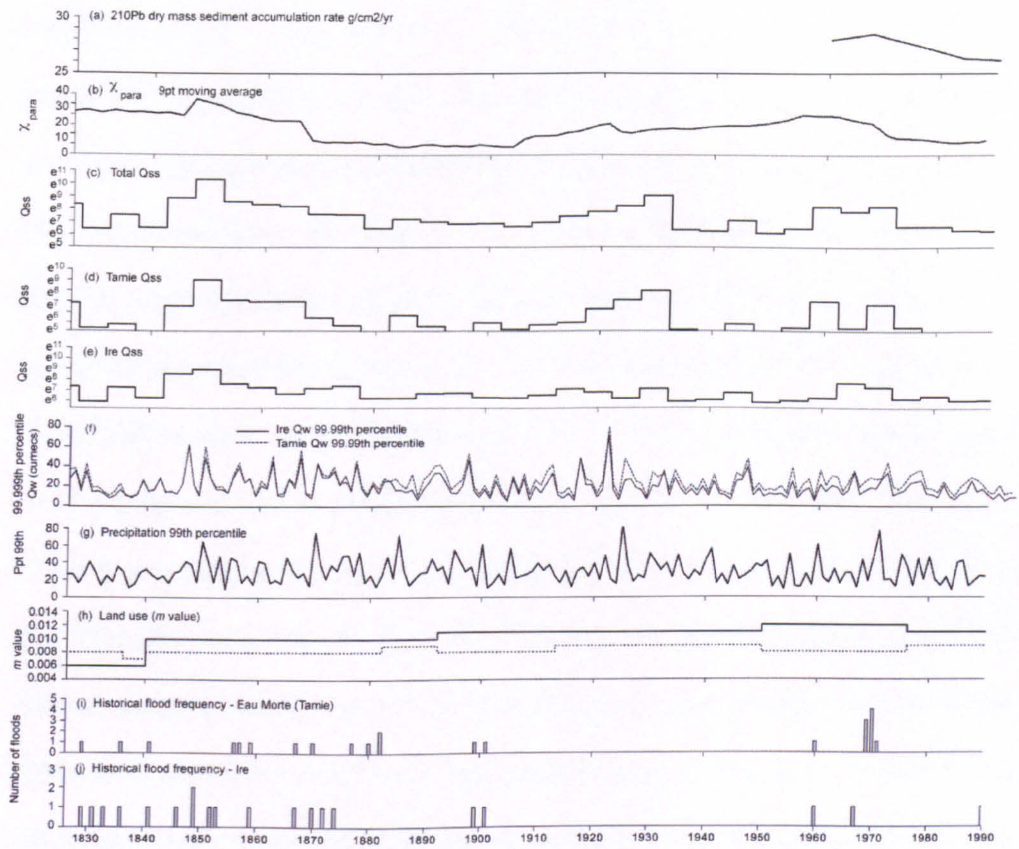

Figure 6.5 *– Comparison of modelled sediment and water discharge data with lake sediment proxies and sediment accumulation rate (a)^{210}Pb sediment accumulation rate (Foster et al., 2003), (b) χ_{para}% 9point moving average, magnetic proxy from lake sediment record (Dearing et al., 2001, Jones et al., in prep.) for detrial sediment, (c) Total 5 year summed suspended sediment discharge for Tamie and Ire (combined), (d) 5 year total suspended sediment discharge, Tamie, (e) 5 year total suspended sediment discharge, Ire, (f) 99th percentile water discharge, (g) 99th percentile precipitation, (h) m value for Ire and Tamie representing land use, (i) Historical flood frequency records for Eau Morte (Crook et al., 2002), (j) Historical flood frequency records for Ire (Crook et al., 2002).*

6.4.5 Model validation

In order to improve the comparability of the model outputs and lake sediment magnetic parameters, the total 5 yearly modelled sediment records for the Ire and

Tamie sub-catchments have been totalled, representing ~30% of the total catchment area draining into the Petit lac. Several mineral magnetic, geochemical and bulk density parameters were available as potential detrital sediment proxies for this lake sediment record. Lake sediment measurements of χ_{LF} (low field magnetic susceptibility), SIRM (saturated isothermal remanent magnetisation) and soft remanence are significantly affected by the presence of bacterial magnetosomes or atmospheric pollution (Dearing et al., 2001). Paramagnetic susceptibility χ_{para} (Fig.6.5b) appears therefore the most suitable magnetic proxy for the transport of Fe-containing detrital soil/sediment to the lake. The effects of rapid urbanisation and accelerated eutrophication during the early to mid-20th century also mean rejecting the available lake sediment geochemical and density measurements. However, a relatively precise sediment mass accumulation rate is available for 1960 -1995 using the ^{210}Pb chronology for the Petit lac (Fig.6.5a; Foster et al., 2003). The ^{210}Pb-derived sediment accumulation rate tracks the χ_{para} pattern closely over the period 1960-1990 (Figure 6.5) with both records displaying a peak around 1960 before declining around 1970. This internal consistency strongly implies that the two proxies reflect broad variations in the supply of catchment soil-sediment and can be used to validate the modelled Q_{ss} records. It should be noted that the chronology of the lake sediment record shown (Fig.6.5b) has an error of ±10 years which means that model validation has to focus on the long term pattern of change rather than the synchrony of peaks and troughs in the time series.

Modelled total Q_{ss} (Fig.6.5c) rises from 1826 to a peak around 1850 before declining around 1885. The peak appears to coincide with the χ_{para} record (Fig.6.5b), which shows a maximum around 1850 before declining to a minimum around 1865-1880.

The period from 1835-1880 has the lowest forest cover (Fig.6.5h) which may have led to larger floods, thus amplifying the peaks in Q_{ss} and increasing the amount of detrital soil-sediment. These conditions provide a plausible mechanism for explaining the high levels of χ_{para} around 1850. Modelled Q_{ss} (Fig. 5c) increases from c.1910 to a smaller peak around 1935 before declining, a pattern paralleled by the χ_{para} values (Fig.6.5b). A decline in Q_{ss} (Fig.6.5c) occurs from 1940 with a slight increase around 1950 which tails off before another increase in Q_{ss} around 1960-1980, again broadly tracking the χ_{para} curve (Fig. 6.5b). This peak is also mirrored in the ^{210}Pb sediment accumulation rate (Fig. 6.5a). All three sets of broad peaks match within the errors of the age-depth model.

Whilst the precipitation inputs remain high between 1925 and 1960 (Fig.6.5g), the response of the fluvial system (Fig.6.5f) is diminished, which is a result of increased forest cover (and therefore *m* value) within the modelled catchments (Fig.6.5h). Whilst a reduction in water discharge was expected due to the nature of the hydrological model, the peaks in sediment were also reduced at this time. The match between Q_{ss} (Fig.6.5c) and χ_{para} record (Fig.6.5b) between 1925 and 1960 perhaps reflects that the increased forest cover did diminish the response of the fluvial system and therefore the sediment discharge. The Ire Q_{ss} record (Fig.6.5e) appears more comparable with χ_{para} than the Tamie Q_{ss} record (Fig.6.5d). This is particularly noticeable around 1920 where the peaks in Ire Q_{ss} and χ_{para} were less pronounced. These differences may reflect that the Ire was more efficient in delivering sediment with an almost instantaneous response in Q_{ss} to flood events. For the Tamie the sedimentary response is likely to be masked and more complex due to the greater opportunities for sediment storage. The similarity between the χ_{para} and Ire-style Q_{ss}

in turn perhaps signifies that the lake was more sensitive to inputs from efficient flashy systems (e.g. Ire) and that the lake sediment sequence is strongly coupled with flood sediment supply.

Crook et al., (2002) collated documentary evidence showing flood frequency (but not magnitude of events) within the Tamie (Fig.6.5i) and Ire (Fig.6.5j) since the late 17^{th} Century. These events are regarded as severe enough to cause damage to land and thus were reported in order to receive compensation for loss (Crook et al., 2002). There are inaccuracies in the historical record as not all events may have been recorded as some land owners may not have reported the incident or damages to their land. No documentary records are available specifically for the Tamie, but records exist from the Eau Morte catchment into which the Tamie flows downstream of Faverges (Fig.6.1). These may be used as a substitute record for flood events in the Tamie although localised flash flooding does not always propagate downstream with the same intensity (Crook et al., 2002). The modelled water discharges (Fig.6.5f) are a function of the engineered precipitation input series (snow melt pulses) and also *m* value, nevertheless there are some striking parallels between high χ_{para} and historical floods around 1850 and 1870 strengthening the suggestion that χ_{para} is a proxy for detrital catchment soil/sediment that is mobilised under the highest flows. Similarities in peak Q_{ss} (Fig.6.5c) and the historical records for (large) floods in the Tamie (Fig.6.5i) and Ire (Fig.6.5j) occur at 1850, 1865-1875, ~1900, 1960 and ~1970 in particular, which suggests that CAESAR has captured the behaviour typical of the larger flood and sediment events.

6.5 Discussion

By approaching the movement of sediment in a catchment within a modelling framework, it is possible to separate system drivers (in this case precipitation and land use) and use them to gain some insight into the drivers of the geomorphological change (e.g. Fig.6.3 and Fig.6.5). The study demonstrates the advantage of using CAESAR to explore the impacts of meteorological, long term climate and land use change within a catchment. The model produces both high temporal resolutions (aggregated to hourly) over relatively long periods of time (10^0-10^4) with a series of sequential raster data (maps of erosion and deposition) that allow analysis of the temporal and spatial patterns of change.

6.5.1 Model validation and understanding hydro-geomorphic processes

In order to apply a numerical model to any real-world problems, validation and sensitivity testing of the model is essential. Previous attempts to validate long term landscape evolution models have assessed valley long profiles (Hancock et al., 2002; Tucker and Slingerland, 1994) and undertaken hillslope plot experiments (Hancock et al., 2002), both of which are fraught with problems when modelling beyond the recent past (Coulthard et al., 2005). Coulthard et al. (2005) used a well constrained (radiocarbon-dated) investigation of both sedimentation style or rate discerned from the timing of geomorphologically significant changes in river activity for comparison with the CAESAR record for 4 tributaries of the River Ouse, Yorkshire (England). An alternative to validation or corroboration by comparison with geomorphological histories is to compare lake sediment evidence with (CAESAR) modelled sediment discharge. The good correspondence between the Petit lac magnetic proxy (χ_{para}) and modelled sediment record for both sub-catchments is encouraging, despite the fact that the lake sediment record is a catchment-wide record of erosion and deposition, whereas the totalled sediment record (Q_{ss}) presented here depicts time series for two sub-catchments. The lake sediment record has been smoothed using a moving average (nine point) to eliminate short term variability and highlight the long term trends that are the focus of this paper. As the resolution of the lake sediment chronology is +/- 10 years, it is logical to smooth the sediment record to some extent; we regard sub-decadal as the best possible chronological resolution. In the case of non-annually laminated sediments precise comparisons cannot be made owing to the chronological uncertainty in the lake sediment proxies, but the parallelism shown here suggests the model is capturing the timing and magnitude of changes over longer timescales (bi-decadal) and to some extent on an event basis. In addition these Q_{ss} data incorporate neither the 3 other sub-catchments (Montmin, St Ruph and Bornette) nor sediment transmission through the sensitive Eau-Morte floodplain/delta (Fig.6.1).

The overall match between the modelled and lake sediment proxies highlights the potential for producing a modelled catchment-wide sediment record for comparison with lake sediments. The match appears slightly better for the Ire sub-catchment, which perhaps suggests that this style of system and sediment transport behaviour dominate the lake sediment detrital record. The characteristics of this system are flashy flood induced sediment transmission and it is likely that the other mountain-torrent sub-catchments (e.g.Bornette, St. Ruph) may behave in a similar manner. The lower reaches of the Ire traverse across a flood-plain (not yet modelled) and may have had significant storage potential in the 19th century before dyking and canalisation of the reach was introduced in the early 20th century. The management schemes put in place since the early 20th century certainly reduced flooding as the lower canalised reaches allowed direct transporting of sediment from the Ire to the lake. The match between χ_{para} and total Q_{ss} suggests a similar forcing; therefore one may conclude that higher flows drive the χ_{para} record in the lake sedimentary sequence. The lake may also be more strongly coupled to mountain torrent hydrology/sediment movements than the behaviour of storage-release-storage systems like the Tamie. Alternatively, the close proximity of the Ire system to the lake (4 km v 12 km) may mean that the lake record is more sensitive to local inputs. However, resolving these questions requires further modelling of sediment storage on flood plain reaches.

The role played by forest cover on the sediment regime is at present captured through the moderation of the flood hydrograph; CAESAR perhaps at present fails to capture this completely due to the simplistic and static representation of hillslope processes. This impression is perhaps compounded or obscured by the lack of dramatic changes within the forest cover of the two sub-catchments. However, the magnetic proxy records suggest that as forest cover increases and stabilises in the late 19th century a reduction in sediment delivery to the lake is reflected in the χ_{para} record, a trend that has been replicated using CAESAR. In addition, the greatest amounts of modelled sediment are moved during the periods of lowest forest cover which suggests that even minor changes in forest cover can have significant perturbations on the sediment regime. The geomorphology of the catchment plays a large role in modulating the impacts of meteorology, long term climate and land use. The Tamie displays a more complex set of responses due to longer duration sediment storage in that sub-catchment. The response of total sediment flux from the Ire is rapid, whereby sediment is moved from the hillslopes into the channel through well coupled tributaries with the only storage in the channel and of a temporary nature producing efficient transmission of sediment from this sub-catchment. Understanding systems like the Ire should be easier than systems like the Tamie where storage and release of sediment must be considered. Although the changes in forest cover are subtle for both sub-catchments, there is some evidence that forest cover conditions the response of the hydrological system and sediment regime even for small scale perturbations. When forest cover is increased in CAESAR by 10-15%, peak flood discharges appear to reduce by around 35%. These values are within a similar order of magnitude and broadly realistic when compared to previous research into the effects of land use on fluvial systems which suggests that reducing forest cover increases water yield (Brown et al., 2005). Should the changes in forest cover be more pronounced, perhaps the response in the sediment regime within CAESAR will also be more pronounced, but longer temporal simulations with greater switches in forest cover are needed.

At present the representation of hillslope processes within CAESAR is relatively simplistic, which reflects a necessary reduction in complexity to allow computational time to be spent on fluvial processes, the focus of research and development in the past (Coulthard and Van De Wiel, 2006; Coulthard et al., 2007). Clearly there is scope to increase the level of slope process representation in CAESAR for studies such as this where hillslope to fluvial system coupling is very important. Once the hillslope processes are developed, this may allow a more sensitive analysis of the role of changing land use at catchment scale. The importance of spatially distributed numerical modelling cannot be overestimated. If a DEM had not been used to represent topography for this type of study, then important sediment conditioning effects such as storage and release of sediment (e.g. Tamie) or seemingly more rapid sediment transmission (e.g. Ire) would not be identified and the temporal data alone would have provided only part of the story

6.5.2 Implications

The study shows that a cellular hydro-geomorphological model can realistically simulate the complex interactions between climate, meteorological events, land use, flooding, and sediment transport and storage for a sub-alpine catchment over decadal timescales. This modelling tool may become vital for anticipating future changes. IPCC reports (2007) suggest that for European Alpine regions, 90% of the Atmosphere-Ocean General Circulation Models (AOGCMs) concur that winter precipitation 2090-2099 will increase by between 10-20% compared with the same period in 1980-1999. For the same period, summer precipitation will decrease by approximately 20%. If these projections are met or exceeded, this would have significant implications for the hydrological and sedimentological regimes of even sub-alpine catchments, like Annecy. At present, a large proportion of the winter precipitation is stored as snow and consequently released during April and May. This study shows that snow cover and snowmelt are critical conditioning components of the sub-alpine hydro-geomorphological regime. If winter precipitation increases 10-20% as suggested by the IPCC (2007), the predicted increase in the frequency and magnitude of large flood events in winter and spring can be expected to be amplified further as water storage, in the form of snow, is reduced. These climatic projections affect a landscape that is experiencing changing land use needs (e.g. declining numbers of winter ski resorts, changes in EU Agricultural Policy) and CAESAR allows modelling of these drivers in an integrated framework.

6.6 Conclusions

CAESAR has allowed the investigation of drivers of landscape change; meteorological events, climate and land use, and assessed the impact that they have on hydrological and sedimentological regimes. Conclusions from this research suggest that:

1. Catchment morphometry and sediment distribution play a significant role modulating the pattern of sediment transmission. Thus spatially distributed numerical modelling is of fundamental importance to understand the pattern of sediment transmission at catchment scale and to identify storage and release within the hill slope – fluvial system.
2. Peaks in sediment discharge appear to be largely a function of flood magnitude driven by precipitation and snow melt, but this may be in part conditioned by the control forest cover (land use) exerts over the hydrological regime.
3. Data-model comparisons suggest that the Petit lac d'Annecy lake sediment record may be more sensitive to systems where the linkages between hill slope and lake basin are well-coupled whereas significant sediment storage interrupts the sediment conveyor and encourages this relationship to breakdown.
4. Palaeorecords such as lake sediments and in particular environmental magnetic proxies for detrital sediment supply can provide a robust methodology to validate the modelled sediment flux. Conversely modelled sediment discharge provides an experimental framework for testing hypotheses about drivers of the lake sediment record.

5. There is a clear need for further model development, for example improving and coupling hill slope processes with land use, storm events and long term climate as this would capture the hydrological and sedimentological regime more effectively.

Future research will attempt to improve representation of hillslope processes within CAESAR, to allow a fuller understanding of the role of forest cover. Here the capability of CAESAR has been tested to be used as a tool to predict geomorphological change and sediment flux from a catchment under varying meteorological events, climate and land use scenarios. To assess longer term impacts of change, for example wet/dry shifts in climate and more dramatic changes in land use, model runs are needed that address longer time periods (>2000 years) to comprehend system drivers over these time scales. The eventual aim is to provide and test a framework for understanding the annual to decadal changes in flooding and sediment transmission that may take place in the *future* so that approaches to adaptation and mitigation of flood hazards and shifts in sediment movement can be assessed.

Chapter 7

Integrated hillslope – fluvial modelling of sediment delivery since AD 1500

7.1. Introduction

Sediment records preserved in the lacustrine basins of Europe during the late Holocene reflect the combined impacts of increasing anthropogenic pressure on the landscape, climatic change and flood events (e.g. Edwards and Whittington, 1993; Foster et al., 2003; 2008; Chiverrell 2006). The detrital components of lake sediments typically reflect the averaged contribution from a catchment, where the materials derived from upland hill-slope, alluvial and deltaic environments are combined to deposit an integrated record of catchment sediment supply. This smoothing is a response to the behaviour of the different components of the catchment sediment conveyer (e.g. Fryirs & Brierly, 2001; Lang et al., 2003; Fryirs et al., 2007), with cycles of storage and release punctuating the transmission of materials from erosive source zones to eventual deposition within the lake. More rapid sediment supply to a lake or changes in the composition of the sediments (e.g. particle size, geochemical and magnetic mineralogy) may reflect increased erosion from alternative sources (eg. different geologies, or topsoil v subsoil) or improved transport efficiency but it is difficult to discern further information about the spatial pattern of geomorphological processes from the lake sediment record.

Hydro-geomorphological models (e.g. CAESAR, Coulthard et al., 2002; Van de Wiel et al., 2007, Welsh et al., 2009) can be used to test hypotheses about the spatial pattern of catchment sediment delivery (e.g. Coulthard et al., 2002), and data-model comparisons between modelled sediment discharge and lake sediment proxies allow some assessment of system sensitivity to external drivers (e.g. land use change, precipitation events). Initial hydro-geomorphological modelling (CAESAR) undertaken in the Petit lac d'Annecy catchment, in the French pre-Alps (Welsh et al., 2009), showed the importance of flood magnitude and frequency in controlling sediment delivery, but also highlighted that it appears to be strongly regulated by the morphometry of the catchment. In addition, changes in land use within CAESAR have an indirect role in moderating sediment discharge through augmentation or diminution of flood magnitudes. The spatial pattern of erosion and deposition within these model simulations highlighted the importance of both sediment delivery from the hillslopes and variation in the efficiency of coupling between hillslope and fluvial zones within the hydro-geomorphological model is important.

Much of the recent focus in the development of CAESAR has been on fluvial processes (e.g. braiding and meandering) and simulating channels at a reach scale (e.g. Coulthard et al., 2007), thus there has been little requirement to refine the representation of hillslope processes. For catchment scale modelling with a focus on simulating sediment supply to a lake basin (e.g. Welsh et al., 2009), there is a clear need for more robust representation of hillslope processes. The aim of this chapter is to improve and test the representation of selected hillslope processes within CAESAR and to simulate the effects of land use and climate change on sediment delivery to a lake basin. Currently the representation of hillslope processes in CAESAR is simplistic but few other numerical models satisfy the requirements of process representation needed for this research. Thus the objectives here are to produce a more realistic and spatially distributed representation of soil creep, soil erosion and slope failure processes in CAESAR, so that they are coupled with soil moisture and land use. Here a long 500 year time series of land use and climate are used to identify past patterns of erosion/deposition and sediment flux/delivery for two sub-catchments in the Petit lac d'Annecy drainage basin (Ire and Tamie) (Fig.7.1). The outputs from CAESAR simulations will be compared to detrital sediment indicators in the lake sediment record from the Petit lac d'Annecy (Dearing et al., 2001; Jones et al., in prep.) and to the geomorphology of the catchment (e.g. Foster et al., 2003).

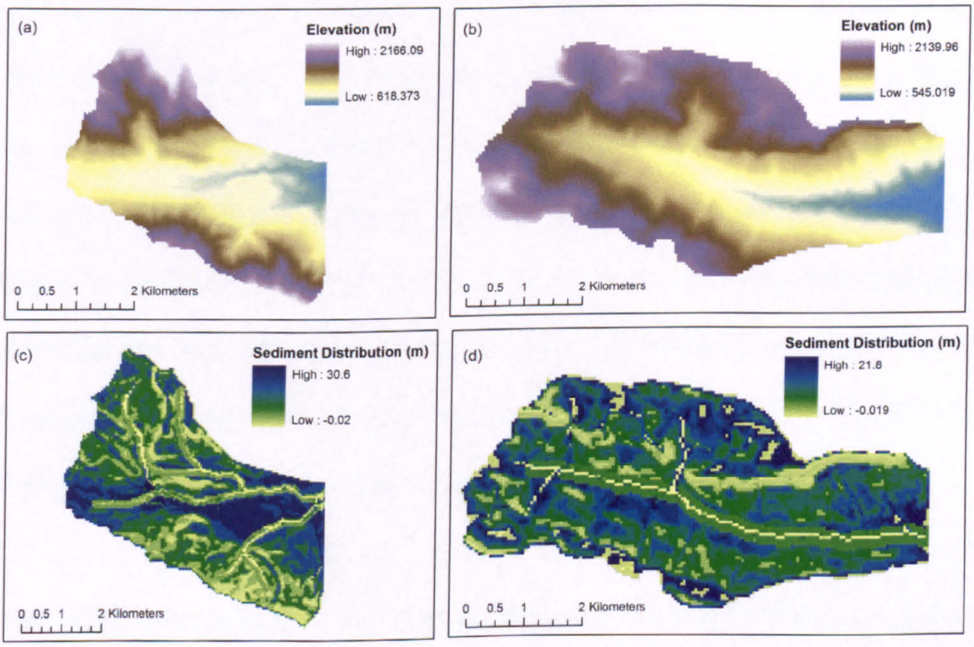

***Figure** 7.1(a) Tamie DEM (b) Ire DEM (c) Sediment distribution in Tamie after run-up period (d) Sediment distribution in Ire after run-up period.*

7.2 Geomorphic processes in mountains environments

In mountain regions, hillslopes are a vital sub-system within a catchment, and provide sources of and contribute sediment through various geomorphological processes such as mass movement (soil creep and slope failure), sheetwash, rilling and gully erosion. Lake sediments in part reflect ordered sequences of eroded catchment soils mixed with varying amounts of autochthonous material (e.g. biogenic silicic, aquatic macrophytes and authigenic carbonate). Sediments can therefore be investigated to provide a history of catchment erosion (Mackereth, 1966) and changes can be compared with climate and land use histories. Variations in the rate of erosional processes in river systems are largely determined by variations in the flow regime (e.g. flood discharges) and sediment supply (Bull, 1979; Bull, 1991). More runoff will increase critical stream power, thus encouraging erosion and overcoming the thresholds for sediment storage at tributary junctions (e.g. alluvial fans) and elsewhere in the alluvial system (e.g. floodplains). Excess sediment supply within a system may lead to aggradation of coarse grained sediment in alluvial fans (Harvey, 2002), with the fine grained material similarly retained in flood plain alluvium. The sediment supplied from the hillslopes may also eventually be transported to and deposited in lacustrine basins (e.g. Dearing, 1991), particularly if hillslope-channel coupling is effective in the catchment (Harvey, 2002).

Rates of hillslope processes are often amplified on the steep slopes of mountain zones and minor perturbations in climate or land use can produce substantial responses in the sediment regime. Natural relief (Weisshaidinger and Leser, 2006) and higher rainfall intensities (Auzet et al., 2006) in alpine zones enhance the potential for greater erosivity of slope materials. Fragile rendzina soils are commonplace across the Alps due to poor regeneration of soil at high altitudes (Strauss & Klaghofer, 2006). In conjunction with slow soil formation (~<0.1mm/year, Mosimann et al., 1999) these slopes are rendered susceptible to erosion. Weathering of black marls (Descroix and Olivry, 2002) and freeze-thaw weathering (Auzet et al., 2006) take place due to wide temperature ranges at high altitudes and consequently increase soil/sediment yield. The recent past has seen increasing demographic pressure due to mountain settlement (Auzet et al., 2006); subsequent urbanisation and poor management of grasslands (Weisshaidinger and Leser, 2006) have increased the threat of landslides (Tasser et al., 2003), exacerbated soil erosion in mountain regions and caused entrenchment of riverbeds (Brochot & Mernier, 1995). This has had significant implications for infrastructure and aggradation of material downstream, further augmenting flood risk. Whilst soil erosion may be perceived to pose the greatest threat to slopes in the Alps, Weisshaidinger and Leser (2006) suggest that creep and mass movement processes produce greater problems, although landslides remain the greatest hazard (Macquaire, 2005). In addition to this, variable rates of erosion, weathering and denudation may occur under different precipitation intensities and land use conditions. For effective management of river systems, it is vital to gain a better understanding of hillslope-channel coupling under different climate and land use scenarios, particularly identification of upstream perturbations which may cause large-scale aggradation of sediment downstream, and thus augment the likelihood of flood inundation and sedimentation (Descroix and Mathys, 2003) in these highly non-linear systems.

Currently available numerical models which handle sediment delivery such as SEDEM (Van Rompaey, 2001) require local level parameterisation of soils and land use and can only offer annual resolution data outputs. Landscape evolution models such as GOLEM (Tucker & Slingerland, 1994) and SIBERIA (Hancock and

Wilgoose, 2002, Willgoose et al., 2004) implement hillslope processes within the models, but operate at a much longer time scale or require high levels of parameterisation. CAESAR (Coulthard et al., 2002, 2005; Van de Wiel et al., 2007) provides an integrated model that encapsulates the sediment production, transmission and depositional zones (Schumm, 1977) within a single spatially distributed model and is thereby able to operate at reach or catchment scale at an sub-hourly resolution for thousands of years (e.g. Coulthard et al., 2002). A major advantage of operating at a sub-hourly resolution is that the variations in sediment delivery caused by individual flood events can be captured, which is vital as these singular events can often do more in terms of geomorphic work than several months or years of lower magnitude events (Selby, 1993). Robust handling of sediment sources, sediment release, and within system connectivity is an essential precursor to the simulation of the catchment-scale response to changes in both land use and climate. The representation of hillslope processes is required to be simplified enough so that model simulation time is not affected, yet complex enough to broadly represent relationships between various parameters.

7.3. Hillslope process representation in CAESAR

The hillslope processes represented in CAESAR 5.9 (used in Welsh et al., 2009) are slope failure, soil creep and soil erosion (a rill/wash term). However, Coulthard et al. (2002) concede that one of the main controls on sediment discharge in an upland cool maritime temperate catchment could be soil creep. They suggest that the present representation of soil creep may be suitable for short simulations (1-2 years) but may be a short coming of CAESAR 5.9 when model simulations exceed hundreds of years as the volumetric importance of sediment supply from soil creep increases over time. In CAESAR 5.9 soil creep and soil erosion operate to provide a regular supply of sediment, and simulations (not shown) that exclusion of these processes fails to produce sediment output comparable to the lake sediment record (e.g. some simulated years and decades where there was no transmission of material out of the catchment). Given the contribution of soil derived organic and inorganic materials to lake sediment archives (Mackereth, 1966; Edwards and Whittington, 2001), access to slope derived sediment and dynamic coupling between the slope and fluvial systems has encouraged a re-evaluation of the hillslope component of the CAESAR model (version 5.9).

The current representations of hillslope processes in CAESAR 5.9 (e.g. Coulthard et al., 2002, 2005; Van de Wiel et al., 2007) are as follows:

1. **Slope failure** is governed by a user defined failure slope angle, above which all the available materials in cell *n* move to the adjacent cell(s) down slope. For the Petit lac d'Annecy catchment, slope failure and rapid mass movement of sediment are important components of the geomorphic system (Foster et al., 2003), as rock fall, landslides and debris flows induce rapid sediment movement processes. Implementation of these processes in CAESAR 5.9 is constrained spatially to an arbitrary user-defined zone determined solely by slope angle (e.g. >45°), materials can only move to the adjacent cells (50m), and the *rates* of sediment movement are not governed by slope gradient or soil moisture. To prevent too rapid movement of materials there is a defined limit (0.001m) to the materials that can be moved during each user-defined time-step.

2. **Soil Creep** is determined by a user defined rate that is fixed over time and is typically based on a local average soil rate derived from empirical data (e.g. 7mm/year in the Alps c.f. Jahn, 1989). Soil creep takes place in CAESAR on a daily basis and the rates are not linked to soil moisture or land use, but do vary on a cell by cell basis controlled by slope angle. Creep takes place in every cell in the DEM and material is moved down-slope to the adjacent cell(s).

3. **Soil Erosion** acts as a wash/rill term within CAESAR (Coulthard, pers. comm., 2008) and behaves in a similar manner to RUSLE (e.g. Renard et al., 1991), although in CAESAR 5.9 there is no coupling of erosion with soil erodability, land use or practices (the K, C and P factor). In CAESAR 5.9 the rates are governed by slope angle (S factor) and the upslope drainage area of cell *n* (similar to slope length term: L factor). However, in CAESAR 5.9 these rates are not currently linked to precipitation, soil moisture or land use. Soil erosion takes place daily within a model simulation and similarly to creep, materials are moved down slope to the adjacent cell(s).

There are numerous individual process models that simulate landsliding (e.g. Hungr, 1995), soil creep (e.g. Kirkby, 1967) and soil erosion (e.g. Renard et al., 1991) that are available for use at a high level of complexity, there are no numerical models which handle these processes at a realistic level of complexity whilst being computationally simple enough to simulate multiple processes at high temporal (hourly) and spatial resolution over long (~2000 year) time periods. For catchment scale simulations (e.g. 3x4km) computer capacity necessitates the use of coarser grids (e.g. 25-50m), and as a consequence much of the finer detail, for example channel planform change and evolution, is obscured. The objective was not to write models for individual hillslope processes, thus a key constraint over the development of the hillslope component was integration with the CAESAR distributed fluvial model. Processes were therefore required to be time-stepped in a similar manner to the fluvial system with precipitation and land use as the two input time series used to drive simulations. CAESAR simulations generate a catchment wide soil moisture parameter (j_mean) that could be used to vary slope processes over time.

7.4. Methods and Approach

This section details the approach taken, the theoretical background and the practical application of exactly how the hillslope processes were improved from CAESAR 5.9. It then moves on to describe model setup and the simulation matrix used for rigorous testing of the newly implemented hillslope processes.

7.4.1. Mass movement (revised soil creep and slope failure)

Soil creep has long been studied, but many of the factors that influence soil creep have been difficult to test and quantify (Selby, 1993). The empirical and theoretical studies that have taken place suggest that creep rates are fastest near the soil surface and increase with slope angle (Davis, 1892). This can be attributed to freeze-thaw action under the certain climatic regimes (Davis and Snyder, 1898, Young 1972, Carson and Kirkby, 1972, Swanson and Swanston, 1977, Jahn, 1981, 1989) or may be exacerbated by burrowing animals (Davis and Snyder, 1898, Boyle, in prep.). Wetting and drying and heating and cooling of soil can cause expansion and contraction within the soil affecting creep rates (Kirkby, 1967, Selby, 1993) and rain-splash erosion is thought to be important in semi-arid poorly-vegetated environments (Parsons et al., 1994). Vegetation cover is also an important factor as soil is resistant to creep near plant roots (Jahn, 1989) which bind the soil near to the ground surface (Selby, 1974) therefore less creep is likely to take place under a full forest cover than under pasture land. There are many factors that influence soil creep and these vary between climatic zones.

Monitoring studies have recorded average rates of soil creep (4mm/year) for the Sudettes mountain range in the pre-Alpine zone of the Czech Republic (Jahn, 1989). Depending on slope angle and land use, he suggests soil creep can range from 2-3mm/year in low altitude meadows, up to 9mm/year in high altitude meadows, 0.5mm/year in forested zones and 12-20mm/year in areas disturbed by human activity (Jahn, 1989). Measured rates of landslides for mountain zones of similar altitude to the Petit lac d'Annecy (Selby, 1993; page 266) show how a single precipitation event can encourage debris flows and landslides of up to 3m deep and 30m wide. Simulations using CAESAR 5.9 have applied user-specified rates (e.g. 7mm/year) varied against slope angle, thus the magnitude of creep processes was spatially distributed. Given the focus on the drivers of catchment scale sediment delivery, there is a need for incorporation of moisture (antecedent weather) and land use to impact on creep rates without reducing model efficiency.

Here CAESAR has been developed to link rates of soil creep with both precipitation and land use. Creep rates have been scaled to a range of values typical of pre-Alpine zones (Jahn, 1989) using an inverse decay relationship, whereby the rate of creep, increases with slope angle up to a slope failure threshold (Fig 7.3a). Landuse and climate impact on this relationship by varying the failure threshold each timestep (daily) which in turn changes the range and gradient of the inverse decay curve

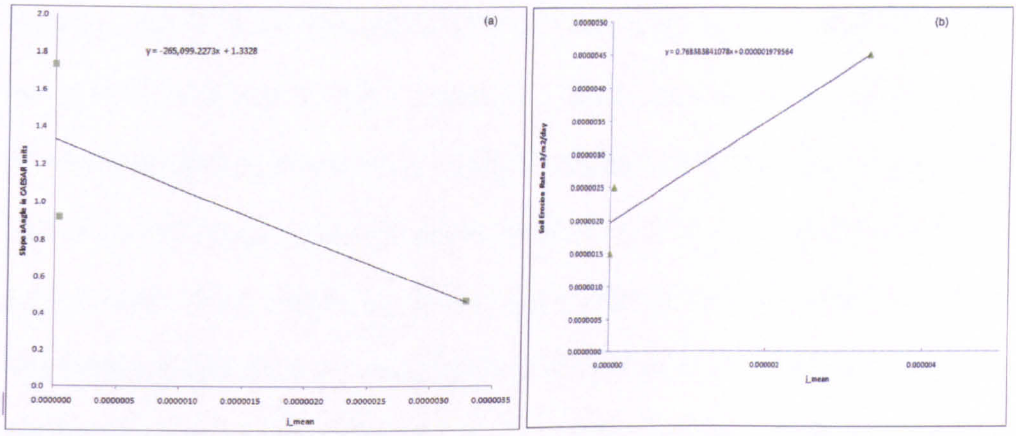

Figure 7.2 *(a) Relationship between j_mean and slope angle (b) relationship between j_mean and soil erosion rate.*

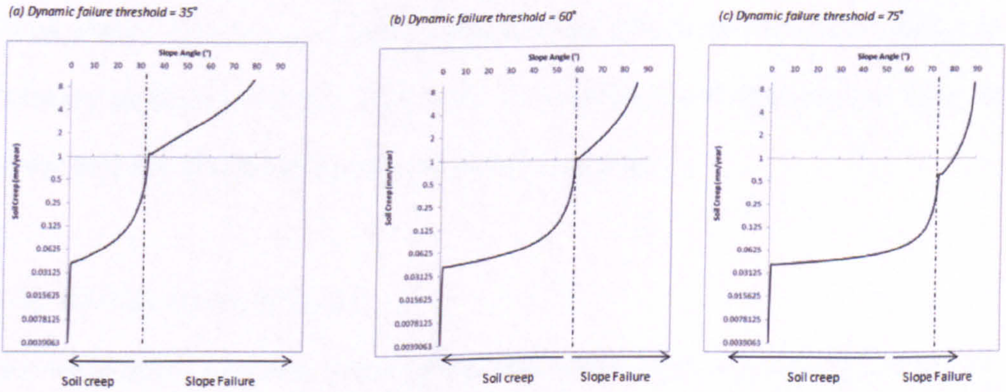

Figure 7.3 *(a) As soil moisture increases, the dynamic failure threshold (DFT) is reduced to 35, therefore more cells will fail than creep, (b), the soil moisture has reduced so DFT has increased to 60, therefore all slopes below 60° will creep at an increasing rate, all slopes above 60° will fail, (c) Soil moisture is greatly reduced therefore DFT is now 75, only cells with a slope angle greater than 75 will fail, all cells with slope angle below this will creep.*

(Fig.7.3a-c). This failure threshold is determined by soil moisture (j_mean) (Fig. 7.2a) and is then applied uniformly across the catchment. The soil moisture parameter is controlled by the *m* value (section 5.5) which imitates the effect that vegetation has on the flood hydrograph; thereby encapsulating the effects of both soil moisture and land use within a single parameter. Thus under wetter conditions the failure threshold is lowered and the rate at which soil creep occurs increases; with the reverse occurring in drier conditions. Above the failure threshold, rates of movement continue to increase at a defined inflection rate based around the rate of increase at the dynamic failure threshold which is extended up to infinity; essentially greater rates of movement occur in steeper cells and under wetter conditions. Imposing a failure threshold and applying soil creep in this manner replaces the original slope failure threshold and mechanism in CAESAR which now becomes defunct. Thus mass movement processes are encapsulated in a single equation and are henceforth referred to as the 'mass movement continuum'. Both CAESAR 5.9 and this new creep function do not deal with differences between creep at the surface or at depths, nor does it consider diffusion and advective behaviour within soil. Instead operating in a more 'black box' style, where losses and gains of sediment occur on a cell by cell basis governed by the slope angle.

7.4.2 Revised soil erosion rates

In CAESAR 5.9 soil erosion rates (SER) were user defined and applied in a spatially distributed manner across the catchment underpinned by the slope component of RUSLE (e.g. Renard et al., 1991), but this excludes precipitation, erodibility and vegetation related parameters. SER affects each cell in the DEM and used an upslope drainage area similar to the RUSLE slope length term which broadly replicates slope wash (Coulthard, pers. comm.). Thus in CAESAR 5.9, both up-slope area (slope length) and slope angle control the spatial pattern of SER. Given that our overall aim is to identify changes in sediment delivery under different precipitation and land use regimes, a logical step was to link the soil erosion parameter to soil moisture (j_mean) (Fig.7.2b). The relationship between soil erosion rates and soil moisture has been scaled on a linear basis. This is similar to adding a P-factor from RUSLE. Thereby land use (vegetation) changes (Prosser & Dietrich, 1995) and water availability (rain or snow melt) would impact on the rates of soil erosion. Linking SER to the j_mean term produces greater erosion at peaks in the flood hydrograph, the intensity of which is governed by both water input and land use (*m* values). Thus SER varies through time, though at this stage spatial patterns of water input and land use change are not incorporated. For the simulations of the Petit lac d'Annecy typical pre-Alpine daily soil erosion rates ($0.000005 m^3/m^2/day$) from empirical studies (Boardman & Poesen, 2006) have been scaled on a linear basis against soil moisture (j_mean) and then factored against the elevation model (slope and slope length) in the existing SER equation. Thus under wetter conditions greater soil erosion takes place and for a similar magnitude and duration wetter period (j_mean) the rate of erosion would be higher for catchment conditions of reduced forest cover (increased human impact) and vice versa.

7.4.3 Model Setup

A DEM of 50m x 50m resolution has been used to represent the topography of the two sub-catchments in the Petit lac d'Annecy: the Ire and the Tamie (Fig. 7.1a-b). A second DEM is used to represent bedrock elevation and therefore constrains and limits soil/sediment depth and availability for erosion. Further detail on the sediment distribution in the catchment is available in chapter 5, but in summary it comprises a thin (1m) sediment/soil cover on the mountain summits, a shallower fill (3m) on the hillslopes thickening at shallower gradients (10m) and on the valley floor (20m) (Fig.7.1c-d). Model simulations were driven using 500 year modelled hourly time series for water input, with a snow mass balance adjusted time series (see chapter 5) produced from precipitation (Pauling et al., 2004) and temperature (Luterbacher et al., 2006) data that were both scaled to Petit lac d'Annecy catchment averages. Land use data impact on CAESAR simulations by varying the *m* parameter, a modification of a key parameter used in TOPMODEL (Beven & Kirkby, 1979). *M* varies the recession limb of the flood hydrograph imitating the affects of vegetation on water movement and water storage. Hourly land use data were interpolated from arboreal/non-arboreal pollen (AP/NAP) ratios (Jones et al., in prep.), calibrated against contemporary forest cover discerned from satellite imagery (Google Images, 2008) and converted to *m* values (see chapter 5). The appropriateness of the range of *m* values was assessed through comparison of modelled and observed water discharges, and informed scaling the of palaeovegetation data (section 5.5).

7.4.4 Matrix of CAESAR simulations

In order to test the sensitivity of the hillslope processes, a series of simulations (Fig.7.4) were ran to test the behaviour of hillslope process rules as applied in the most recent version of CAESAR (version 5.9: Van de Wiel et al., 2007). A second series of simulations with the new parameterisation of hillslope processes implemented were then run to assess model performance. CAESAR has the facility for simulations to be undertaken excluding certain processes (e.g. just fluvial processes on their own or fluvial plus soil creep but without SER). The input time series: precipitation and *m* value (land use) were identical for all simulations unless otherwise stated; with the precipitation series pre-adjusted for snow storage (c.f. Welsh et al., 2009).

1. For the Tamie sub-catchment, a valley in part with a large, wide, flat floodplain with steep sides (Fig.7.1a) both the original CAESAR 5.9 (A2 & A3) and the new CAESAR (D1 & D4) were used. The objective was to test the sensitivity of each slope process individually alongside the fluvial model (e.g. fluvial + mass movement and fluvial + SER) and to assess their impact on the geomorphology and sediment movement across the catchment.

2. For the Tamie, the integrated hillslope processes (e.g. fluvial + creep + SER) were tested for original CAESAR (A4) and enhanced CAESAR (D6). Further simulations for 33% increased precipitation (D7) were made in order to assess the response of enhanced CAESAR under different climatic conditions.

3. For the Ire, a valley flanked by steep slopes and with less opportunity for overbank sediment storage (Fig.7.1b) an identical simulation to D6 was carried out (E6) to assess how the modified hillslope processes operated and interacted with the fluvial system for a valley with a different morphometry (Welsh et al., 2009).

All forms of CAESAR output (section 7.5) were used to assess how realistic our parameterisation of hillslope processes has been in terms of the spatial pattern and rates of erosion and deposition across each DEM, the total sediment discharge, and the pattern of change in response to variations in both water input and land use. Firstly the slope processes were tested individually and then in combination (section 7.5) in the Ire and Tamie sub-catchments to identify system response to land use change.

Simulation Name	Fluvial Erosion and Deposition	Slope Failure Angle*	Soil Creep (mm/year)	Soil Erosion Rate (mm/year)	Testing Description	Sensitivity Details	Total Sediment (m3)	Total Suspended Sediment (m3)
A1	ON	89	OFF	OFF	Fluvial erosion & deposition only	Original Ppt	354191.11	99637.73
A2	ON	89	0.007	OFF	Creep test	Original Ppt	751556.78	155613.36
A3	ON	89	OFF	0.00005	Soil erosion test	Original Ppt	317311.19	102575.92
A4	ON	89	0.007	0.00005	Creep & Soil erosion	Original Ppt	559299.21	165552.57
D1	ON	N/A	Advanced DFT	OFF	Creep & New J_mean	Original PPT	899393.61	325632.92
D2	ON	N/A	Advanced DFT	OFF	Creep & New J_mean	Ppt x1.33	1296953.85	451877.10
D4	ON	N/A	OFF	Advanced SER	SER & New j_mean	Original PPT	1169752.12	362962.52
D5	ON	N/A	OFF	Advanced SER	SER & New j_mean	Ppt x1.33	1247126.03	437158.45
D6	ON	N/A	Advanced DFT	Advanced SER	Creep &SER New j_mean	Original PPT	1116885.96	473270.40
D7	ON	N/A	Advanced DFT	Advanced SER	Creep & SER New j_mean	Ppt x1.33	2601659.68	892297.67
E6	ON	N/A	Advanced DFT	Advanced SER	Creep & SER New j_mean	Original Ppt - IRE	1789768.55	543446.87

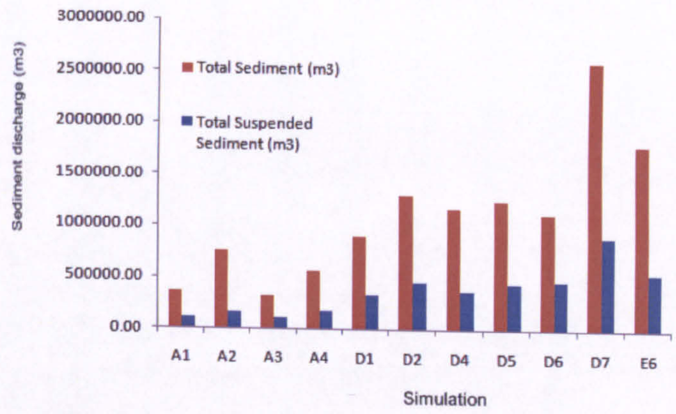

Figure 7.4 *(a) Simulation matrix and description, (b) Total sediment yield and total suspended sediment yield for each simulation.*

7.5. Testing the slope processes

7.5.1 Sediment yield

The output of simulations that exclude certain processes are a little contrived: for example, one could simulate creep processes (e.g. passing materials downslope) but have local depo-centres that are not flushed by SER (slope wash) processes. Nevertheless simulations with each of the slope processes switched off in turn for CAESAR 5.9 (A2 and A3) and the new version (D1 and D4) facilitate some assessment of sediment movement rates on the hillslopes using time-stepped DEMs (sections 5.2 and 5.3). However, the total sediment yields leaving the catchment (Fig.7.4b) should perhaps be regarded with caution. That said, simulations of enhanced mass movement and SER with fluvial processes only produced 2x and 3.6x increases in sediment yield respectively. Operating alongside the fluvial processes in CAESAR at a 50m grid resolution, SER (D4) appears to produce more rapid and uninterrupted sediment routing; in fact D4 appears more rapid and less interrupted in this regard for the integrated mass movement and SER simulation (D6). Similar circumstances apply for equivalent simulations using CAESAR 5.9, but vary with changes in the uniform rates applied for soil creep and SER.

7.5.2 Spatial pattern of mass movement

Time-stepped DEMs from CAESAR output allow the spatial pattern of sediment movement to be assessed through time. These data provide a more logical method for assessing the appropriateness of the simulation of hillslope processes than just sediment yield alone. Here the CAESAR simulations that just include fluvial and soil creep or mass movement processes, with no SER, are instructive because the rates of changes on the slopes away from the active fluvial zone are of interest. Thus to examine the movement of materials on the slopes (in the net erosion/deposition output grids) (Fig. 7.5) we have focused on particular parts of the DEM; steep areas for failure style behaviour and more gentle slopes for creep processes.

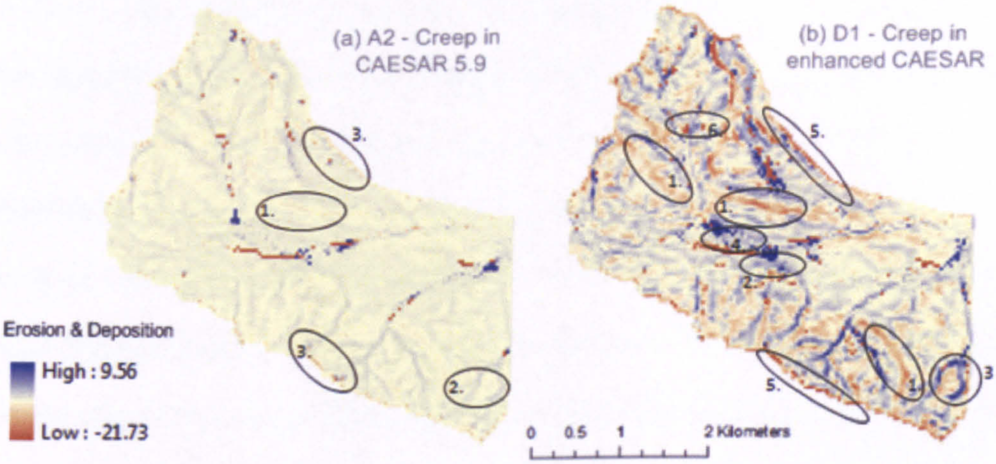

Figure 7.5 *(a) Net erosion and deposition over 500 years for simulation A2(creep processes in CAESAR 5.9), (b) Net erosion and deposition over 500 years for simulation D1 (enhanced creep processes)*

In CAESAR 5.9 (Area 3: Fig.7.5a) slope failure was restricted to cliffs (in A1-4 this was restricted to slopes >89%), and there the 0.5m of erosion over 500 years is reflective of the arbitrary upper limit (0.001m per timestep) and the time step.

Lower failure thresholds (not shown) have been tested, and they produce similar amounts of erosion and deposition but for a larger area of the DEM. With creep and slope failure replaced by the new mass movement continuum in CAESAR (D1, Fig.7.5b), the rates of change are up to ~2m of erosion in watershed zones (Area 5: Fig.7.5b). Locally around areas with repeated slope failure (Area 6: Fig.7.5b) up to ~6m of sediment has been lost from some slopes. These rates appear realistic for mountain environments where debris flow activity is widespread and are in keeping with observations for the Petit lac catchment (Foster et al., 2003). These simulations remain constrained by the pass-the-parcel nature of sediment movement to adjacent cells.

In terms of the distribution of erosive and depositional activity, in the CAESAR 5.9 simulations, slope failure has been limited to the cliffs (Area 3: Fig 7.5a) and creep takes place slowly with the pattern governed by slope angle (Area 3: Fig.7.5a). Deposition of material occurs in the gullies (Area 2: Fig.7.5.a) but at very low amounts ~2-3cm over the 500 year period. With the new model processes (D1: Fig.7.5b), more rapid movement occurs on the steepest slopes (Area1: Fig.7.5b) and on the ridge crests (Area 1:Fig.7.5b). Deposition is focused towards the bottom of slopes (Area 2: Fig. 7.5b) and areas where the gradients are reduced, for example in the gullies or valleys (Area 3: Fig.7.5b). CAESAR 5.9 simulations (A2, Fig.7.5a.) of soil creep applied a constant rate of 0.007 mm/yr (factored against slope angle) and produced net changes in elevation of -20cm (0.4mm per year) in low angle shedding areas and +10cm (0.2mm per year) on deposit slopes. Note the accumulation and deposition figures are net and provide no information about throughput, i.e. materials lost to the river or down slope. For the new mass movement equation (D1, Fig.7.5b) these rates have increased by an order of magnitude with net changes in elevation of -1.2m for shedding areas and +0.5m of deposition on deposit slopes. These rates equate to 2.5mm yr-1 and 1mm yr-1 respectively on average. The zones for failure were identified by slope angle, and the zones for identifying creep were on gentler gradients with reduced flow paths (e.g. not in a gully where fluvial processes may dominate). In summary, the mass movement processes were more intensive on steeper slopes, and both the area affected and the amount of material moved appear realistic (Area 1:Fig. 7.5b, Area 5: Fig. 7.5b). Field evidence suggests that the rates from mass movement simulations are typical of those in the Petit lac d'Annecy catchment (Foster et al., 2003). Large zones of sediment accumulation occur in the alluvial fans (Area 4: Fig 7.5b) and at tributary junctions, but are absent from the A2 simulation (Fig. 7.5a) which suggests that there is a reduced sediment supply in A2 with original creep and failure in CAESAR 5.9.

7.5.3 Spatial pattern of soil erosion processes

For the new model, soil erosion rates (SER) were amended so that they also incorporated response to the soil moisture term. Similar to the investigation of mass movement (section 7.5.2), time-stepped DEMs have been analysed for the spatial pattern of erosion and deposition.

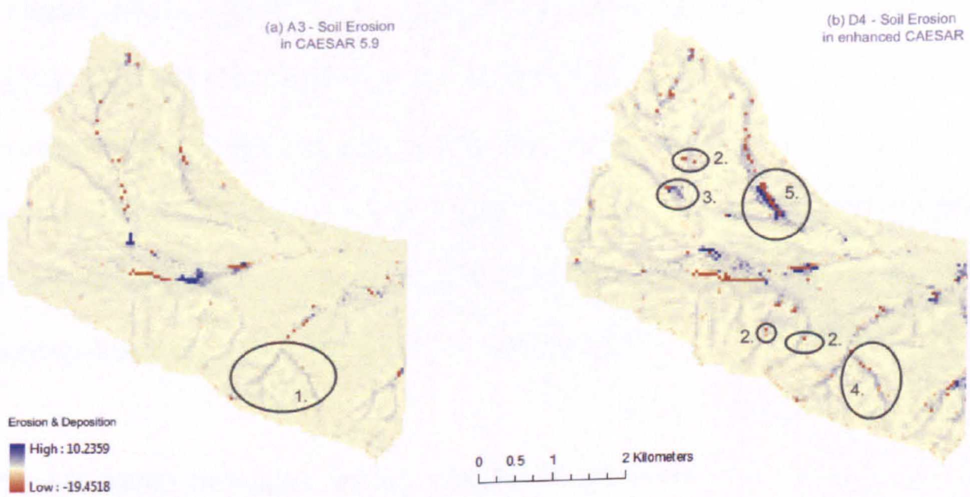

Figure 7.6 *(a) Net erosion and deposition over 500 years for simulation A3 (soil erosion processes in CAESAR 5.9), (b) Net erosion and deposition over 500 years for simulation D4 (soil creep processes)*

In CAESAR 5.9, SER has most prominently affected channel networks and gullies (larger upslope areas) in settings up-system of the fluvial zone (Fig.7.6a), where net changes in elevation of +30cm (0.6mm yr-1) and 50cm (1mm yr-1) have occurred. Comparatively minor amounts of net erosion (-5cm or 0.5mm yr-1) or deposition (+15cm or 0.3mm yr-1) have occurred elsewhere on the hillslopes (Area 1: Fig.7.6a). Under the new model (D4, Fig.6.6b) there are locations with net elevation changes of up to ~3m (6mm yr-1) erosion (Area 2: Fig.7.6b) and ~2m (4mm yr-1) deposition (Area 3: Fig.7.6b) on ridge crests. Typical rates of net hillslope erosion in D4 over 500 years is a loss of 3cm-30cm on the main slopes, and an additional ~5cm – 50cm net on deposit slopes. This equates to 0.06mm-0.6mm erosion per year depending upon slope angle. These values are similar to empirical evidence for Alpine zones (Boardman et al., 2004). There is up to ~0.5m (1mm yr-1) deposition of sediment in gullies in D4 (Fig. 7.6b) which is in the same order of magnitude as A3 (Fig.7.6a), but amounts of erosion are increased to ~2m (4mm year-1) in D4 (Area 4: Fig.7.6b) which is much higher than A3 (Fig.7.6a) where there was virtually no gully erosion taking place. The enhanced soil erosion processes have locally produced stronger coupling between the hillslopes and the fluvial system, for example the tributary stream (Area 5: Fig.7.6b) which is a feature of the contemporary Petit lac catchment (field evidence) but was not recognised in the A3 simulation (Fig.7.6a) due to the process being represented too simply in CAESAR 5.9.

7.6. Modelling integrated hillslope and fluvial processes

Welsh et al. (2009) concluded that there was scope for improvement of hillslope process representation within CAESAR as initial modelling showed that the hillslope processes were not represented well enough to fully capture the role of changes in land use. The behaviour of individual hillslope processes have been tested and described in section 7.5. Here, the combined hillslope processes (both creep and SER) and fluvial processes are all switched on for simulations in the Ire and Tamie sub-catchments. This allows the results of the sediment discharge from the combine process simulations to be assessed and the sensitivity of the landscape to changes in land use to be explored.

7.6.1 Spatial patterns of erosion and deposition

The movement of sediment on slopes, for simulations incorporating all slope and fluvial processes for the CAESAR 5.9 (A4) and the new model (D6, E6, D7) can be assessed from the patterns of erosion and deposition displayed on net erosion/deposition maps (over 500 years) (Fig.7.7). Mass balance changes along key longitudinal transects (Fig 7.8) can also be a useful method of investigation into behaviour of hillslopes. These transects have been selected to represent different morphologies of slope and they show net changes in erosion and deposition over the 500 year simulation.

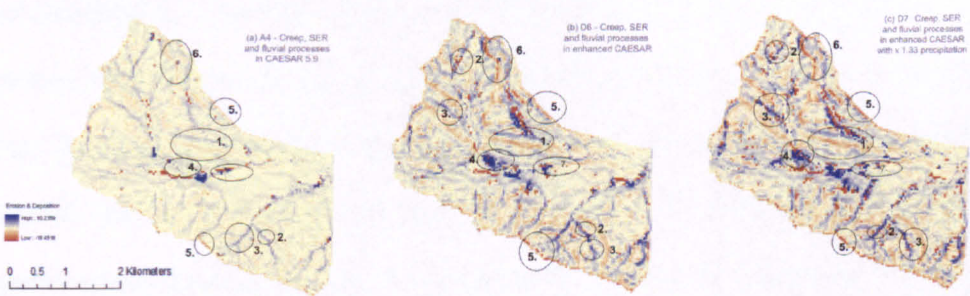

***Figure** 7.7 (a) Net erosion and deposition over 500 years for simulation A4 (all slope and fluvial processes in CAESAR 5.9), (b) Net erosion and deposition over 500 years for simulation D6 (all slope and fluvial processes), (c) Net erosion and deposition over 500 years for simulation D7 (all slope and fluvial processes with x1.33 precipitation in simulation)*

7.6.1.1 Hillslopes

Realistic rates of erosion and deposition and sediment yields do not occur in CAESAR 5.9 due to under development of hillslope processes. Rates of soil creep (net erosion and deposition) in simulation A4 are around 15cm over the 500 year period (Area 1: Fig.7.7a) compared to 60-120cm in D6 (Area 1: Fig.7.7b). Increased precipitation (D7, Fig.7.7c) seems to have little effect on the overall amount of creep erosion and deposition that takes place, as this is also around 60-120cm (Area 1: Fig.7.7c). Activity in hillslope gullies was not extensive in A4 with typical rates at +/- 0.5m (Area 2: Fig.7.7a), yet this increased by an order of magnitude to ~1.5m-8m in D6 (Area 2: Fig.7.7b) and a further increase in maximum erosion values to ~12m occurs when precipitation is increased by 33% (Area 2: D7, Fig.7.7c). Gully deposition is only ~0.5m in A4 (Area 2: Fig.7.7a), yet this level of storage increases to ~0.75m – 2m in D6 (Area 3: Fig.7.7b) and is once again increased to maximum values of ~4m deposition in gullies in D7 (Area 3: Fig.7.7c) when precipitation is increased by 33%. Failure zones are all situated on the steepest slopes of the catchment and ~15cm of erosion takes place in A4 (Area 5: Fig.7.7a), increasing substantially to ~1m-2.5m of erosion in D6 (Area 5: Fig.7.7b) with maximum erosion values across the three simulations peaking at around 3m in D7 (Area 5: Fig.7.7c). Other slope failures occurred in parts of the catchment that are susceptible to rock and snow avalanches (Area 6: Fig.7.7a) and sediment movement of approximately 25cm of erosion which increase by an order of magnitude to between 50cm and 2.5m of erosion in both D6 (Area 6: Fig.7.7b) and D7 (Area 6: Fig.7.7c) are not unrealistic.

Overall there was significantly more sediment moved in the new model simulations (D6, D7) compared to CAESAR 5.9 (simulation A4). The rates of hillslope processes are amplified under enhanced CAESAR (D6) and amplified further when there is an increase in precipitation (D7), this is particularly true for the ridges (Area 1: Fig.7.7a-c), and for the failure zones. Slope failure is particularly interesting as the rates increased in enhanced CAESAR simulations (D6, D7) both at the watershed boundary (Area 5: Fig.7.7a-c) and for localised slope failures (Area 2: Fig.7.7a-c). The magnitudes of erosion were even greater in D7 (Fig.7.7c) and there were an increased number of cells that had this level of high magnitude erosion (Area 2 & 3:Fig.7.7c). These rates seem realistic when compared with field evidence from the Petit lac d'Annecy catchment and are representative of hillslope process behaviours (Foster, 2001; Foster et al., 2003; Jones et al., in prep).

In addition to realistic rates of hillslope processes, it was important to ensure that the spatial patterns of sediment are behaving in a similar manner to the 'real-world'. A series of five downslope longitudinal profiles (Fig. 7.8) were derived from the net erosion and deposition grids for the Ire sub-catchment for 500 years of simulation (E6: new model processes combined with normal precipitation). The profiles that were selected represent different lengths and zones of the slopes, with the focus on slopes rather than channel settings. All of the profiles show that steeper zones have greater erosion; profile 1-2 shows 40-50cm net erosion (Fig.7.8) and a downslope aggradation of up to ~ 50cm at either the foot of the profile or on a part of the slope where the gradient becomes shallower e.g. Profile 2 at 200m and Profile 5 at ~75m-125m. Profile 5 is a rather stepped profile whereby sediment is removed from the top of the slope and deposited further downslope where gradients are shallower. This

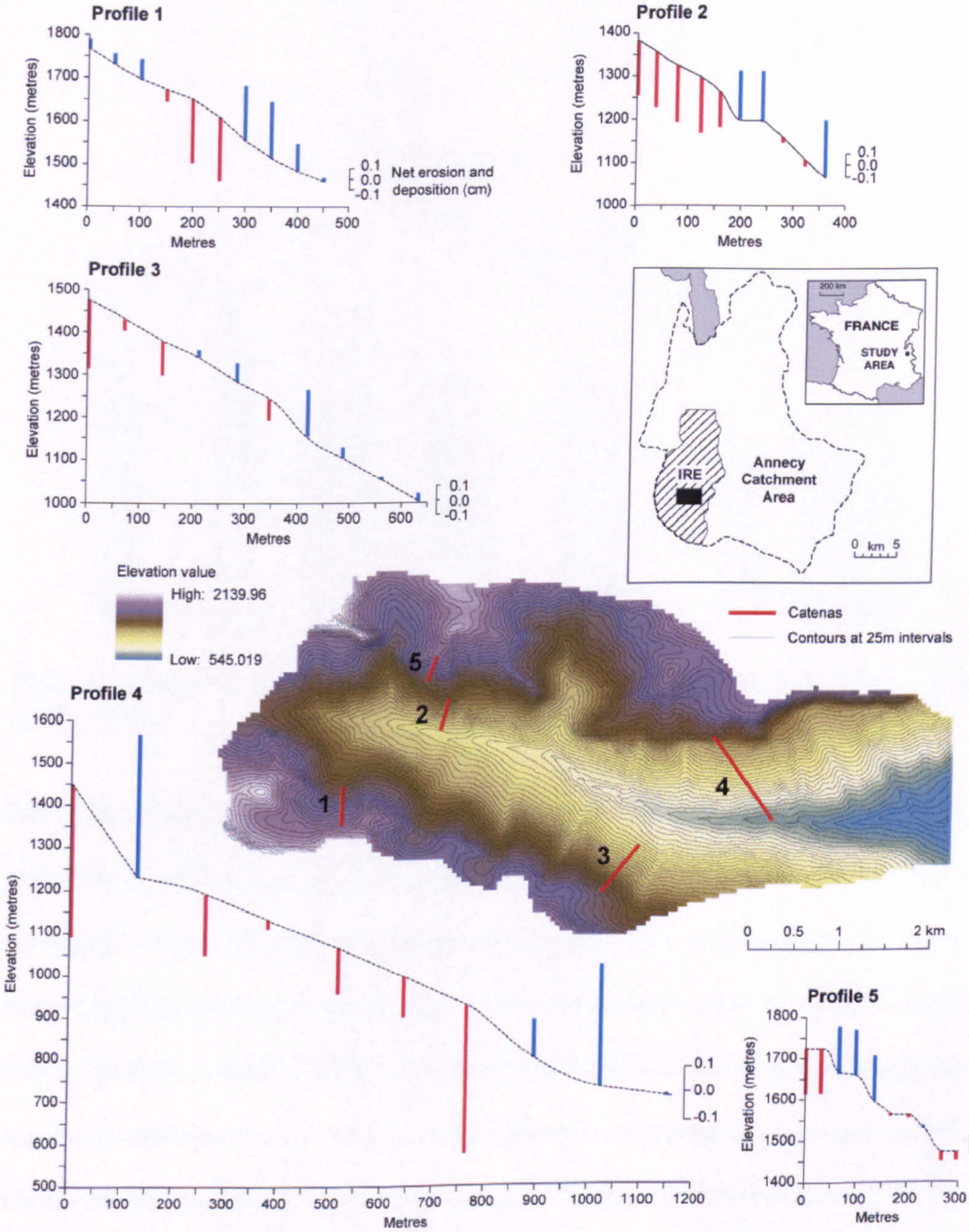

Figure 7.8 – *E6 simulation for the Ire sub-catchment. Five profiles across the slopes have been taken (main location map) and erosion and deposition have been analysed across the profiles.*

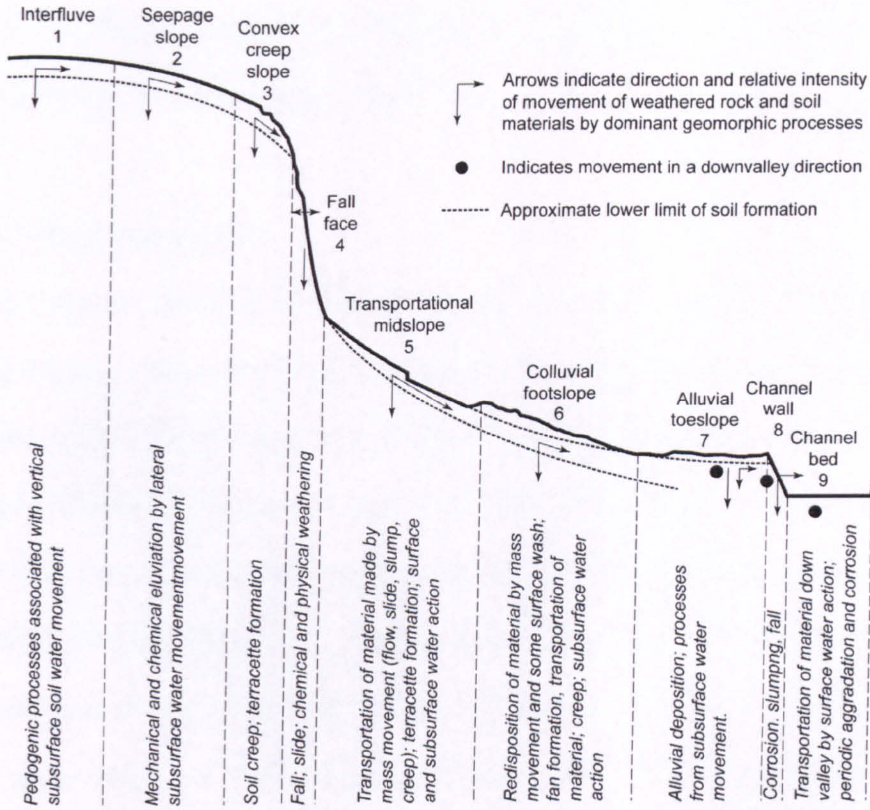

Figure 7.9 *Hypothetical nine-unit landsurface model (After Dalrymple et al. 1968) (Selby, 1993)*

behaviour is also exhibited in profile 4. In order to assess whether this modelled behaviour is reflective of the real world, it is possible to compare these profiles to a theoretical model of slope behaviour, constructed from empirical observations. Selby (1993) constructed a hypothetical nine-unit land surface model (Fig.7.10) after Dalrymple et al., 1968) in which some but not all units can be recognised along the simulated hillslope profiles. Many of the units refer to sub-surface processes which are not readily identifiable in CAESAR, however three of the notable slope units; (4) fall face, (5) transportational midslope and (6) colluvial footslope zones (Fig.7.9) are exhibited within the CAESAR model output. Slide/fall (Fig.7.9) is exhibited in profile four and five. The transport zone is most shown clearly in profiles one, two and three (Fig.7.8) and Selby/Dalrymple's 're-deposition of material' unit is exhibited in profiles two, three and four.

7.6.1.2 The fluvial system

Storage patterns also change between simulations; particularly in the fan zone (Area 4: Fig.7.7a-c). There is little fan storage in A4 other than in the main channel (outside of area 4), yet in D6 and D7 (Fig.7.7b-c) there is a large zone of sediment storage (~25000m^2) with up to ~5-10m of fill. However, the increased precipitation in D7 has increased water capacity in the channel, and some fan head incision of 6m-7m has begun (Area 4:Fig.7.7c). A4 shows that there is more incision in the main channel (Area 4: Fig.7.10a) than in the new model (Area 4: Fig.7.10b-c) which is likely to be due to the lack of fresh sediment supply from the hillslopes in A4 therefore the main channel is eroding. The new CAESAR model simulations show progressively less incision (Area 4: Fig.7.10b-c) as precipitation increases, suggesting that increased rates of hillslope erosion have provided a greater supply of sediment and therefore less channel erosion and more in-channel deposition occurs in D6 and this is further enhanced in D7 where there is much less erosion and increased deposition occurring at this part of the channel. This suggests that sediment supply is exceeding the capacity of the channel here.

The connectivity of these systems varies at different places on the hillslope and under the three simulations A4, D6 and E6. Typical hillslope profiles are a feature of D6 and E6 (Area 1: Fig.7.10b-c) with erosion on steeper zones and deposition at the foot of slopes, rendering a store of sediment available to the tributary channel in D6 (Area 1: Fig.7.10b) and to the main channel in E6 (Area 1: Fig, 7.10c). There are small amounts of deposition at the foot of slopes in A4 (Area 1:Fig.7.10a) yet very little erosion has taken place upslope, therefore the slopes in this simulation were not as well connected to the tributary here as D6.

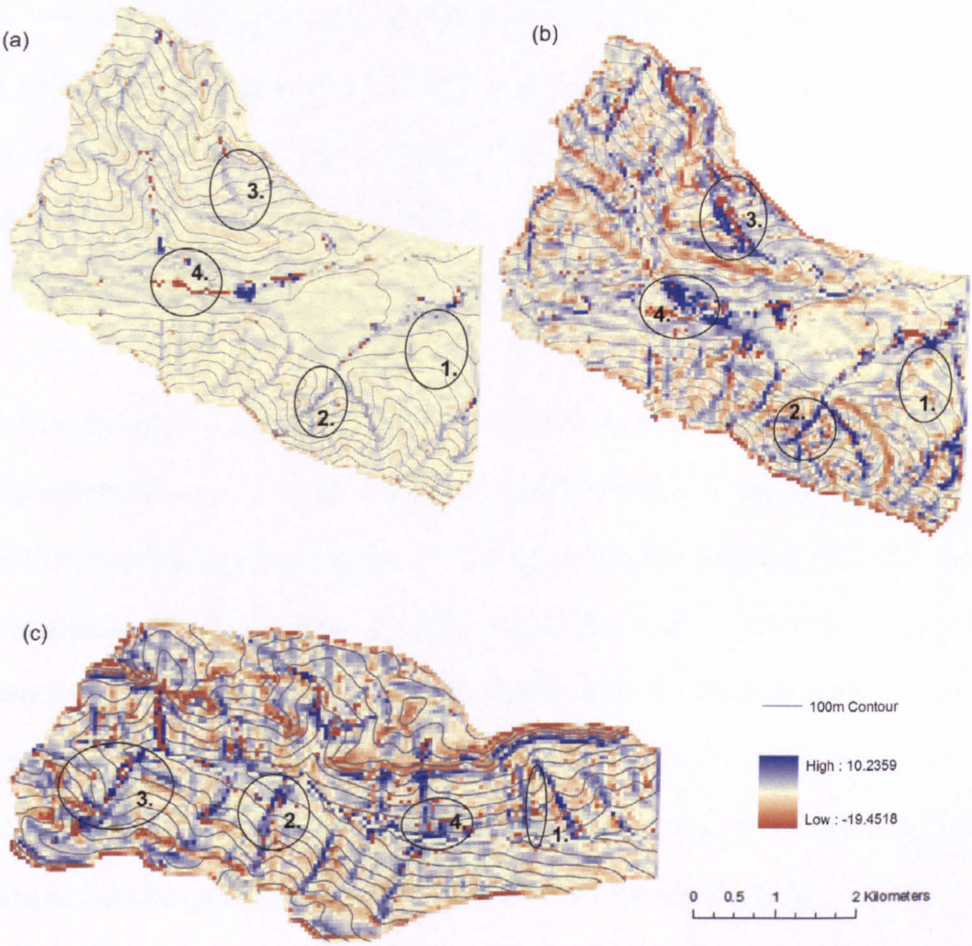

Figure 7.10 *(a) Net erosion and deposition in Tamie A4, CAESAR 5.9, (b) Net erosion and deposition in Tamie, D6, enhanced CAESAR, (c) Net erosion and deposition in Ire, E6, enhanced CAESAR.*

Gullies or tributary stream systems (Area 2:Fig.7.10a-c) are a major zone of storage in the Tamie, and there is significantly more storage of sediment in D6 than A4; a function of increased availability of sediment from the slopes moving into these storage zones. Hillslopes are also fairly well connected to the tributary in both simulations (D6 and A4) which allows delivery of sediment to the channel. Whilst E6 does have similar amounts of storage in gullies as D6, the gullies here also show that some erosion is taking place (Area 2: Fig.7.10c) which is not a characteristic of the Tamie gullies. There are many gullies in the Ire (E6), all of which are well connected to the main channel as there are large depo-centres at the foot of slopes. Field evidence shows that the channel and slopes are well connected in the Ire.

The larger tributary in the Tamie that developed in D6 (Area 3: Fig.7.10b), is much more restricted in A4. There were some small amounts of deposition (Area 3: Fig.7.10a) and limited development of a very small channel/gully, but this was under-developed when compared to D6 (Area 3: Fig.7.10b). There the simulated channel was much better developed with incision of up to 10m taking place on the steeper parts. This allows this zone of the Tamie to become active whereas this part of the catchment is relatively dormant in A4. The development of this tributary stream has encouraged greater delivery of sediment to the channel in E6.

Fan zones are a big area of sediment storage for D6 and E6, yet there is almost no storage in A4 (Area 4 Fig.7.10a-c). For D6 and E6 there are large zones of sediment accumulation at the foot of the slope (Area 4, Fig.7.10b-c) which lie above the main channel yet these could be re-mobilised during high flood magnitudes. There are many alluvial fans in the Ire (E6) all of which are truncated and trimmed by the axial stream; large amounts of sediment accumulate at the foot of these gullies and are available for transport via the main channel (Area 4, Fig.7.10c).

Overall there is more opportunity for storage of sediment and particularly overbank deposition in the Tamie than in the Ire due to the morphometry of the catchment. There are greater numbers of deeply incised tributary gullies feeding tributary junction alluvial fans in the Ire compared to the Tamie which operate as a significant temporary store for sediment but become a source of sediment during high magnitude flood events.

7.6.2 Comparing the new and original CAESAR

The main aims of implementing new hillslope processes is to allow realistic process rates and realistic patterns of erosion and deposition to occur in the model so that the sensitivity of landscapes to changes in land use can be explored more fully. In addition to this, it is vital that CAESAR has adequate process representation to produce sediment yields that are comparable to lake sediment records which is a fundamental aim of this thesis. Therefore, here the sediment discharge records for A4 and D6 are presented in a manner that is comparable to lake sediments and catchment geomorphology to assess how realistic the new processes are.

Implementing the new slope processes (mass movement and soil erosion rate) for the Tamie has doubled the amount of total sediment over the 500 year simulation compared to A4 (Fig. 7.4). In order to identify what impact these modifications will have when compared to lake sediment accumulation rates, an estimated hypothetical lake sediment accumulation rate was calculated for each from the suspended sediment grain sizes. Over a 500 year period (run-up excluded), there would be approximately ~0.08m of accumulation in the lake for A4 and ~0.4m for D6. Given that these are estimates from only one of the five sub-catchments in the Petit lac d'Annecy, approximate calculations suggest that if the rates were the same in all sub-catchments this would equate broadly to 0.4m accumulation for A4 and 2m of sediment accumulation in D6 over a 500 year period. The lake sediment accumulation rate for LA13 (Jones et al., 2001) equates to 1.96m of sediment over the last 500 years. Clearly, the improved hillslope processes (D6) presented here produce a similar rate of sediment discharge to that of the LA13 core but more importantly they are within the same order of magnitude, unlike the original slope processes (A4).

There are no differences between these two simulations except that in D6 the soil moisture term is allowed to exert control on the rates of both mass movement and SER. Thus the increase in total sediment flux reflects slope processes varying with temporal changes in both land use (e.g. sparser vegetation cover) and hydrology (e.g. wetter/increased frequency of extreme precipitation events). In the A4 simulation slope processes are not coupled with soil moisture, and do not respond to changes in either land use or hydrological input. D6 and E6 produce sediment discharge time series that are comparable with a lake sediment record e.g. continuous sediment discharge through the 500 year period (Fig. 7.11) whereas there are frequent gaps in the sediment discharge record in A4 due to greatly reduced sediment supply from the hillslopes. This alteration to the simulated sediment discharge reflects the new representation of mass movement and soil erosion processes.

7.6.3 Landscape sensitivity to land use change

One of the key aims of this chapter is to identify how land use change impacts on sediment delivery over time. Changes in modelled sediment flux AD 1500-1700 can be largely attributed to either catchment morphometry (storage and release) or precipitation drivers as land use during this period remains unchanged at around 55% until around 1725 when there was a 20% drop to 35% cover (Fig.7.11h). A second reduction in forest cover takes place around AD 1885. Despite the low forest cover 1725 AD – 1805 AD, this 100 year period represents the lowest period of sediment discharge in the 500 year time series for both catchments (Fig.7.11a-b). It is likely to be a result of low total annual precipitation (Fig.7.11e) during this period which reduces sediment supply and water discharge therefore leading to lower erosion rates and reduced channel capacity. Low erosion rates result in a reduced supply of sediment to the system thereby reducing the ability of the stream to transport sediment out of the catchment.

After 1800, there are a series of years with high total annual precipitation (Fig.7.11e) e.g. 1837, 1881. In addition, discharge statistics (Fig.7.11f-g) show that although 1850 did not have a high total annual precipitation, there was an extremely large flood event during this year which explains the increase in sediment discharge during this year. During these years of increased precipitation, with the new model processes, there was an increase in sediment supply due to increased erosion rates and also a greater channel capacity due to increased flood magnitudes. The high precipitation events show a large increase in sediment delivery in the Ire almost immediately (Fig.7.11b), but the sediment discharge record for the Tamie

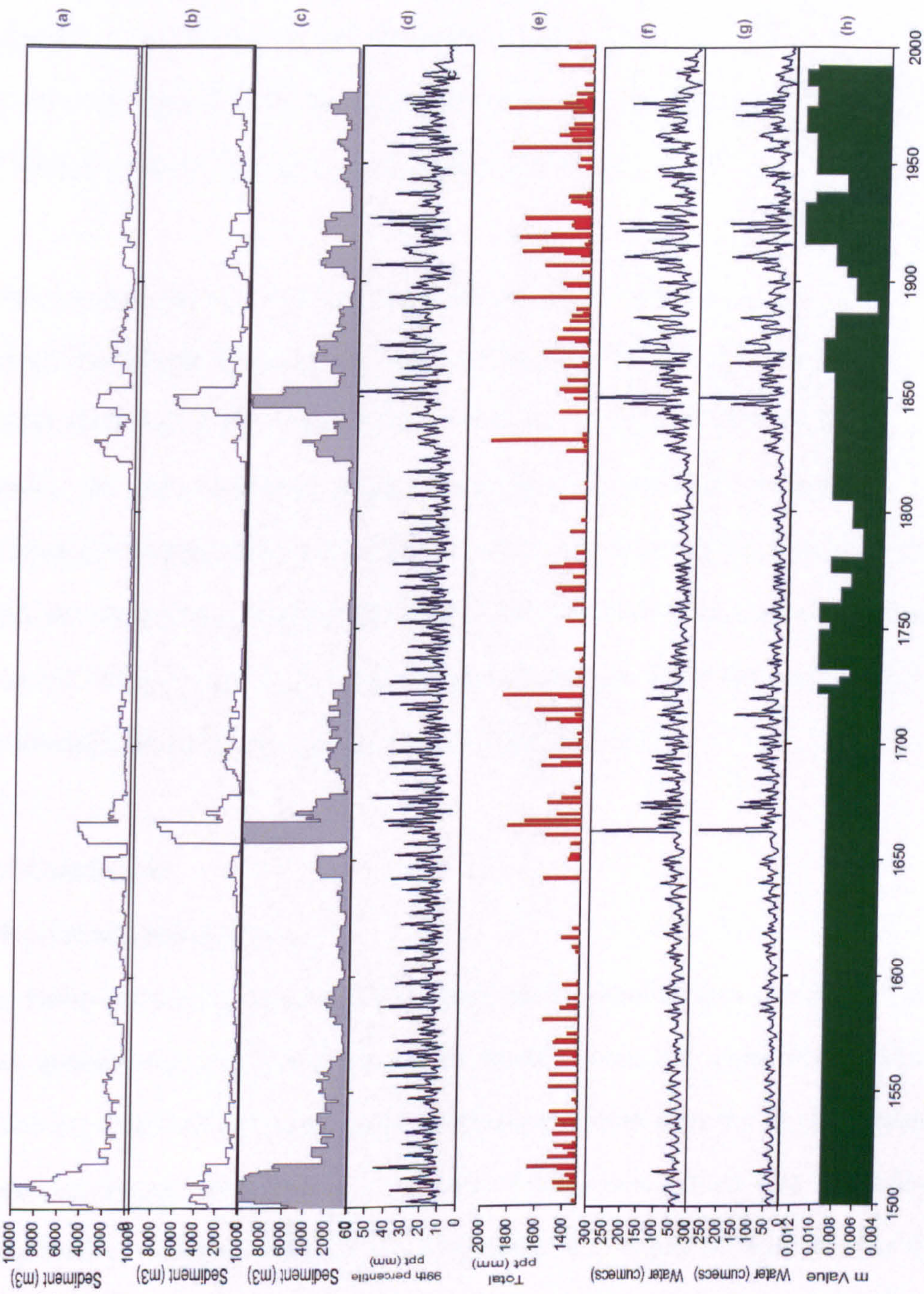

Figure 7.11 *(a) Total sediment discharge Tamie D6, (b) Total sediment discharge Ire E6,(c) Combined total sediment discharge Ire and Tamie (D6 & E6), (d) 99th percentile precipitation, (e) Total annual precipitation, (f) 99th percentile water discharge Tamie, (g)99th percentile water discharge Ire, (h) Land use values (m)*

shows that there are several smaller peaks (Fig.7.11a). This suggests that the Tamie may have responded to the first event, but a legacy of sediment may remain in and around the channel to be transported out of the catchment during the next event. This is confirmed in the net erosion/deposition maps (not shown).

The reduction of forest cover around AD 1880 (Fig.7.11h) corresponds to a period after which a large amount of sediment was flushed out of the catchment. The time series show that there is no immediate increase in sediment delivery during this period, but there is perhaps a lag time in both sub-catchments as there are two increases in sediment delivery around AD 1900. During the period 1900AD – 2000 AD, the catchments remain largely forested, and two small peaks in sediment 1907 and 1927 (Fig.7.11a-b) are the major features of this time series that are likely to be attributed high precipitation events in these years (Fig.7.12d).

7.7 Conclusions

This chapter concludes that:

(1) Following the developments of hillslope process representation, both the rates and spatial patterns of hillslope processes are now more comparable to empirical evidence, field evidence and lake sediment accumulation rates for the pre-Alpine zone (e.g. Foster, 2001; Foster et al., 2003, Dearing et al., 2001) than before the modifications outlined in this chapter were made. The improved representation of hillslope processes allows a better assessment of the role that forest cover has in governing sediment delivery out of the system as the processes are now linked to soil moisture and slope angle.

(2) The spatial patterns of net erosion and deposition in addition to the slope profiles show emergent behaviour within sediment such as alluvial fans and down-slope profiles resulting from small-scale process law interaction.

(3) Sensitivity testing with increased precipitation series (33% more) has show that gully incision can increase by around 50% (8m in D6 to 12m in D7). Failure rates are broadly unaltered under more precipitation conditions, but the area affected by failure covers a greater extent. Additionally, some level of fan-head incision took place in the increased precipitation simulation, suggesting that the channel capacity is much greater and exceeds supply, therefore erosion of stored sediment on the fan occurs). In the main channel for D6 and D7 there was much less incision than in A4 where sediment supply was essentially limited. This again shows the importance of the balance between capacity and supply and the level to which it determines whether a system erodes or deposits.

(4) Over the 500 year period, there was 1.6 times more total sediment discharge in the Ire than the Tamie. The spatial maps of net erosion and deposition suggest that the Ire is a much better connected system, with well-coupled gullies to the main channel providing more sediment discharge out of the catchment. In addition to this, the gullies in the Ire erode, a feature which does not occur in the Tamie. It could be suggested that the Ire is extremely efficient at moving sediment out of the system, therefore rendering the system 'supply-reduced' and consequently incising.

(5) The legacy of sediment within a channel, or indeed the availability of sediment for re-mobilisation later in the time-series is important in determining future periods

of increased sediment delivery over decadal to bi-centennial time periods. Implications of this suggest that any simulations of future sediment delivery should have at least several hundred years of hind-casting to account for any storage and re-mobilisation of sediment.

(6) The temporal role of forest cover could not be as fully explored as hoped over such short time scales, largely due to the fact that during the lowest period of forest cover, total annual precipitation was also low, and therefore there were no increases in sediment delivery at this time. There was only one other reduction in forest cover which was precluded by several large precipitation events, which suggests that much of the channel sediment and sediment available for re-mobilisation had been moved out of the system in previous years. However, the main conclusion from the temporal series suggests that if total annual precipitation is low, there will be no increase in sediment delivery regardless of forest cover. Additionally, a series of high magnitude precipitation events may have flushed a large amount of the sediment out of the catchment, which led to a limited supply of sediment available to the channel for transporting out of the catchment. The reduction in forest cover may have increased erosion rates and produced a lagged response in increased sediment delivery 20-30 years later, once again re-enforcing the idea that the history of sediment delivery plays an important role on determining future responses.

In order to investigate the role of forest cover more fully, longer multi-scenario simulations (similar to that of precipitation demonstrated here) are likely to be more useful, as any response will be a function of forest cover alone if all other variables are held constant. Chapter 9 explores these long-term multi-scenario simulations in more detail. The results presented in chapters 8, 9 and 10 make use of the new hillslope processes outlined within this chapter.

Chapter 8

Landscape instability in the pre-Alpine zone under simulated 'snow-free' conditions.

8.1 Introduction

Snowfall and winter snow cover are significant components of the annual hydrological cycle in mountain regions and these areas are particularly vulnerable to projected increasing temperatures as a result of climate change. Typical seasonal characteristics of snow cover are the storage of winter precipitation followed by snow melt in the spring when temperatures increase (Beniston, 2003). The likelihood of snow accumulation and duration of the cover throughout a season increases with altitude due to lower temperatures at high altitudes. However, as temperatures increase, the seasonality, distribution and amount of precipitation in these regions may change and could impact on the seasonality, frequency and magnitude of water discharge in rivers (Beniston, 2005). Beniston (2003:136) states that "*it is important to know.....the period of the year when peak runoff can be expected to occur*" as hydrological management strategies are likely to have to alter as temperatures increase.

IPCC (2007) have projected that it is 90% likely that the snow season will shorten and there is a 60% probability that the depths of snow lying will decrease across most of Europe. In addition to the hydrological and geomorphological changes that may occur as a result of this, snow tourism is also a vital component for the economy of these regions through winter ski resorts. Beniston et al. (2003) suggest that the duration of snow cover could be reduced by 50% (at altitudes of 2000 m) and by 95% (below 1000 m) for the 4°C shift in mean winter temperatures projected for the

European Alps (IPCC, A2 emissions scenario, 2007). In the Petit lac d'Annecy catchment the ski resort at Station de la Sambuy lies between 900m and 1800m close to Seythenex. This resort would be particularly vulnerable to the projected reduced duration of snow. Many of the lower altitude ski resorts are already experiencing poor snow retention which is proving costly for tourism. If the warming trend continues as projected; only 44% of ski resorts in the Alps would still be snow-reliable (Elasser & Bürki, 2002).

Projected increases in temperature could result in significantly reduced snow storage and the early onset or even the lack of a snow melt phase, which would significantly alter the hydrology and sediment regime in the Alps, particularly at lower elevations and in the foothills. These changes would produce radically different flood regimes, including altered flood frequency and magnitudes, with significant potential for altering the nature and rate of geomorphic change and sediment movement. Minor perturbations in system dynamics are often amplified in mountain regions, thus they are a critical zone for studying the sensitivity of ecological, hydrological and socio-economic structures in response to climate change (Beniston, 2003).

Though there are many global and regional future scenarios available from the climate modelling community, these have rarely been coupled to geomorphic process models to predict or simulate the impacts of climate change on the landscape. Thus the need for realistic, robust simulation models is greater than ever before, particularly those that handle complexity and feedbacks in non-linear systems (Dearing, 2005, see chapter 3). Whilst snow storage/melt models, such as Alpine3D (Lehning et al. 2006) exist, they are snow-process intensive, highly parameterised and often used over short periods of time (< 10 years). Alpine3D modelling has a focus on the distribution, amount and duration of snow cover, and the impacts on the hydrological regime, rather than any impacts on sediment. It cannot therefore simulate hydro-geomorphology or process rates in the same way that CAESAR can. Rather than a focus on the distribution of snow cover, CAESAR simulations (see Chapter 5) have been adapted to take into account the effects of temperature, through the use of a snow mass balance model that creates a precipitation series that imitates the annual temporal pattern of snow hydrology (Chapter 5). Here CAESAR has been used to simulate the hydro-geomorphic regime for a hypothetical Petit lac d'Annecy catchment lacking snow-storage. This chapter presents simulations for two scenarios; one with a 'typical' pre-Alpine snow-regime, and one lacking any snow storage/melt. The simulations were driven using precipitation (1. adjusted and 2. not-adjusted for snow storage) time series for the last 1973 years, and the simulated sediment and water discharge data were examined for differences in the hydro-geomorphic regime. These simulations also incorporated the impacts of land use change (the *m*-value time series for the last 1973 years (see chapter 5.5) and differences in catchment morphometry and relief by using both the Ire and Tamie sub-catchments.

8.2 Model Setup

The overall methodology and model setup are detailed in chapter 5. Here simulations used the steep-sided bedrock-constrained mountain torrent of the Ire and the also steep sided Tamie valley which has a much wider floodplain (section 4.4). Simulations used two contrasting climate scenarios; firstly with a precipitation series moderated by temperature (section 5.5.2) to imitate snow store and snow melt, and secondly with a 'snow free' precipitation series that was not moderated by temperature, to reflect what perhaps could be expected if the pre-Alps were unable to sustain snow cover.

The input files used were an hourly 1973 year precipitation series derived from a combination of bog surface wetness palaeodata (AD 0-1500: Charman et al., unpublished) and a regional multi-proxy gridded precipitation dataset (AD 1500-1973: Pauling et al., 2006) both scaled to Petit lac d'Annecy averages (section 5.4). The two series were joined together to produce a continuous 1973 year time series of hourly precipitation. The snow-adjusted series was moderated using a multi-proxy temperature series (AD 0-1973: Moberg et al., 2005) also scaled to Petit lac d'Annecy averages (section 5.4.4.3). Forest inventory and pollen data (section 5.5) were used to generate the land use history. The land use data were converted into an hourly m-value time series (section 5.5), which within CAESAR is used to modulate the flood hydrograph. In addition, these simulations incorporate hillslope processes (soil creep and erosion rate), which respond to the soil moisture (j_mean) parameter (section 5.5) which is governed by both the rainfall and *m* value series (detailed in chapter 7). An identical land use record was used for both sub-catchments given that local sub-catchment histories cannot be discerned from the regional vegetation signal in the Petit lac d'Annecy pollen record (sediment core LA13) (Jones et al., in prep.).

8.3 Hydrology and the flow regime

8.3.1 Flood magnitude and frequency

Significant differences in annual hydrology such as frequency and magnitude of flood events are evident between the 'snow' (T1, I1) and 'no snow' (T2, I2) (Fig.8.1) simulations for both sub-catchments. The principal differences between the snow and no-snow simulations for both catchments is that the snow-melt episode with

	T1	T2	*factor*	I1	I2	*factor*
Catchment	Tamie	Tamie		Ire	Ire	
Hydrological input	WITH SNOW	NO SNOW		WITH SNOW	NO SNOW	
Number of flood hours above 8 cumecs	91693	156861	1.7	50947	95622	1.9
Number of flood hours above 12 cumecs	29552	53588	1.8	17683	38261	2.2
Number of flood hours above 20 cumecs	7257	15357	2.1	4740	11964	2.5
Total sediment discharge for simulation (m3)	996907	1406605	1.4	1796903	2404501	1.3

Table 8.1 *Number of flood hours with a magnitude greater than 8, 12 and 20 for snow and no-snow conditions in the both the Tamie(T1,T2) and Ire (I1, I2) and total sediment discharge for each simulation. 'Factor' denotes the factor of difference between snow and no-snow.*

peak flows was removed and high magnitude flow events become more evenly spaced throughout the year. Peak discharges for each simulated year are always higher under 'no snow' (T2, I2) than 'snow' (T1, I1). Under the no-snow conditions the magnitude of floods events has been increased probably due to the additional runoff from intense winter precipitation (T2, I2). Thus with the catchments unable to sustain snow storage and melt the frequency of hours of floods with a magnitude greater than 8 m3s-1 is 1.7 times and 1.9 times greater than when there was snow (for the Tamie and Ire respectively) for the 1973 year time period. The frequency of flood events under 'no snow' regime with a magnitude greater than 12 m^3s^{-1} is 1.8 and 2.2 times more frequent (for the Tamie and Ire). Floods with a magnitude greater than 20 m^3s^{-1} are 2.1 and 2.5 time more frequent. Therefore, the simulations for no-snow conditions are underpinned by greater numbers and more regular high magnitude floods and the absence of sustained lower magnitude flows through the snowmelt months (March-May).

8.3.2 The annual hydrological cycle

To examine in greater detail the annual hydrological pattern for both snow (T1) and no-snow (T2) conditions, for the Tamie two five year periods were selected containing a series of years with low total annual water discharge (AD 808-813) (Fig.8.1a) and high total annual water discharge (AD 916-921) (Fig.8.1b). Under low total annual discharge conditions (AD 808-813) for the snow simulation winter (January-March) discharges are 2-3 m3s-1 in all 5 years, which contrasts with the more flashy hydrograph containing peak discharges of 10-30 m3s-1 for winters under no-snow conditions (Fig.8.1). During the April-May sustained snowmelt period, river discharges fluctuate around 8 m3s-1 (T1: Fig.8.1). This snowmelt period is a feature that is part of the annual pre-Alpine hydrological cycle (Beniston et al., 2003). The no-snow simulation (T2) does not have the same sustained discharge peak (Fig.8.1d), but instead has individual rain-induced flood events that vary in magnitude. Differences between snow and no-snow discharge between July to September are negligible, and perhaps reflect occasional and limited duration snow storage and release. The differences between the two simulations become obvious again between October and December when I2 (Fig.8.1d) clearly shows more frequent and higher magnitude flood events than I1 (Fig.8.1b) due to water storage as snow.

Very similar annual patterns of hydrological differences occurred between snow and no-snow simulations for a period with high total annual discharge (AD 916-921) (T1 and T2: Fig.8.1). Differences comprise the extremely high discharges in the no-snow regime (T2) between January to March, and reflects a combination of the increased run-off, the higher total annual precipitation and also lower *m*-values (forest cover) amplifying the flood hydrograph (Fig.8.2g). Despite these discrepancies the pattern of alterations to the hydrological cycle between the snow and no-snow simulations is similar for different total annual precipitation and catchment land use.

Therefore if there is more total annual precipitation, the temporal pattern of the annual hydrological regime does not differ, only the magnitude of extreme events. Removing the snow melt/store substantially alters the annual hydrological regime, with more high magnitude flows between October to March, and a flashy regime during April and May replacing the sustained lower magnitude discharge peak produced by the snow store/melt (Fig.8.1f). Under both scenarios, the annual sediment discharge patterns largely follow the water discharge patterns (e.g. Fig.8.1a,c,f,g).

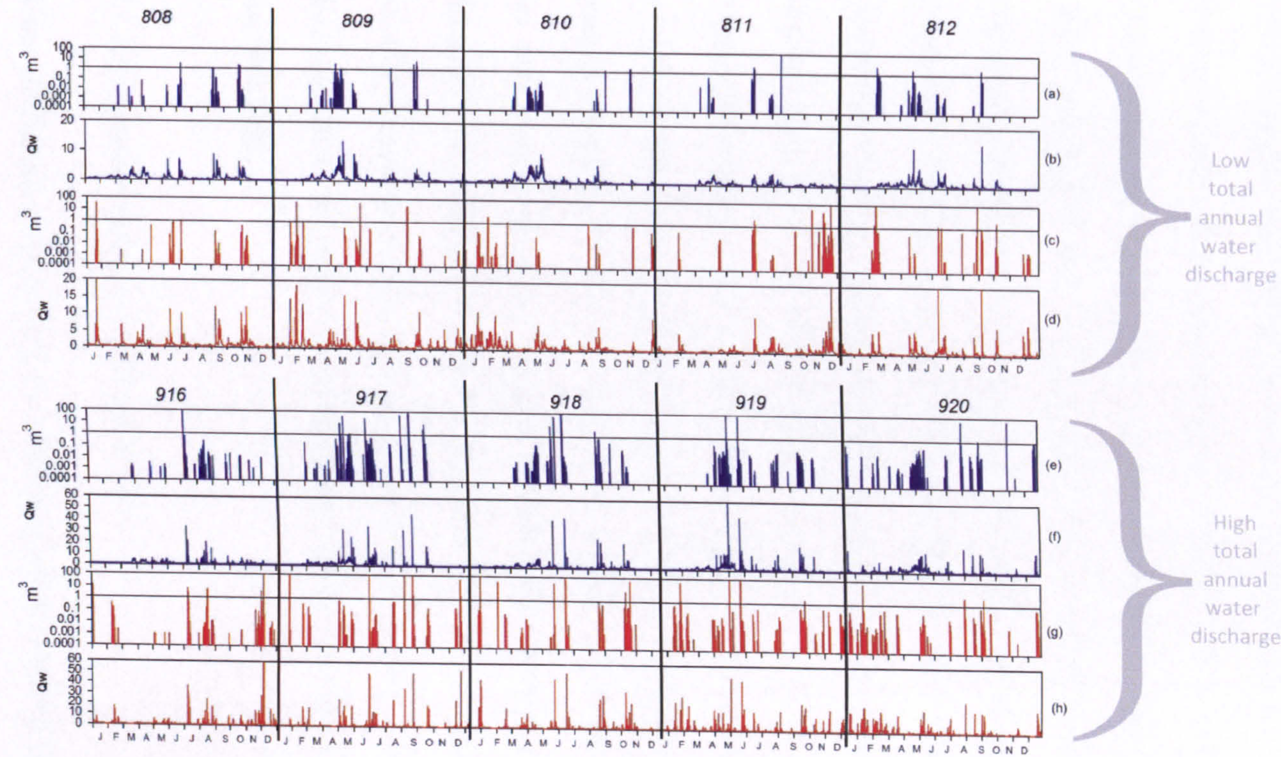

Figure 8.1 *(a) Tamie 'snow' hourly sediment discharge under 'low' total annual water discharge (b) Tamie'snow' hourly water discharge under 'low' total annual water discharge (c) Tamie'no snow' hourly sediment discharge under 'low' total annual water discharge (d) Tamie'no snow' hourly water discharge under 'low' total annual water discharge (e)Tamie 'snow' hourly sediment discharge under 'high' total annual water discharge (f) Tamie 'snow' hourly water discharge under 'high' total annual water discharge (g) Tamie 'no snow' hourly sediment discharge under 'high' total annual water discharge (h) Tamie 'no snow' hourly water discharge under 'high' total annual water discharge*

These simulations show similar behaviour to that described by other snowmelt-ruonff models and empirical data (e.g. Gleick, 1987; Lettenmaier and Gan 1990; Hamlet et al. 2005, 2007; Mote et al. 2005, and Stewart et al. 2005). Beniston (2003) suggests that reduced snow cover in mountain regions will have particular implications for early seasonal runoff. A hydrological characteristic modelled in these simulations is change in the alteration of seasonal runoff, in particular during winter and typical 'melt' months. Horton et al., (2006) show increases in winter discharges and an earlier and decreased snowmelt-induced peak in their modelled projections of hydrological impacts to reductions in snow cover. These projections corroborate the annual hydrological behaviour of the simulations presented here. The findings from a more recent modelling study by Bavay et al., (2009:102) also support the findings presented here. They suggest that changes in water discharge as a result of reduced snow cover is "*severe*" with a much shorter snow melt discharge peak in spring, in addition to which, heavy precipitation events in later months of the year would fall as rain instead of snow, thus increasing the probability of flooding. This suggests that the hypothetical simulations under conditions of no-snow presented here follow the general trend seen in other snow-melt models, and thus some level of confidence can be given to annual behaviour seen in CAESAR modelled data.

8.3.3 Long-term climate and land-use drivers

Water discharges produced by the 1973 year simulations for both sub-catchments (Fig.8.2) under snow and no-snow conditions can be assessed in several ways. Firstly, the total decadal Qw mirrors the precipitation series and is identical for both catchments. The minor offset between snow and no snow simulations for each sub-catchment reflects the storage of snow across the end of each decade (Fig8.2a&b). The high magnitude flood events for both snow and no snow simulations in both sub-catchments are closely tuned to both incidents of large precipitation events (Fig 8.2e&f) and the decadal pattern of total rainfall (not shown). The most striking difference between snow and no snow conditions is that for both the Ire and Tamie higher peak annual discharges occur in every year of the 1973 year series under no-snow conditions.

Whilst long-term frequency of peak flood events seems to be largely controlled by the climate series, forest cover also exercises some control over flood magnitude because the land use parameter (*m* value) in CAESAR simulations regulates the flood hydrograph. For example at AD 280 forest cover was reduced by ~20% (Fig.8.2g), which when coupled with an increase in precipitation at the same time, produced the highest discharges of the entire series (Fig.8.2c&d) in both snow and no-snow simulations.

However, both high levels of precipitation and reduced forest cover are necessary, for example AD 480-680 the forest cover was further reduced by 10% (Fig.8.2g) and floods did not reach these magnitudes because there are insufficient high magnitude precipitation events. Peak flows under both snow and no-snow simulations are sensitive to changes in land use and the highest magnitude flow events coincide with heavy rainfall and reduced forest cover regardless of snow or no-snow conditions.

The pattern of long term change in sediment discharge is broadly similar between the four simulations (Fig.8.3), with more rapid accumulation taking place in the sediment accumulation curves at AD 0-150, 300-400, ~600, 900-1100 and 1600-

1700. The rapid accumulation at the base of the simulations is likely to be a function of model

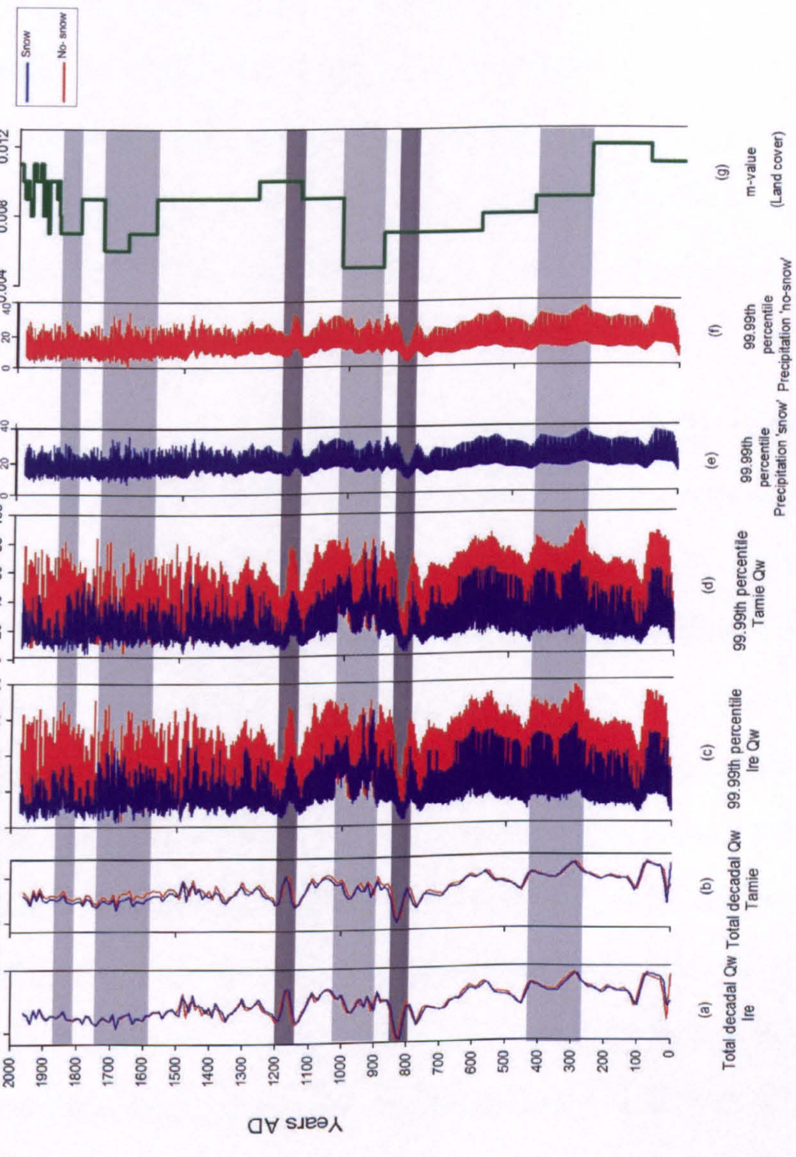

Figure 8.2 *(a) Total water discharge for snow (blue) and no snow (red) in Ire (b) total water discharge for snow (blue) and no snow (red) in Tamie (c) 99*

equilibration at the beginning of the model run, notwithstanding the 60 year run-up period (not shown). The other increases in sediment accumulation rate all coincide with significant land use changes, i.e. expansions in agriculture and reductions in forest cover. Human impact is frequently cited as possible cause of increased landscape sensitisation (Chiverrell et al., 2007) and thus increased sediment supply due to increased landscape instability and gradual erosion of soils and sediment (Edwards and Whittington, 2001). Coarse grained sediment field studies coupled with modelling have also suggested that localised reforestation is capable of reducing sediment delivery rates (Lane et al., 2008).

As forest cover reduces in CAESAR (through the *m* parameter), the response of the hydrological model is altered and the magnitude of water discharge is augmented, thereby providing the system with increased stream capacity. At the same time, due to the increase in water discharge magnitudes and therefore available water, rates of hillslope processes such as soil creep and soil erosion also increase (chapter 7), thereby providing the system with an increased supply of sediment. Both simulations in the Ire respond to the drop in forest cover around AD 280 (Fig.8.3n), yet there is more sediment discharge in 'no snow' (Fig.8.3i) than 'snow' (Fig.8.3j). Both simulations have the same total annual precipitation input (Fig.8.2e, Fig.8.2f) and forest cover input (Fig.8.3n), yet the way in which the water is delivered to the catchment (e.g. the changes in annual hydrology from lack of snow) has a profound effect on water discharge peaks which are higher in each year for the 'no snow' simulation throughout the 1973 year period. The increase of water in the system leads to increases in capacity and stream power and results in greater ability to transport sediment.

The amount of reduction in forest cover may also play a role in amount of sediment discharge. For example, between AD 630 and AD 870, despite being the lowest period of forest cover (Fig.8.3n) since before AD 0, there are no increases in sediment discharge in 'snow' (Fig.8.3i). This suggests that perhaps by reducing forest cover in a step-wise manner by 6% at each of the previous drops, the system begins to stabilise and there are no real responses to land use reductions. The biggest sediment discharge events take place when there is a drop in land use of 15-20% where the sediment peaks seem to be much higher. Sediment discharge stablises after 1000 AD with the exception of a small increase in both simulations at AD 1150 which is a response to a climate event as there are increased flood discharges and increased capacity of water here. Supply is limited as forest cover has increased therefore, the increases in sediment discharge are probably not as high as they would be if this magnitude of precipitation occurred under lower forest cover.

8.4 Catchment geomorphology and sediment delivery

8.4.1 Sediment delivery to the lake

The total volume of sediment delivered from both catchments differs substantially between snow and no-snow conditions. There is 1.4 times more total sediment under the no-snow simulation in the Tamie and 1.3 times more total sediment under no-snow in the Ire (Table 8.1), which shows some similarity between the response of sub-catchments. A hypothetical sediment accumulation rate for each modelled sediment record has been calculated for deposition in a focussed area of the Petit lac d'Annecy ($2km^2$) assuming both rivers flow directly into the lake (i.e. no floodplain

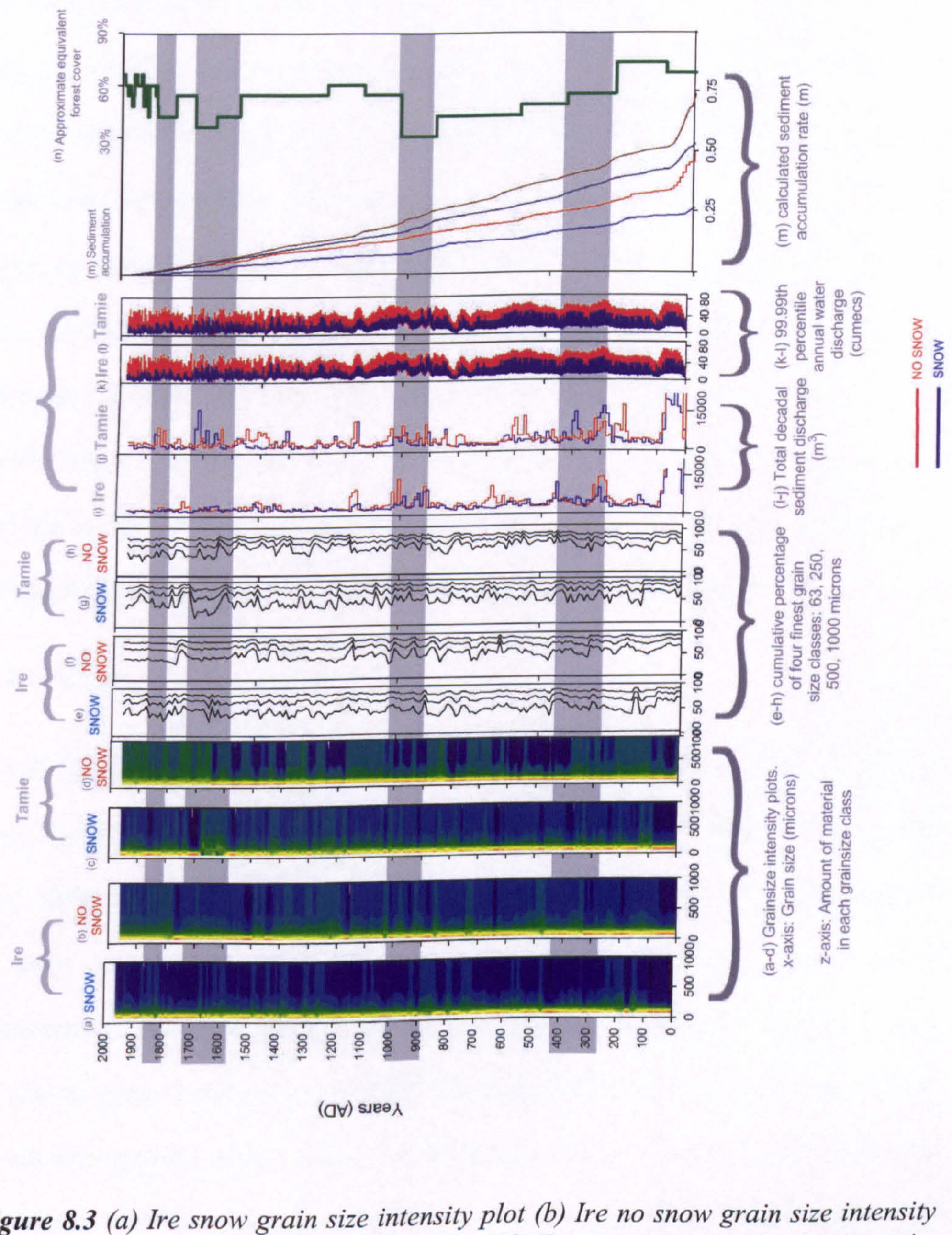

Figure 8.3 *(a) Ire snow grain size intensity plot (b) Ire no snow grain size intensity plot (c) Tamie snow grain size intensity plot (d) Tamie no snow grain size intensity plot (e)Ire snow cumulative percentage of four finest grain size fractions (63 μ, 250 μ, 500 μ, 1000 μ), (f)Ire no snow cumulative percentage of four finest grain size fractions (63 μ, 250 μ, 500 μ, 1000 μ), (g)Tamie snow cumulative percentage of four finest grain size fractions (63 μ, 250 μ, 500 μ, 1000 μ), (h)Tamie no snow cumulative percentage of four finest grain size fractions (63 μ, 250 μ, 500 μ, 1000 μ),(i) Ire total decadal sediment for snow (blue) and no-snow (red), (j) Tamie total decadal sediment for snow (blue) and no-snow (red), (k) Ire 99th percentile annual water discharge for snow (blue) and no-snow (red), (l) Tamie 99th percentile annual water discharge for snow (blue) and no-snow (red),(m)calculated sediment accumulation rate for Ire snow (blue) Ire no-snow (red), Tamie snow (dark blue), Ire no-snow (dark red), (n) percentage of forest cover*

storage). Sediment accumulation rates are higher in both sub-catchments under no-snow conditions. Over the 1973 year period there is approximately 20cm and 17cm more sediment accumulation in the Ire and Tamie respectively under no-snow conditions compared to when snow storage/melt can occur. This equates to approximately 8mm-1cm additional sediment accumulation over a year. If values were similar for the other three sub-catchments in the Petit lac d'Annecy catchment, there would be an estimated 1m of additional sediment accumulation under no-snow conditions or ~10cm additional accumulation per year which is approximate to several hundred years of accumulation based on the last 2000 years of lake sediment accumulation rates (LA13 – 4m in 2000 years; Jones et al., in prep.).

8.4.2 Flood behaviour and sediment delivery

CAESAR outputs temporal data for 9 user-defined grain size fractions and here the four finest grain size fractions 63μ (~silt), 250μ (~fine sand), 500μ (~coarse sand) and 1000μ (~fine gravel) are presented (Fig.8.3 a-h). Cumulative plots of percentage of grain size fractions (Fig.8.3e-h) are useful to determine the composition of sediments through time and to identify phases of behaviour such as increased periods of coarser grained sediments. In addition to this, contour plots which are essentially 'grain size intensity' plots (Fig.8.3a-d) have proved to be a useful way of displaying abundance of different grain size fractions over time (y-axis). More 'intense' (or 'warmer') colours such as red/yellow/orange (z-axis) at a particular grain size class (x-axis) suggest a greater percentage of this grain size fraction. 'Cooler' colours such as green and blue show a low percentage of the grain size fraction. Displaying the data in this way makes it easy to see phases or peaks in relatively coarser grained sediments.

General patterns show that there are more phases of coarser material in no-snow simulations for both sub-catchments (Fig.8.3b; Fig.8.3d). This is likely to be due to the alteration of the annual hydrological cycle which has greater frequency and higher magnitude flood events (Fig.8.1) and the capacity of the channel has increased thus there is increased stream power to transport coarser grained materials. The increase in coarser grained sediment relative to fine grain during quiescent periods in both sub-catchments suggests that the overall nature of grain size composition would become coarser e.g. AD 1300-1500. There is also more variability and shifts from coarse to fine and vice versa in no-snow simulations (Fig.8.3f&h). A temporal characteristic that is true for all four simulations is a fairly sharp (20-30 year) increase in fine grained material relative to coarse grained (Fig.8.3e-h) e.g. AD 900 (Fig.8.3e), AD 1160 (Fig.8.3f), AD 1740 (Fig.8.3g) and AD 790 (Fig.8.3h). The peak in fine grained material relative to coarse grained material coincides with a peak in total sediment discharge (Fig.8.3i-j) which is followed by a sharp drop in fine-grained material relative to coarse. This suggests that there is a build up of fine grained material in the system which is regularly flushed during peak flood discharges. There appears to be no direct control over this behaviour other than during peak discharges, whether that is through reduced forest cover or increased precipitation.

The coarser phases in the Ire no-snow (Fig.8.3b) simulations broadly correspond with the major phases of reduced forest cover. However, there is a long phase of

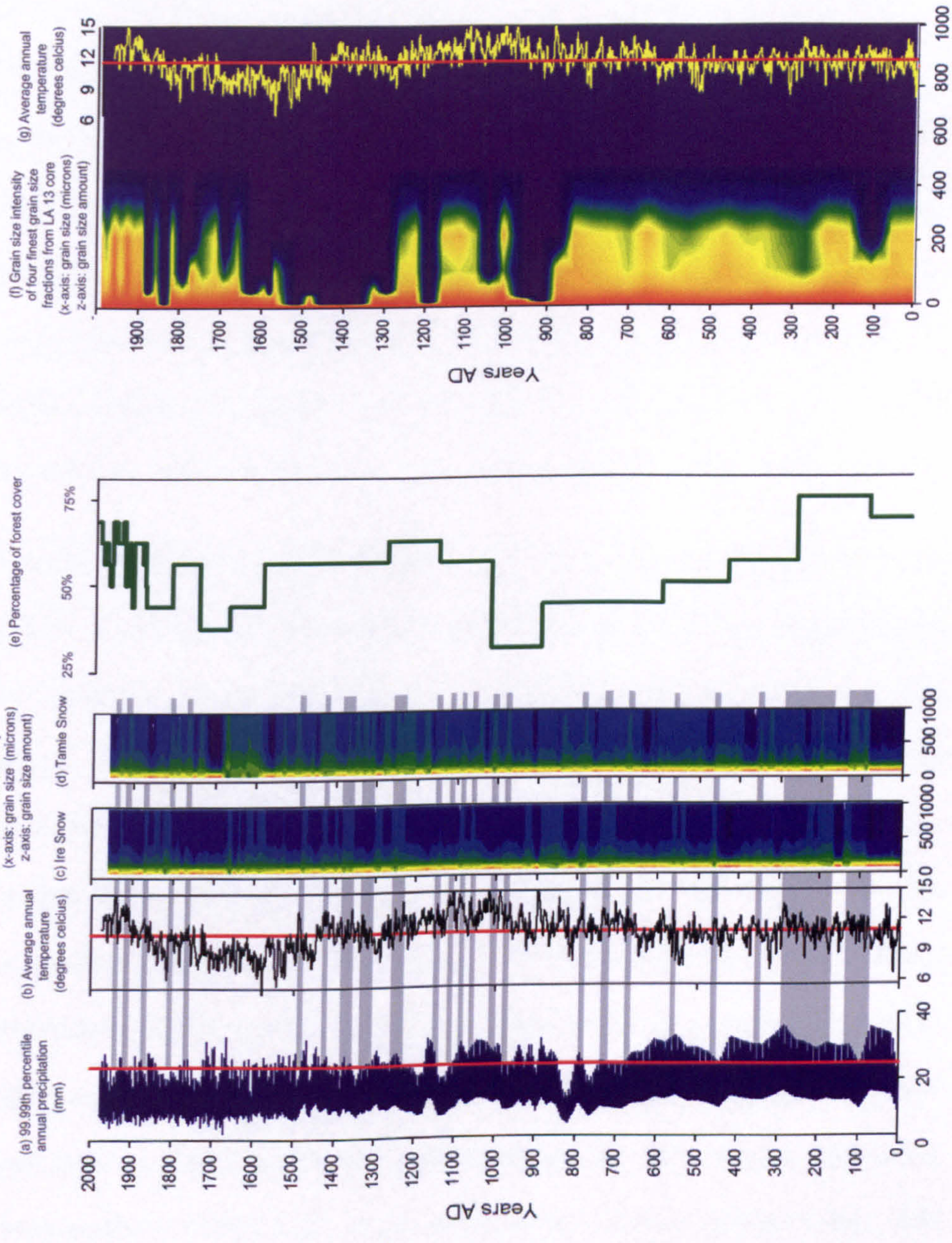

***Figure 8.4** (a) 99.99th percentile annual precipitation with a constant of 25mm marked in red, (b) Average annual temperatures for Northern Hemisphere (Moberg et al., 2005) scaled to the average of the Petit lac d'Annecy catchment with a constant average annual temperature of the series marked in red (10.94°C), (c) grain size intensity plot for Ire 'snow', (d) grain size intensity plot for Tamie 'snow' (e) percentage of forest cover (f) grain size data from the Petit lac d'Annecy LA-13 core (Jones et al., in prep.) (g) Average annual temperatures for Northern Hemisphere (Moberg et al., 2005) scaled to the average of the Petit lac d'Annecy catchment with a constant average annual temperature of the series marked in red (10.94°C).*

coarse grained material during the period of *increased* forest cover AD 1200-1500 which does not fit with the general trend and are not easily accountable for. The coarse phases in the Tamie no-snow (Fig.8.3d) have a lesser match with reductions in forest cover although there is some broad correspondence at AD 300-400, AD 1600-1700 and AD 1800-1850. However, there are other phases of coarse grained material that are not in synchronicity with reduced forest cover and may be a function of a lagged response due to storage e.g. AD 1000-1100 or perhaps increased incision due to reduced sediment supply under high forest cover e.g. AD 1300-1350.

The fluctuations in grain size for the snow simulations require a more detailed study (Fig. 8.4). The behaviour of the coarse grained material for no-snow suggests that it moves under the highest flood discharges, and appears to correspond well with reductions in forest cover. A similar style of behaviour must be occurring in the snow simulations, yet they seem to correspond less well to reductions in forest cover, suggesting that something other than just increased flood magnitudes under low forest cover is exerting control over them. The coarse grained events correspond loosely to some high magnitude precipitation events (Fig.8.4a) e.g. AD 1680, but not enough to suggest that they are strongly linked. The grain size data were compared to mean annual temperature for the Northern Hemisphere (Moberg et al., 2005) that has been scaled to Petit lac d'Annecy averages (see chapter 6) (Fig. 8.4b). The match between above average temperatures (average for 1973 year series is 10.94°C and is marked on Fig.8.b with a constant line) and peaks in coarse grained sediment events in both sub-catchments is remarkable (Fig.8.4c-d). The precipitation input data has been modulated by this temperature series to 'imitate' typical characteristics of annual snow melt/store. The match between the peak annual temperatures and coarse grain size events suggests that fluctuations in temperature alter the rate of snow melt, and warmer than average years experience a faster rate of melt. This increases melt rate then increases flood magnitudes over short periods, thus enabling greater carrying capacity of the channel and ultimately transport of coarse grained sediment. During periods when temperatures are lower e.g. the climatic deterioration around AD 1500-1700, snow melt is likely to be much slower, and thus resulting in a reduced incidence of coarse grained events. However, coarse grained events do still occur during lower temperatures, but have greater correspondence to low forest cover e.g. AD 1600-1700 (Fig. 8.4c-d) or to years where maximum annual precipitation is higher than 25mm (Fig. 8.4a) e.g. AD 1158. This hypothesis corresponds well to a study undertaken by McDonald and Lamoureaux (2009) who observed that the transport of coarsest grain sizes occurred in conjunction with increased stream velocity during the snowmelt flood. They conclude that the timing, magnitude and importantly the intensity of the snowmelt flood were primary controls over river discharge and suspended sediments.

To test this hypothesis further, a broad comparison was made between grain size data for core LA13 (Fig.8.4f) (Jones et al., in prep.) and average annual temperature for the Northern Hemisphere (Moberg et al., 2005) scaled to averages of the Petit lac d'Annecy catchment (Fig.8.4g). The general trends compare well, with lower temperatures broadly corresponding to periods of less coarse grained material e.g. AD 1400-1500 and periods with warmer average temperatures corresponding to phases of coarse grained material e.g. AD 800. Perhaps most interesting, is the coarsening up sequence near the top of the core, from AD 1800. During this period where temperatures are increasing, there is increasing intensity of grain sizes around 250μ and 500μ.

Overall these results suggest that warmer than average years control the rate of snowmelt; suggesting that warmer years increase the rate of melt but reduce the duration resulting in more focussed high magnitude discharge events which seem to play an important role in delivery of coarse grained sediment.

8.4.3 Channel capacity and sediment storage.

There tend to be more regular, high magnitude sediment discharge events in the Tamie than the Ire throughout the 1973 year record (Fig. 8.3i-j) some of which do not correspond to high magnitude precipitation events or reductions in forest cover e.g. AD 1500 (Fig. 8.3j). These events are likely to be a function of storage and release cycles due to the nature of the catchment morphometry in the Tamie where there is more opportunity for sediment storage and the alluvial fan is effective at trapping sediment (this is discussed further in chapter 10). There is also a tendency for these high magnitude events to be more frequent under no-snow conditions than when there is snow. Grain size data (Fig. 8.3a-h) have show that these peaks correspond to increases in fine grained material relative to coarse grained material which suggests that there is a build up of fine grained material during lower water discharges. These fine grain sizes are then flushed out of the system during peak flood discharge.

The sediment discharge in the Ire declines over the 1973 years towards the present day (Fig.8.3i). In addition to this, the pattern of sediment discharge in the Ire is generally similar for snow and no-snow but the latter has higher sediment discharge magnitudes. This is particularly noticeable during periods of reduced forest cover when there is increased sediment supply and increased in flood magnitudes. The Tamie has more frequent high magnitude discharge events during no-snow conditions which are likely to be due to higher flood magnitudes enabling re-mobilisation of sediments.

The peak in sediment discharge at AD 450 (Fig.8.3j) may be a response to the small drop in forest cover (6% = 0.001m) around this time, yet there is more sediment in 'snow' than 'no snow' which suggests that this peak is more likely to be previously stored material being re-mobilised and moved out of the catchment at this time.

Another explanation for this is that the discharge inputs begin to reduce due to decreases in total annual water discharge and there are therefore lower magnitude peak flood events, which means that there is simply not the capacity available in 'snow' to move the material. The same does not apply for 'no snow' which records sediment discharge in a large peak around 650 AD. Flood magnitudes do not drop off as rapidly or by as great a magnitude as they do in 'snow', therefore 'no snow' simulation still has the capacity to move sediment. As the forest cover drops again at AD 900-1000, there is again both an increased supply of sediment and capacity of the stream in both simulations due to increased discharge magnitudes. There is a small increase in sediment discharge in 'snow' during the reduction in forest cover at 1650 which is likely to be an increase in sediment supply and channel capacity at this period, yet this is not seen in 'no snow' as water discharges have reduced significantly, despite being during a period of low forest cover. The same pattern occurs in the Tamie which suggests that there may be some large snow melt events occurring in 'snow' which are not seen in 'no snow' therefore limiting the capacity of the stream. These observations tie in well with empirical observations from a recent study which suggests that during periods of low flows, there may have been sediment available to the system for transport but delivery is limited by reduced stream competence (McDonald and Lamoureaux, 2009).

8.5 Conclusions

(1) The CAESAR simulations here show that under no-snow regimes there are more frequent high magnitude floods through winter months and the large spring snowmelt peak is replaced by individual high magnitude flood events. Increased sediment delivery coincides with the highest magnitude flood events. These findings correspond well with empirical data and other snow melt models.

(2) Frequency and magnitude of flood events are closely linked to frequency and magnitude of precipitation however, reductions in forest cover correspond to increases in the magnitude of discharge events which is largely a function of the control of the *m* parameter in CAESAR. Both sub-catchments show that peak annual water discharges are higher under no-snow regimes which have significant implications for the impacts of projected climate change as there will be an increased incidence of flood events and consequently increased sediment discharge. However, in addition to this the percentage and nature of forest cover reductions may have implications for sediment delivery. The greatest increases in sediment discharge occur under 20% reductions in forest cover. In addition to this, during one of the lowest periods of forest cover (AD 600-900), there is little increase in sediment, which suggests that a step-wise reduction in forest cover (e.g. small reductions from AD 280- 900) may have less of an effect on sediment delivery than a sharp clearance. These ideas are explored further in chapter 9.

(3) One of the major controls on sediment delivery is the relationship between supply of sediment and the capacity of the channel. In the Tamie particularly, supply often exceeds the capacity of the channel to transport material, and sediment is often

stored until a later flood event re-mobilises it. This renders systems with increased opportunity more difficult to predict than well coupled systems with reduced opportunity for storage. This suggests that sediment legacy over decadal-centennial scales in a system plays an important role in the future of sediment availability and delivery. In addition to this, under periods of reduced sediment supply, fine grained sediments have been flushed out during peak discharges and there is then greater incision then takes place. There is also a greater amount of coarse grained material relative to fine in the sediment discharge. Water discharges are higher under no-snow, but along with this there is an increase in sediment from augmented hillslope processes.

(4) Perhaps one of the most interesting findings is the control exerted by higher than average annual temperatures on coarse grained sediment delivery under snow conditions. The findings suggest that warmer annual average temperatures increase the rate of snow melt, thus increasing the magnitude of discharge over a shorter period. If rising temperatures cause snow in mountain regions disappears altogether, then more frequent high magnitude flood events occur throughout the year which then (in conjunction with amplification of flood events through reductions in forest cover) control the delivery of coarse grained material in the system. An increase in coarse grained sediment transport and delivery has massive implications for rivers such as increased channel aggradation and increased braiding which ultimately leads to increased flooding and has implications for fauna in gravel-bedded streams These suggestions are speculative, but a more in-depth study is a consideration for future research.

Chapter 9

Sensitivity testing changes in land use.

9.1 Introduction

As land use reduces either through human interference such as deforestation, or naturally through shifts in climate, empirical data suggests that water yield increases due to reduced soil moisture storage and hydrological change induced by forest cover are variable and unpredictable (Hibbert, 1967; Bosch & Hewlett, 1982). These results come from studies within paired catchments whereby two catchments of similar size and characteristics are studied side by side to identify the effects of land use change as one catchment is deforested. There are many limitations of paired catchment studies including a heavy focus on hydrological change rather than changes in sediment regime and often there are few reports of longer term studies (Brown et al., 2005). In addition to being a primary control on hydrology land use has also been identified as a primary control on sediment supply and also catchment hydrology (Starkel et al., 1998; Knighton, 1998). Removal of forest can increase system sensitivity to climate and cause disproportionate impacts in sediment dynamics (Madej, 1995; Marron et al., 1995).

Although inferences can be made about sediment delivery (i.e. sediment moves under highest discharge events) in paired-catchment studies and from empirical evidence, the differences are difficult to quantify in the absence of comparable data. Palaeoarchives from lake sediments offer an alternative approach and are continuous records of change, ideal for long term studies. Proxy records for flooding (e.g. Thorndycraft et al., 1998) and catchment erosion (e.g. Foster et al., 2008) discerned from lake sediment records have been useful for identifying system responses during periods of perceived flooding, and for recognising past analogue conditions. Yet it is extremely difficult to quantify these changes as other variables such as total annual precipitation and the frequency and magnitude of stochastic events are not held constant when looking at analogue conditions. In addition catchment land use and channel management strategies to mitigate extreme flows and erosion of sediment can seldom be excluded from observations.

The only real alternative to improving investigation into the effects of land use on landscapes is through the use of numerical modelling. Modelling allows individual variables (for example precipitation, catchment geometry or land use history) to be held constant, which ensures that any changes in sediment regime and hydrology reflect changes from a single variable. In CAESAR land use exercises control over the hydrology and geomorphology through the manner in which the *m*-value modulates the flood hydrograph and through j_mean, the soil moisture term (indirectly linked to *m*-value) which has a control over hillslope processes. In other chapters (chapter 6-8, chapter 10), time series of variations in land use (*m*-value) have been used to identify responses of the system to human activity over time. This chapter explores the impacts of systematically adjusting the entire 1973 year series for land use upwards by 20% and downwards by 20% to encompass less and more intensively used landscapes (Fig 9.1). The focus here is on the implications for sediment delivery to the lake. Each *m* value equates to approximately 6% forest cover (section 5.5).

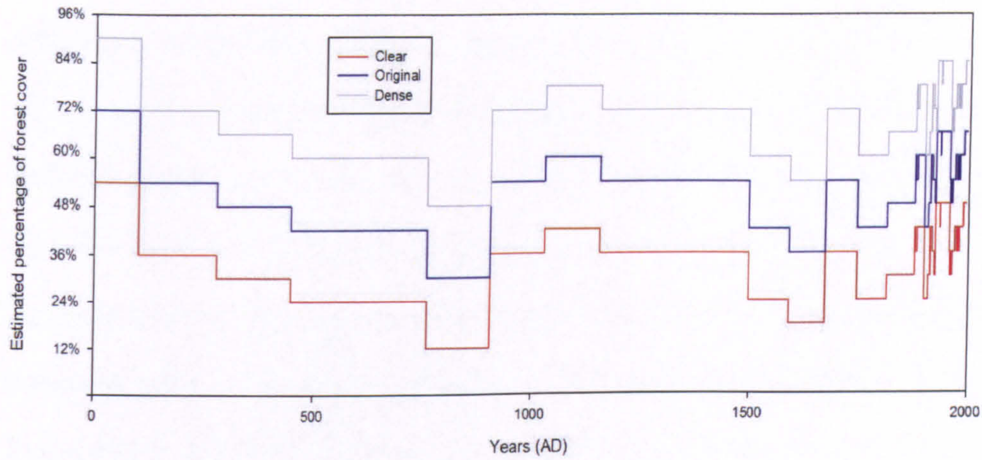

Figure 9.1 *Approximate percentage of forest cover for each simulation, original (blue), clear (red) and dense (grey)*

9.2 Method and Model Setup

In order to investigate sensitivity in different morphological settings, two contrasting styles of sub-catchment the Tamie (section 4.4.1) and the Ire (section 4.4.2) are used here. The input files used were an hourly 1973 year precipitation series derived from a combination of bog surface wetness palaeodata (AD 0-1500: Charman et al., unpublished) and a regional multi-proxy gridded precipitation dataset (AD 1500-1973: Pauling et al., 2006) both scaled to Petit lac d'Annecy averages (section 5.4). The two series were joined together to produce a continuous 1973 year time series of hourly precipitation. The snow-adjusted series was moderated using a multi-proxy temperature series (AD 0-1973: Moberg et al., 2005) also scaled to Petit lac d'Annecy averages (section 5.4).

Pollen data (section 5.5) were used to generate the land use history. The land use data were converted into an hourly *m*-value time series (section 5.5), and in

CAESAR this is used to modulate the flood hydrograph. In addition, these simulations incorporate hillslope processes (soil creep and erosion rate), which respond to the soil moisture (j_mean) parameter (section 5.5) which is governed by both the rainfall and *m* value series (as detailed in chapter 7). The same land use records are used in both the Ire and Tamie as local sub-catchment histories cannot be discerned from the regional vegetation signal in the Petit lac d'Annecy pollen record (sediment core LA13) (Jones et al., in prep.). The pollen data described as 'original' in this chapter are identical to the pollen series used for chapters 7-8 and chapter 10. In order to test landscape sensitivity, the 'original' land use was reduced by ~20% (three *m*-values) throughout the entire series to create a land use record indicative of a more open landscape. This is referred to as 'clear' throughout this chapter. In order to test the sensitivity of the landscape to a more dense forest cover, the 'original' land use was increased by ~20% (three *m*-values). This is referred to as 'dense' throughout this chapter. By adjusting the 'original' series up and down throughout the 1973 year record (Fig.8.1), it is possible to quantify the effect of forest cover on sediment delivery as land use is the only variable.

9.3 Results and Interpretation

9.3.1 Total Sediment Yield

For each land use scenario, total sediment yield (normalised for catchment area) for the four finest grain size fractions (Fig.9.2a) was 2 times (clear), 1.7 times (original)

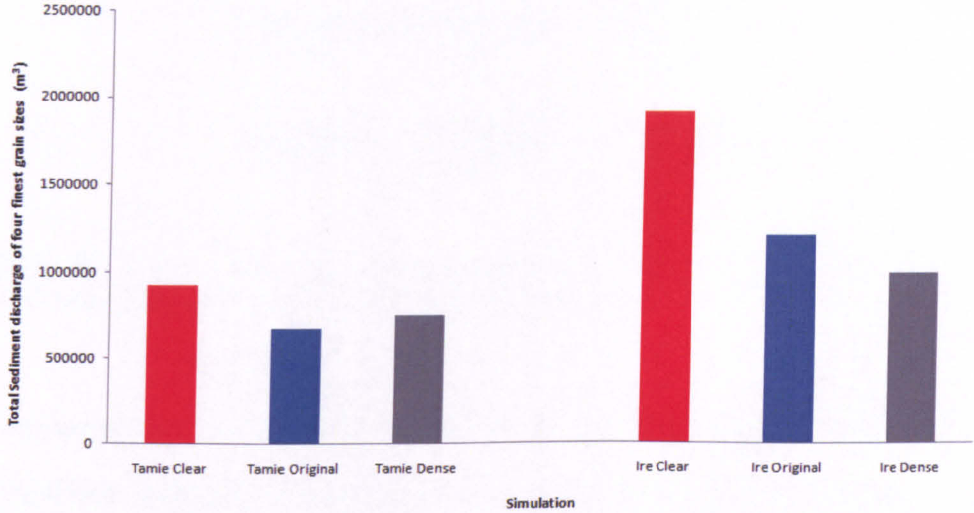

Figure 9.2 *Total sediment yield of 4 finest grain size fractions in the Ire and Tamie under different land use scenarios for the 1973 year period. (Data normalised for catchment area)*

and 1.3 times (dense) more sediment in the Ire simulations than the Tamie. This suggests that regardless of forest cover the Ire is always more efficient at moving sediment out of the catchment than the Tamie. This is likely to be a function of the steeper gradients, narrower floodplain and less opportunity for sediment storage in the Ire.

As forest cover decreases by 20% for each scenario, the total sediment yield from the Ire increases sequentially (Fig. 9.2) but the rate of this increase is not linear. The second reduction in forest cover (reduced by a further 20%) from 'original' to clear' produces more sediment than the reduction from 'dense' to 'original'.

Tamie Clear	Tamie Original	Tamie Dense
2.65	1.99	2.54

Ire Clear	Ire Original	Ire Dense
0.57	0.36	0.30

Table 9.1 *Approximate calculations of amount of sediment m^3 per m^2 deposited across the Tamie fan and La Ronde fan over 1973 years.*

Equivalent values for the 'original' and 'dense' scenarios results in a 33% decrease in sediment yield and a further 18% decrease respectively (Fig.9.3). There appears to be a threshold or tipping point in the sediment regime between the 'original' and 'clear' scenarios (between 25% and 50% forest cover) in which sediment delivery from the hillslopes to the fluvial system coupled with sufficient sediment carrying capacity in the discharge regime to produce this more substantial increase in sediment delivery. This perhaps reflects that the thresholds for temporary storage in tributary alluvial fans and on the restricted floodplain are being regularly exceeded. In order to gain a perspective of how much sediment this would equate if it were deposited on the La Ronde fan (i.e. directly out of the Ire sub-catchment onto the alluvial fan lying within the Eau Morte floodplain), an approximate calculation was made for each simulation. For the 'clear' simulation there would be approximately 0.5m added to the fan (Table 9.1) over 1973 years. Similar values for 'original' and 'dense' suggest that there would be 0.36m and 0.30m added respectively. Approximate calculations show that a 40% decrease in forest cover (from 'dense' to 'clear') could almost double the amount of sediment deposited across the La Ronde fan ($2500m^2$).

When comparing Tamie 'clear' with equivalent values for the 'original' scenario, there is a 27% decrease. However when comparing Tamie 'original' to Tamie 'dense', there is an 11% increase in the 'dense' scenario compared to 'original' in sediment yield. The increase in sediment yield between the 'original' and 'dense' scenario is intriguing, a possible explanation is that the reductions in the amplitude of the flood hydrograph imparted by the *m*-value time series have confined the majority of the flows under 'dense' to the channel. Thus the sediment delivered through the system remains in the fluvial zone, whereas perhaps under the 'original' scenario there are more flows that deposit materials overbank, but the flow regime is capable of flushing these areas less regularly than under 'clear' conditions. In essence a floodplain sediment store is accreting, whereas under the reduced flows (dense) more sediment is retained in the channel and available for transport. The net result was an 11% reduction in sediment yield moving out of the catchment perhaps reflecting these differences in sediment storage between simulations and this is a hypothesis explored in later sections.

Approximations of the amount of sediment likely to be deposited across the Tamie fan (500m^2) suggest that under a 'clear' scenario there would be 2.65m of sediment deposited across the fan. Similar calculations for 'original' and 'dense' suggest these amounts would be 1.99m and 2.54m respectively suggesting that the 'original' scenario would trap the least amount of sediment across the fan.

9.3.2 Total Sediment Discharge

The temporal pattern of total annual discharge of fine grain sizes is broadly similar in each of the three land use scenarios for both the Tamie (Fig.9.4b) and the Ire (Fig.9.4e). Increases in sediment discharge are largely in synchrony for the four periods of lowest forest cover (e.g. AD 0-100, AD 300-400, AD 900-1000, AD 1600-1700). The lower sediment delivery identified for the Tamie in the 1973 year not as pronounced under denser forest cover where background sediment delivery is higher. This pattern gives further credence to the hypothesis that fine-grained materials are being stored in overbank settings, whereas the lower discharges of the 'dense' simulations restrict water flow and sediment transport to the channel. As flood discharges are higher in 'original', sediment that is perhaps episodically deposited overbank and removed from the system. In 'Tamie dense', the flood magnitudes are lower perhaps not reaching the magnitudes required for overbank deposition. The channel zone is thus sediment rich and materials are continuously moved out of the catchment. For the Ire, sediment discharges are lower under the most dense land use scenario (Fig.9.2e) and there is virtually no response to the reductions in forest cover AD 1600-1700 and 1800-1900.

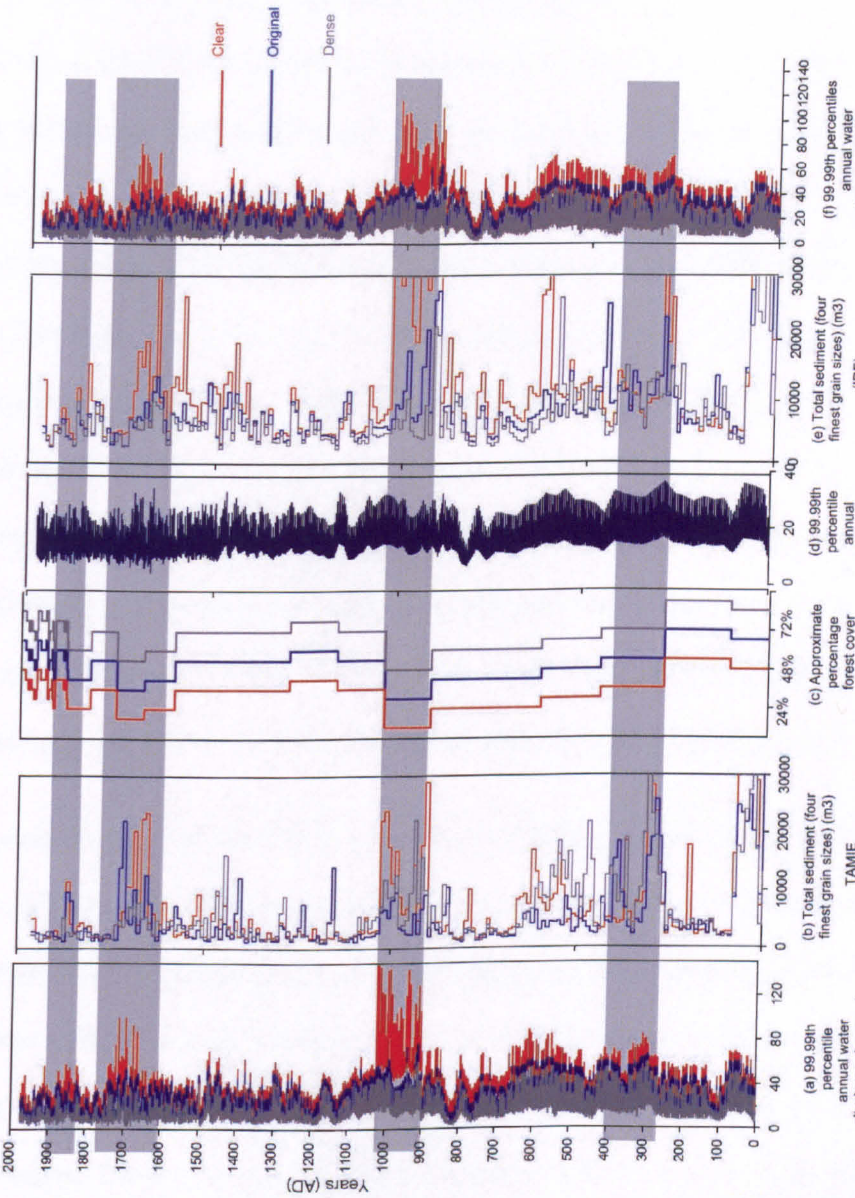

Figure 9.3 *(a) 99.99th percentile annual water discharge Tamie, (b) Total sediment of four finest grain size fractions, (c) Approximate forest cover percentage, (d) 99.99th percentile annual precipitation (e) 99.99th percentile annual water discharge Tamie, (f) Total sediment of four finest grain size fractions,*

The total for the 'original' simulation compared to the 'dense' is replicated in the temporal pattern. The 'original' simulation displays a phasic sequence with episodes of quiescent sediment discharge (e.g. AD 1100-1300). There is also a lack of synchrony in the peaks in sediment discharge for the Ire between the three scenarios between AD 600-450. In different land use scenarios, lower forest cover (controlled by *m*) has increased flood magnitudes and increased rates of hillslope erosion. The discrepancies between the three scenarios in the temporal match of increased sediment delivery in the Ire suggest that the changes in flood magnitude and supply of sediment from slope processes are affecting the way in which the sediment conveyor (Lang et al., 2003) is transmitting materials through the catchment. Varying forest cover up and down in intensity is sufficient to alter cycles of erosion, carrying efficiency, connectivity/coupling, storage and release. The total sediment discharge simply shows net delivery to the edge of the catchment, thus the broad overall pattern is similar, but the finer detail varies thus leading to the small discrepancies between timing of increased sediment delivery peaks.

The results also suggest that there may be a tipping point across the three scenarios, at which point, modelled sediments record a response to change in forest cover. The 'dense' scenario tends not to show any response to the forest cover until it drops below ~55-60% during which time there is a sufficient increase in sediment supply and increase in flood magnitude to increase sediment delivery out of the sub-catchment There is a inert period of sediment delivery at AD 1100-1600 during which time, the 'dense' forest cover is around ~75%. During two periods of reduced forest cover e.g. AD 1670 and AD 900 to ~55-60%, there is an increase in sediment delivery in both sub-catchments. In addition to this at AD 100-250, there is an increase in forest cover to ~55-60% in the 'clear' simulation, during which time sediment delivery is reduced in 'clear' simulations for both sub-catchments. The major exception to this is during high magnitude precipitation events e.g. AD 1150 or wetter periods e.g. AD 300-400 whereby there is an increase in sediment delivery regardless of forest cover. There is also seems to be less difference in sediment delivery during high magnitude precipitation events between the three simulations than when forest cover reduces. At AD 1150 there is a large precipitation event, and there is a response in each of the three land use simulations (Fig.9.3b,d). The difference in volume of sediment between the three simulations in each sub-catchment is approximately 20-30%. However there is a much greater difference between the volume of sediment delivery between the three simulations during reduced forest cover, whereby approximate calculations suggest that the difference between volumes of sediment can vary between 150-350%. This broadly suggests that a high magnitude precipitation event will generate similar volume of sediment delivery regardless of forest cover. However, if forest cover is reduced, the changes in volume of sediment delivery depend and vary greatly on the percentage of forest cover.

9.3.3 Simulated catchment geomorphology

The sediment conveyer style of behaviour of drainage basins is a critical determinant of sediment delivery. CAESAR simulations produce time-stepped elevation models that allow the temporal and spatial pattern of erosion and deposition to be assessed (Fig.9.5). There is clearly a greater area of storage around the channel and in the alluvial fan zone under the 'Tamie Original' scenario (Fig. 9.5b, circled). 'Tamie Dense' (Fig.9.5a) has no storage in this zone but does have some in-channel storage further downstream and in the tributary channel (Fig. 9.5a, circled). It is likely that this stored sediment in the channel will be re-mobilised during the next large flood event as in this scenario the sediment is available to the channel to transport out of the catchment, yet in 'Tamie Original', the sediment is unavailable to the channel as it has been deposited overbank. Therefore, 'Tamie Original' has less sediment available to the channel than 'Tamie Dense' which explains the lower total sediment yield in 'Tamie Original'. The water discharge is likely to have crossed a threshold whereby the flood magnitudes were no longer large enough to deposit the sediment overbank, which results in increased sediment availability within the channel. This is not the case for 'Tamie Clear' which has much larger sediment discharge volumes throughout the time series, particularly during the periods of lowest forest cover due to increased supply of sediment from the slopes and increased flood magnitudes under lower forest cover.

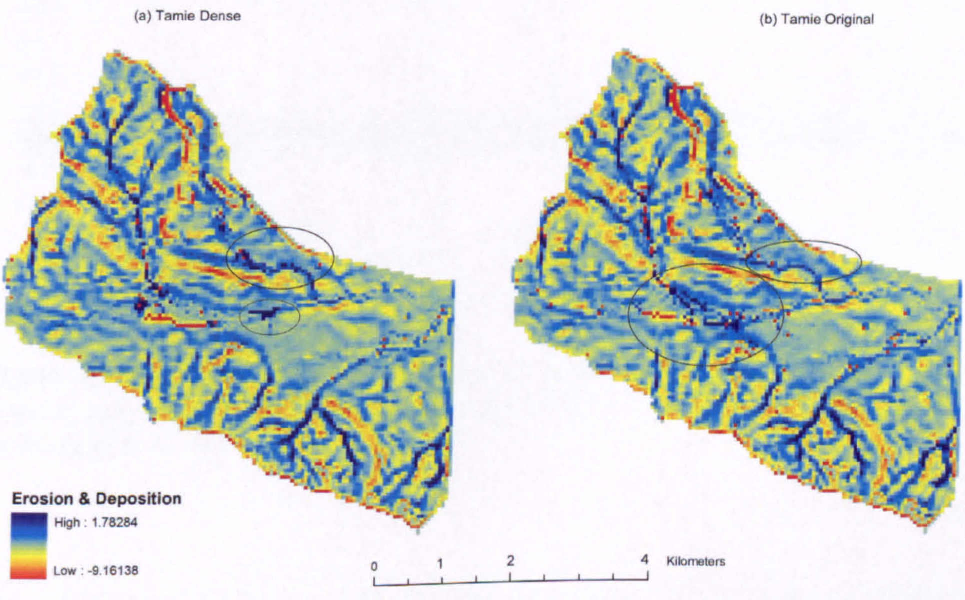

Figure 9.4 *Net erosion and Deposition for Tamie (a) Dense and (b) Original between AD 900 and AD 1000.*

9.3.4 Sediment Accumulation and Impacts on Lake.

Approximate sediment accumulation rates are calculated for each simulation which reflects the amount of sediment accumulation that would take place in a focussed area of the Petit lac d'Annecy assuming no storage of fine-grained deposits on the pre-lake floodplain.

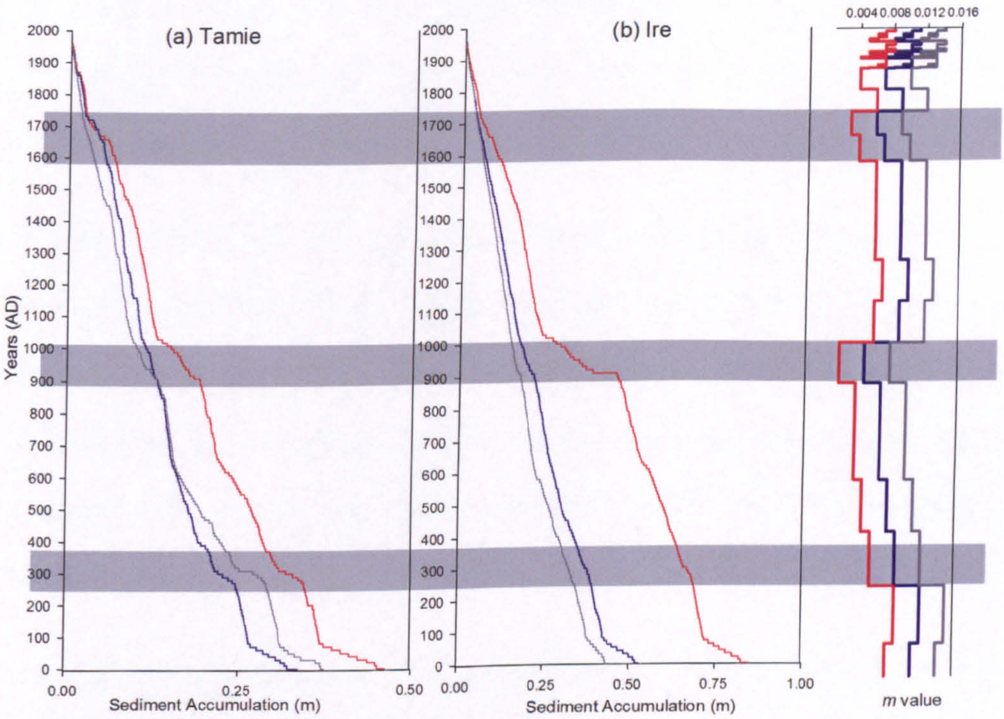

Figure 9.5 *Sediment accumulation rate for each scenario (red = clear, blue = original, grey = dense) for the (a) Tamie and (b) Ire. The time series of m value for each scenario is also shown.*

The sediment accumulations over the 1973 year periods are instructive of sediment responses to changes in forest cover in the three scenarios. Sediment accumulation is faster in the Ire (Fig.9.6b) than the Tamie (Fig.9.6a) in all three scenarios, again suggesting that regardless of forest cover there is always increased storage in the Tamie. The sediment response to decreased forest cover, particularly at AD 900-1000 is clear in each scenario for 'Ire clear', with almost 20cm of sediment accumulation in a 100 year period. Throughout the entire 1973 year series, sediment accumulation rate is much faster in 'Ire clear' than 'Ire original' or 'Ire dense', but this is particularly noticeable between AD 0 – AD 1000. During this 1000 year period, there is 0.5m of accumulation in 'Ire clear' and only 0.2m of accumulation in 'Ire original'. Under the 'clear' scenario the sediment delivery (four finest grain size fractions) is broadly equivalent to adding 0.85m of sediment (Fig. 9.3) to the lake over the 1973 year simulation.

Over the whole 1973 years there would be 0.3m more sediment accumulation if the forest had been 20% lower for a single sub-catchment. If a similar difference in sediment volume occurred in the other sub-catchments, this would equate to an additional ~1.5m of sediment delivery to the lake (assuming no Eau Morte floodplain storage) in the first 1000 years of the 'Ire clear' simulation compared to 'Ire Original'.

Sediment accumulation rate in 'Ire Original' and 'Ire Dense' (Fig.9.6b) are fairly similar and follow largely parallel trajectories through the 1973 year series. The differences between accumulation rates are less noticeable after AD 1600. Both scenarios see increased sediment accumulation rates during the periods of low forest cover, but 'Ire Dense' remains slightly lower than 'Ire Original' and produces ~0.25m less sediment over the series. If all sub-catchments had similar sediment responses to a 20% increase in forest cover (e.g. Ire Dense), there would be around ~1.25m less sediment delivered to the lake (assuming no Eau Morte storage) over the 1973 years.

The pattern of sediment delivery for the three scenarios for the Tamie (Fig.9.6a) is different; with the sediment delivery broadly equivalent to adding 0.46 m of sediment to the lake under the 'clear' scenario. Sediment accumulation is quicker in 'Tamie Clear' scenario than 'Tamie Original' and 'Tamie Dense' and accumulates an additional 0.12m compared to 'Tamie Original' throughout the 1973 year series. Overall, 'Tamie Dense' accumulates around 0.04m more sediment than 'Tamie Original' which is likely to be due to greater overbank deposition in 'Tamie Original' than 'Tamie Dense'.

However, whilst sediment accumulation rates are faster for 'Tamie Dense' than 'Tamie Original' up to ~AD 650, the two sediment accumulation rates then become very similar until ~ AD 900, after which, the sediment accumulation rate in 'Tamie Original' becomes faster than 'Tamie Dense'. This switch may have occurred as 'Tamie Dense' flushed more sediment out of the system during the drop in forest cover between AD 900- AD1000 than 'Tamie Original' (Fig 9.6a). In addition to this forest cover increases at this time, so a fresh supply of sediment to the system in both scenarios is limited. 'Tamie Original' may have had large enough flood magnitudes post AD-1000 to re-mobilise the sediments that had been previously deposited overbank, whereas 'Tamie Dense' has flushed the sediment that was in the channel out of the system during the peak at AD 900 and has no additional supply of sediment as it does not have large enough flood magnitudes to access any sediment stored outside of the main channel. There may be incision in Tamie 'dense' at this point, but reduced flood magnitudes once again render it difficult for the system to move the coarser grained materials out of the channel.

9.4 Conclusions

Here some attempt has been made to quantify the impacts of changing the level of land use in two sub-catchments. The results conclude that:

(1) A 20% reduction in forest cover across all five Petit lac d'Annecy sub-catchments could add a potential ~1.5m of sediment to the lake (assuming no pre-lake floodplain sub-catchment) over 2000 years. This is equivalent to approximately 750 years worth of sediment accumulations (4m in 2000 years in core LA13).

(2) The importance of the 'sediment conveyor' (Lang et al., 2003) has been strengthened here, as the results demonstrate that the timing and magnitude of increased sediment delivery can be a function of alterations in erosion-deposition cycles due to sediment storage, particularly in sub-catchments that have increased opportunity for storage (e.g. wider floodplain, lower gradients). Coupled with this is the issue of sediment supply and channel capacity. Lower forest cover has increased flood magnitudes and increased supply of sediment from the hillslopes. The balance between these can cause sediment to be stored and re-mobilised later in time (decadal-centennial) if the flood magnitudes are larger. Rommens et al., (2006) suggest that sediment sinks can become sources over time depending on changes in climate and human impact and this is clearly demonstrated within these simulations.

(3) Finally Macklin et al. (2006, p.153) question '*to what degree land use must change to amplify or reduce the climate signal in fluvial systems*'. Estimates within this research suggest that if the forest cover is too high (e.g. Tamie 'dense' ~ranges from ~60% - ~90% cover through the simulation), the system reaches a tipping point and flood magnitudes cannot access sediments that are stored overbank, this causes in-channel incision due to lack of overbank access and results in higher sediment delivery overall. In addition to this, all three simulations in both sub-catchments demonstrate that responses to changes in climate and land use occur when forest

cover is less than ~55-60% suggesting this may be the tipping at which slopes and channels may be de-coupled. Below 55-60% cover there is a response to change on the slopes (increased sediment supply due to increased rates) and a response to change in the channel (higher magnitude flood events and increased transport capacity). Above ~60% there is a reduced supply from the hillslopes, reduced flood magnitudes so sediment cannot access over bank deposits and channels begin to incise. These findings could have an impact on catchment management schemes which often promote reforestation to reduce flooding. However, if the catchment is 'too' forested, there will be no supply of sediment from the hillslopes and therefore incision will occur. However, further research to include a wider range of land use scenarios in a wider range of catchment styles would be needed to corroborate these findings. These tentative estimates are only likely to be applicable for the Alpine environment.

Chapter 10

Modelling long term lake sediment records

10.1 Introduction

Whilst facing a future of uncertain climates (IPCC, 2007), modelling has become a widely used and heavily relied on tool to predict how systems will respond to changes. Many researchers have used palaeorecords to talk about analogue conditions e.g. times in the past when there have been similar climatic conditions to the present day. Yet modelling offers the only real alternative to identifying response, behaviours and possible trajectories of systems to future combinations of climate change and human actions. However a fundamental flaw exists within predictive modelling communities: the lack of comparison against empirical data or more precisely, the lack of long-term empirical data to compare model outputs against. The reasons for this shortcoming are explained by the difficulties in accessing records of appropriate resolution, problematic fragmentary records and also instrumental data that do not exist before the last few hundred years. Therefore periods of comparison are short and often not long enough to explore long term changes. The advances in the last 30 years into studies of the palaeoenvironment have ensured we have comprehensive estimates of past climate from multi-proxy records. By combining numerical modelling and palaeorecords, we have long term palaeorecords with which to compare modelled system behaviour against.

The overriding aim of this thesis was to attempt to model a long-term record of sediment delivery from the Petit lac d'Annecy catchment for comparison with proxy records of sediment influx within a continuous lake sediment archive. This would allow behaviour between the modelled record and palaeorecord to be compared with the anticipation that the model could be deemed as robust. The robust model may then be used as a tool for modelling future changes in behaviour and trajectory.

Here, modelled sediment records for the past 2000 years are presented for each of the five sub-catchments in the Petit lac d'Annecy. The main focus of this chapter is identifying sensitivities to climate and land-use divers in different catchment morphologies. The ultimate focus is on comparison of these modelled records to the lake sediment archives.

10.2 Method and Model Setup

The Petit lac d'Annecy DEM was split into five sub-catchments and the large Eau Morte floodplain (Fig.4.3) to aid the speed of modelling and to allow different land use scenarios to be applied where the data permit (e.g. Welsh et al., 2009). Previous chapters have focused on the responses of two contrasting sub-catchments; the Ire and the Tamie. Here the five sub-catchments within the Petit lac d'Annecy are modelled to produce an integrated catchment record of sediment delivery to the lake. Time constraints precluded modelling of the Eau Morte floodplain. An hourly 1973 year long record of precipitation from bog surface wetness data (Charman et al.,unpublished) has been modulated against a 1973 year long multi-proxy temperature record (Moberg et al., 2005) and adjusted to imitate snow store and snow melt. A pollen record for the same period from the Petit lac has been used as a proxy for land use (Jones et al., in prep.). For each sub-catchment, the precipitation and land use have been kept constant, therefore any changes in sediment behaviour or delivery is a function of catchment morphometry. The total sediments from the five simulations are compared to proxies for sediment flux from the lake sediment record.

10.3 Temporal and spatial sediment discharge patterns in sub-catchments

10.3.1 Total Sediment Yield

The difference between total sediment yields of the four finest grain size fractions (63μ, 250μ, 500μ, 1000μ) over 2000 years (Fig.10.1a) is somewhat varied between each sub-catchment, and is largely a function of catchment size, although the sediment yields for each sub-catchment lie within the same order of magnitude. As expected the largest sub-catchments, the Tamie and Ire, deliver the most sediment, yet the smaller Montmin sub-catchment delivers a substantial amount of sediment; more than the larger St Ruph suggesting that the sediment conveyor (Lang et al., 2003) Montmin sub-catchment may be potentially more efficient than the St Ruph.

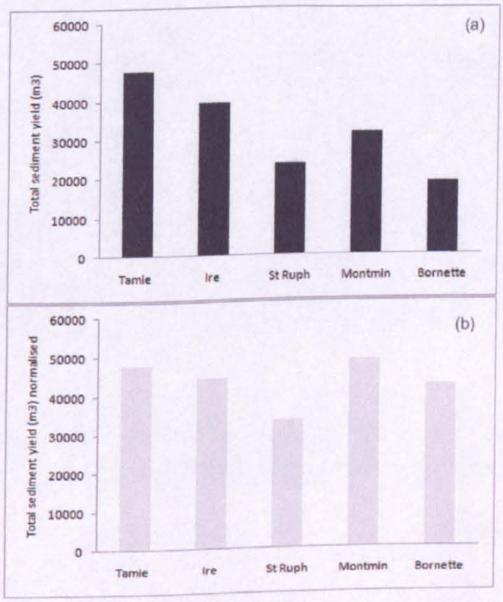

Figure 10.1 *(a) Total sediment yields of the four finest grain size fractions for individual sub-catchments over 2000 years in order of catchment size (largest to smallest),(b)Total sediment yields of the four finest grain size fractions normalised by Tamie catchment area for individual sub-catchments over 2000 years in order of catchment size.*

When normalised by catchment size (Fig.1b), it appears that the Tamie, Ire and Montmin sub-catchments are similar in terms of the efficiency of sediment delivery, with the Bornette only slightly less efficient. The St Ruph sub-catchment is the least effective at moving the finest grained sediment out of the system over the 1973 year period. The pattern is also the same for total sediment yield (not shown). The lower efficiency of the St Ruph is fairly consistent with present day field evidence (Fig.10.2) which shows that despite slopes being reasonably well coupled to the channel, high volumes of sediment are stored in multiple gullies and also within the channel, therefore the sediment supply may often exceed the capacity of the channel to transport sediment efficiently.

Figure 10.2 *(a) Large sediment-rich gully coupled to the main stream of the St Ruph, (b) High volume of sediment stored in the main channel of St Ruph*

10.3.2 Total Sediment discharge

10.3.2.1 Overall patterns

Overall, in terms of temporal patterns of total sediment discharge, the Ire and Montmin sub-catchments behave similarly in terms of sediment discharge patterns and response to changes in land use and precipitation (Fig.10.3 k-l.). They both deliver a high amount of background sediment (i.e. periods of continuous but lower discharge such as AD 600-900) throughout the entire time series, unlike the other sub-catchments. The Tamie (Fig.10.3a) behaves similarly but the sediment delivery is sometimes lagged and the background sediment discharge is much lower relative to peaks in sediment delivery. The Tamie is frequently out of temporal synchronicity when compared to the other sub-catchments which may be a function of storage and release cycles within this sub-catchment due to a much wider, flatter and lower gradient floodplain than the other sub-catchments. The Bornette and St Ruph sub-catchments (Fig. 10.3e&i) behave differently to the other three sub-catchments. Both have lower total sediment discharge relative to the other sub-catchments (Fig.10.1b) and background sediment discharge is significantly lower relative to peaks in sediment delivery in both sub-catchments. The style of sediment delivery in the St Ruph and Bornette is described best as phases of quiescence punctuated by high magnitude sediment delivery which may be a function of sediment supply exceeding capacity of the channel or storage and release within the systems.

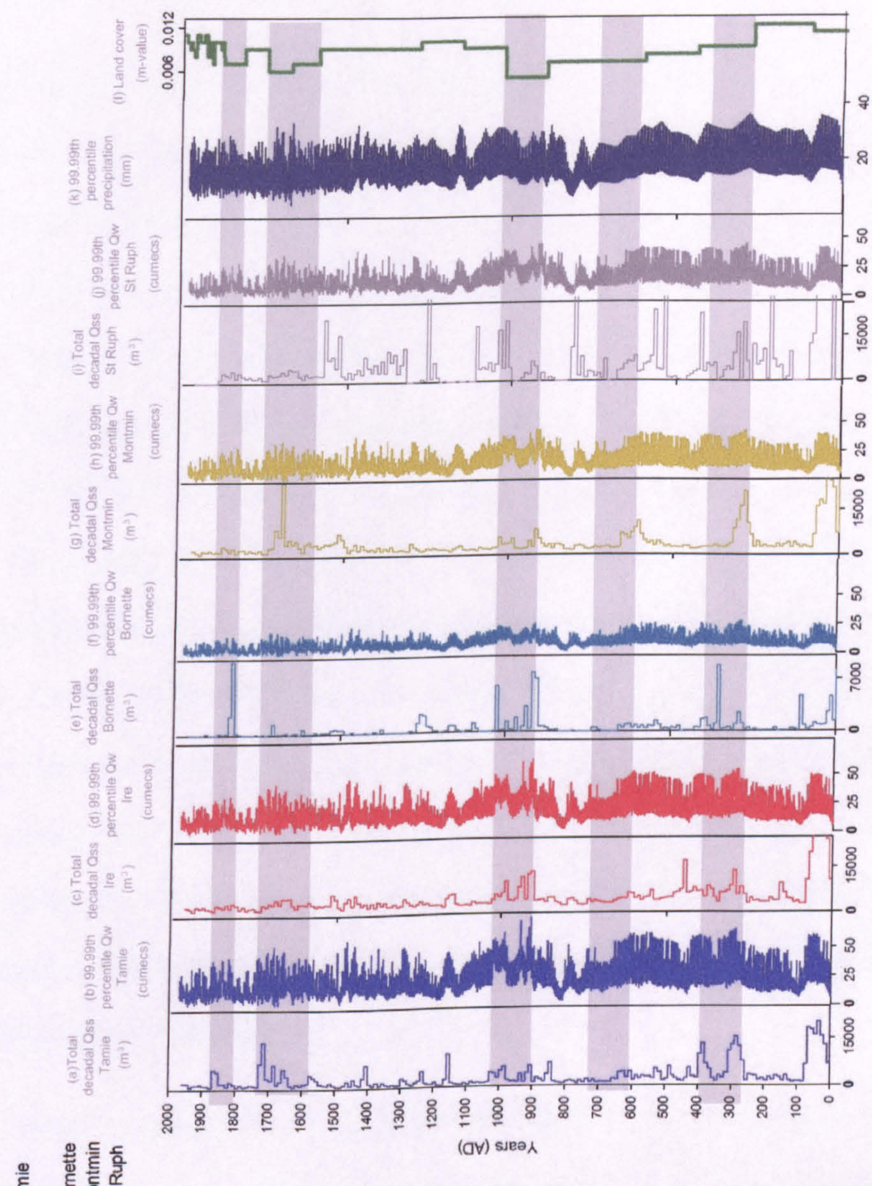

Figure 10.3*(a) Total decadal sediment of 4 finest grain size fractions Tamie (b) 99.99th percentile water discharge Tamie (c) Total decadal sediment of 4 finest grain size fractions Ire (d) 99.99th percentile water discharge Ire (e) Total decadal sediment of 4 finest grain size fractions Bornette (f) 99.99th percentile water discharge Bornette (g) Total decadal sediment of 4 finest grain size fractions Montmin (h) 99.99th percentile water discharge Montmin (i) Total decadal sediment of 4 finest grain size fractions St Ruph (j) 99.99th percentile water discharge St Ruph (k) 99.99th percentile annual precipitation (l) forest cover in m-value terms *NB scale for Bornette total sediment discharge is lower than other sub-catchments*

10.3.2.2 Sensitivity to land use change

There are 4 periods of substantial (15%-20%) forest cover reduction at AD 280-400, AD 900-1000 AD 1600-1700 and AD 1800-1880. The sediment discharge from the Bornette sub-catchment (Fig. 10.3e) seems to be the most responsive to changes in land use. This may be a function of increased water discharge (Fig.10.3f) during lower land use, and therefore the capacity of the channel is increased and may cross a threshold to enable sediment transportation out of the sub-catchment during these periods. Whilst the sediment delivery in other sub-catchments increases during *some* land use reductions, the sediment delivery in the Bornette increases during every phase of reduced land use. However the magnitude of total sediment delivery during these periods is around a quarter of other sub-catchments such as Montmin (Fig.10.3g) which also seems to be sensitive to changes in land use throughout the 1973 years. In fact, Montmin and Bornette (Fig.10.3e&g) are the only sub-catchments that remain sensitive to changes in land use throughout; this is particularly true in the Montmin as the increases in sediment delivery are more pronounced than in the Bornette.

The sediment discharge in the Ire (Fig.10.3c) seems to be most sensitive to the reduction in land use at AD 900-1000 where the Tamie (Fig.10.3a&g) and Montmin are less sensitive. This may be due to an increase in water discharge magnitudes for this period in the Ire (Fig.10.3c). Whilst water discharges also increase in other sub-catchments under lower forest cover the Ire has increased substantially and may have crossed a threshold enabling an increase in capacity to transport the increased supply of sediment that becomes available under lower forest cover. The Tamie also had a substantial increase in water discharge (Fig.10.3b), yet it is likely that much of the sediment mobilised during this phase has been stored and may be re-mobilised under the next large discharge event. This seems to be true in AD 1158 as there is an increase in sediment delivery that coincides with the precipitation event, yet there are no increases in sediment delivery at this point in other sub-catchments (Fig.10.3a). Increased water discharge in the Bornette (Fig.10.3f) may also have re-mobilised sediments at AD 900-1000 that may have been previously stored in the channel or low lying narrow flood plain. Increases in sediment discharge in the St Ruph sub-catchment do coincide with some reductions in forest cover, but good correspondence with drivers is not evident as there are many phases of increased sediment discharge throughout the St Ruph series, which suggests that there may be a complex or large scale storage and release cycle here.

After the first millennia, the Ire and Bornette have fewer periods of increased sediment delivery, and overall the background sediment discharge reduces somewhat in the Ire, which may be a function of reduced sediment supply as a result of increased land use after AD 1050. This reduction is not seen in the Tamie, Montmin or St Ruph simulations, perhaps because the stored sediments are being re-mobilised from the floodplain in these sub-catchments.

10.3.2.3 Sensitivity to precipitation

Annual precipitation maxima occur at AD 1664, 1468, 1153, 878, 588, 533, 403, 273, and AD 58 (Fig.10.3k, Table 10.1). These values are derived from the annual record of 99.99th precipitation percentiles, and represent the highest 0.01% of precipitation magnitudes over the 2000 year period. All sub-catchments respond to the largest precipitation events (Table 10.1 *denotes largest events) even if some of the sub-catchments have either slightly lagged responses or smaller responses. The only exception is the Ire (Fig.10.3c) in AD 878, where intrinsic factors may dominate instead. Some of the other slightly smaller high magnitude precipitation events have little if any effect on sediment delivery particularly in the Tamie and Ire.

Years (AD)	Tamie	Ire	Montmin	Bornette	St Ruph
1664*	Y	YS	YL	YL	YS
1468	N	N	YS	N	YL
1153*	Y	Y	Y	Y	N
878*	YS	N	YS	YS	YS
588	YL	YS	Y	YS	N
533	N	N	YS	Y	Y
403	N	N	YS	Y	YL
273*	Y	Y	Y	YL	Y
58	Y	Y	Y	Y	Y

Key

Y - substantial response

YS - response but small

YL - response but lagged

N - no obvious response

***Denotes largest precipitation events.**

Table 10.1 Major precipitation events throughout the 1973 year series and the sediment response in each sub-catchment.

Montmin (Fig.10.3g) seems most sensitive to extreme annual precipitation values and there is relatively instant (within the decade) response in sediment for all of the precipitation events except for AD 1664 where it is lagged by 30-40 years. The Bornette (Fig.10.3c) and St Ruph (Fig.10.3i) also have increased peaks in sediment discharge when precipitation is high, but they are more frequently lagged and at a much lower magnitude than in Montmin. Montmin is the most efficient system for carrying sediment out of the system (Fig.10b). This may be a function of steep slope gradients and long slope lengths increasing sediment supply in the model due to increased erosion rates. The gullies and channel within Montmin are also well coupled and the main channel flows through a narrow bedrock gorge above the main valley floor with little opportunity for sediment storage.

The temporal sediment discharge patterns of the Bornette moves in synchronicity with the other systems (except for the St Ruph) but the largest sediment discharges here are during some periods of reduced forest cover or increased precipitation. Conversely, the St Ruph has extremely large sediment discharge flushing events e.g. AD 800, AD 1250 which rarely show correspondence to changes in land use or increased precipitation suggesting that the intrinsic system controls may be more dominant here than external forcings.

10.3.3 Sediment Accumulation Rate

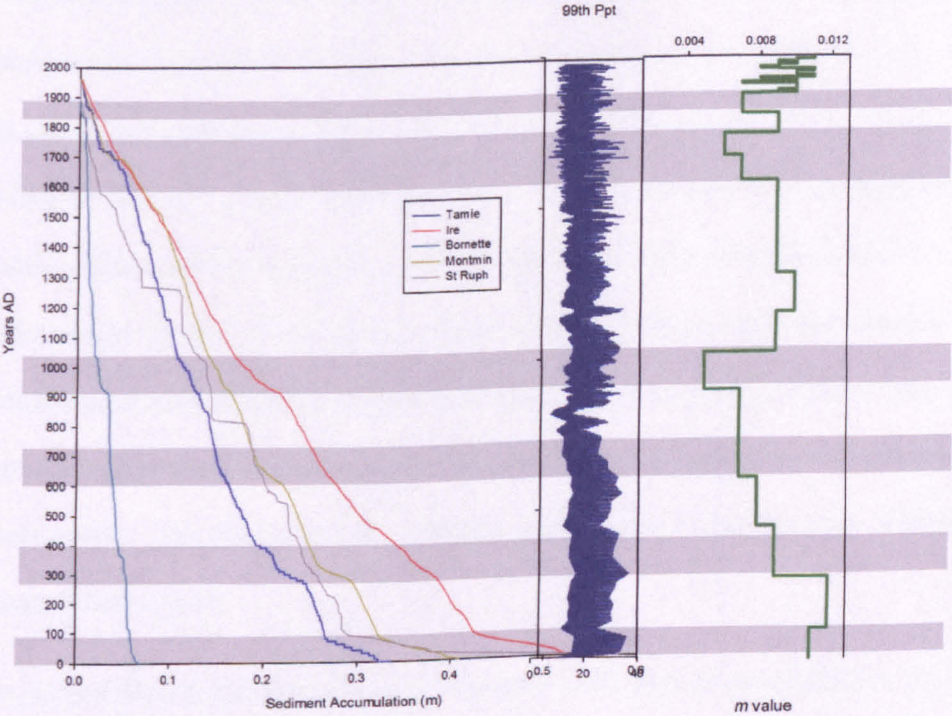

***Figure 10.4** Sediment accumulation of each sub-catchment compared against 99^{th} percentile of precipitation and land use.*

In order to compare these modelled sediments to sediment accumulation rates from the Petit lac d'Annecy, a hypothetical sediment accumulation rate for each of the sub-catchments to a focussed area of the lake ($2km^2$) has been calculated assuming no storage on the flood plain to provide an estimate of sediment accumulation. The Ire and St Ruph (Fig.10.4) sub-catchments accumulated over 0.5m of sediment each over the 1973 year period, followed by the Montmin sub-catchment which accumulated 0.4m. The Tamie and Bornette accumulated ~0.35m and ~0.07m respectively.

The early sediment accumulation rate was fastest in the Ire and Montmin which is likely to be due to sediment delivery from the well coupled gullies to the main channel. Hillslope processes may also be enhanced here as there are long, uninterrupted slopes which increase soil erosion rates (chapter 7). Although the Tamie is the largest sub-catchment, there is less sediment accumulation from this sub-catchment which is likely to be a function of storage and release cycles of sediment due to the catchment morphometry. This is corroborated by the marginally concave nature of the curve after ~1300 which suggests that sediment accumulation is more rapid. It likely shows that the Tamie still has plenty of sediment 'sources' to access even during increased forest cover, suggesting that the other systems may be more affected by the supply-limited nature of sediment associated with increased forest cover. The Bornette is the smallest sub-catchment and has least sediment accumulation overall.

The main difference between the sub-catchments is the style of accumulation; the Ire had a mostly constant accumulation rate throughout the time series, with very slight increases in accumulation during the period of low forest cover around AD 900-1000, yet the St Ruph had periods of slow accumulation followed by intense periods of rapid accumulation, at ~AD 800. This may have been a response to precipitation, whereby flood magnitudes increased and sediment could be flushed from the large stores. This creates a much more stepped profile for the St Ruph (Fig.10.4) sediment accumulation suggesting the magnitude of the storage and release cycles are large.

Similarly to the total sediment discharge graphs (Fig.10.3) the Bornette shows similar responses to reductions in land use whereby the accumulation rate of sediment is increased. The increase in sediment accumulation is particularly noticeable at AD 900-1000 but also during the reduction in land use at AD1800-1850 suggesting once again that the Bornette may be the most sensitive sub-catchment to changes in land use. The sediment accumulation rate in the Tamie and Montmin (Fig.10.4) both increase during periods of low forest cover but are more sensitive before AD 600 and after AD1600. The increases in sediment accumulation rate at AD 100, AD 300, AD 1600 and in the case of the Tamie AD 1800 were much more pronounced than during the reduction in land use between AD 600-700 and AD 900- AD 1000 where small increases in accumulation took place. This may be because land use was lower in the first millennia through the period of AD 280- AD 1000, therefore background sediment would have increased, and periods of increased sediment accumulation may be masked. The comparisons of the totalled five sub-catchment sediment accumulation rate and LA13 lake sediments are discussed in section 10.4.

10.3.4 Net erosion and deposition

In addition to time series of sediment and water, CAESAR outputs raster data of elevation difference over time, allowing net erosion and deposition to be identified.

Even general patterns of net erosion and deposition over millennia can aid explanations of behaviour exhibited in time series.

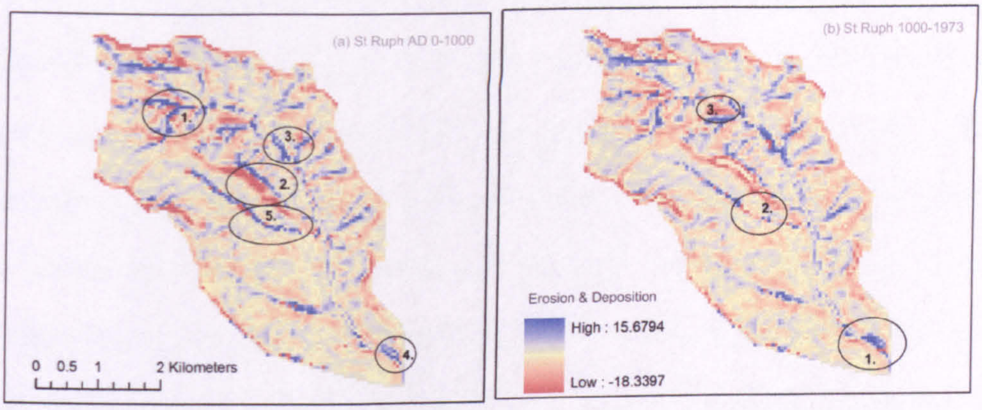

Figure 10.5 *(a) Net erosion and deposition in St Ruph between AD 0-1000, (b) Net erosion and deposition in St Ruph AD 1000-2000*

Time series of sediment discharge has shown that the St Ruph operated out of synchronicity when compared to the other sub-catchments. The peaks in sediment delivery out of the catchment were sometimes short episodes of high magnitude discharge out of the catchment, followed by periods where there was very little sediment moving. It is suggested that this temporal pattern is likely to be a function of storage and release cycles, the CAESAR spatial maps of erosion and deposition can help the understanding of temporal patterns. In the first millennia, there is up to ~4m of sediment storage in gullies (Area 1: Fig.10.5a), tributary streams (Area 5: Fig. 10.5a) and in the main channel (Area 3: Fig. 10.5a) and a large area (240000m^2) of intense hillslope erosion of up to 3m on a steep gradient, which is likely to be a large source of sediment supply for the sub-catchment. This source of sediment, coupled with erosion in watershed zones is likely to have provided the main channel with massive amounts of sediment that it did not have the capacity to carry.

Consequently, this has led to high volumes of sediment being deposited in the main channel on a decadal basis (not shown), which has ultimately reduced the efficiency of the river to move sediment out of the sub-catchment. Eventually the river reached much lower gradients (Area 4, Fig.10.5a) and sediments accumulated in this zone on a cyclic basis and were then flushed out of the sub-catchment periodically. This perhaps explains why the total decadal sediment discharge time series of the St Ruph is characterised by the low magnitude background sediment punctuated by high magnitude sediment peaks. However the peaks in sediment delivery (Fig.10.3i) do not coincide with peaks in precipitation or land use, which is not to suggest that the St Ruph is not responsive to changes in input drivers, but that the volume of storage so close to the edge of the DEM renders is it impossible to find signals of response in the St Ruph as they are masked by massive sediment delivery from release events. Shorter time step DEMs (now shown) show that under lower forest cover or increased precipitation, hillslope processes rates are intensified, but the increased supply of sediment accumulates in Area 4 (Fig.10.5). The second millennia (Fig.10.3i) shows a reduction in both overall sediment and high magnitude sediment discharge events; this may be a function of reduced sediment supply as forest cover increases from AD 1000 to AD 1550, however areas that had previously been zones of deposition now erode more which suggests remobilisation of sediments from the hillslope zones (Area 3: Fig.10.5b). The tributary channel, Ruisseau de Nant (Area 2: Fig.10.5b) is also now eroding which suggests that capacity of the channel is greater than supply of sediment over this period. The zone of deposition for the second millennia is a much greater area and had a much greater fill of sediment of up to 15m of net sediment accumulating over 1000 years (Area 3:Fig.10.5b).

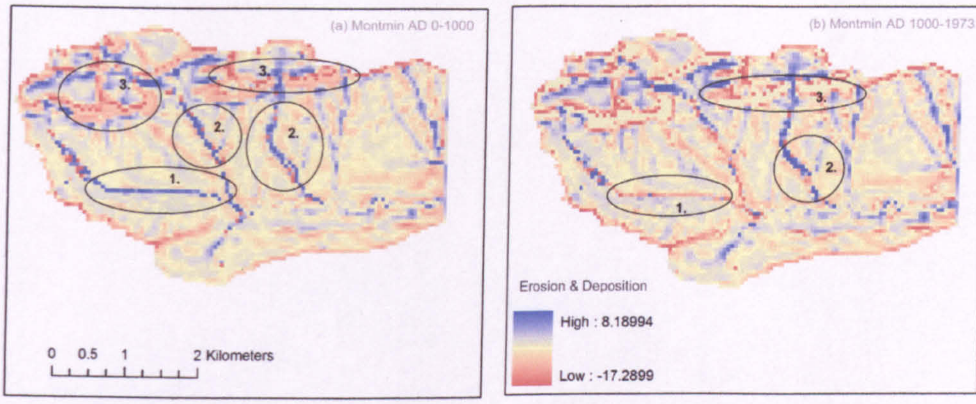

Figure 10.6 *(a) Net erosion and deposition in Montmin between AD 0-1000, (b) Net erosion and deposition in Montmin AD 1000-2000*

The first 1000 years in the Montmin sub-catchment is characterised by large in-channel and gully accumulation (Areas 1-2: Fig.10.6a) due to a high sediment supply from the steep slopes from the slopes below La Tournette, the highest point of the entire Petit lac d'Annecy catchment (Areas 3: Fig.10.6a). The second millennia sees a reduction in sediment supply from the steep slopes (Area 3: Fig.10.6b) as sediment as a source is likely to be depleted on these steep slopes, additionally there is an increase in land use which has reduced the rate of hillslope processes and therefore reduced supply of sediment. Part of the main channel in the second 1000 years is now eroding, with the downstream part of the channel eroding and depositing equally as values here are close to zero (map shows net erosion and deposition). The largest gully still has up to 4m of sediment accumulation. This gully is coupled to the main channel which provides a route for sediment from the steep slopes (Area 2: Fig.10.6b). Overall, the gullies and main channel are well coupled, providing constant sediment supply from the slopes, explaining why the background sediment discharge is high throughout the 2000 years in the time series (Fig.10.3g).

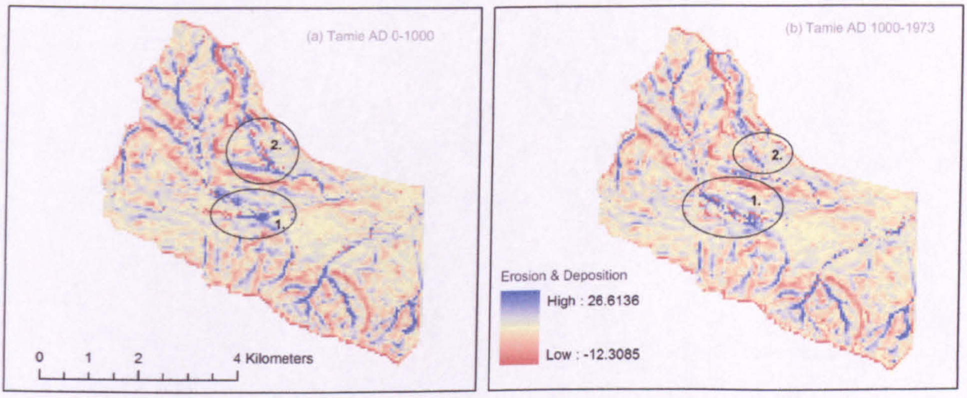

Figure 10.7 *(a) Net erosion and deposition in Tamie between AD 0-1000, (b) Net erosion and deposition in Tamie AD 1000-2000*

The first thousand years in the Tamie is characterised by a large area of up to 4m accumulation in the Tamie fan (Area 1: Fig.10.7a). The fan is particularly effective at trapping sediment during this period, most likely due to a constant supply of sediment from increased hillslope erosion rates due to lower forest cover at this period. There is also a substantial amount of sediment stored in the tributary stream (Area 2: Fig.10.7a) and some degree of erosion here. The second millennia had less sediment accumulation on the fan, and the channel has begun to incise into the accumulated material. This is likely to be due to a reduction in supply of sediment from hillslopes due to increased land use during this period. The channel therefore begins to incise as the capacity of the channel is greater than the supply of sediment (Area 1:Fig.10.7b). The amount of accumulation in the tributary is less in the second millennia (Area 2: Fig.10.7b), which is a function of reduced slope processes and increased capacity relative to supply resulting in increased erosion.

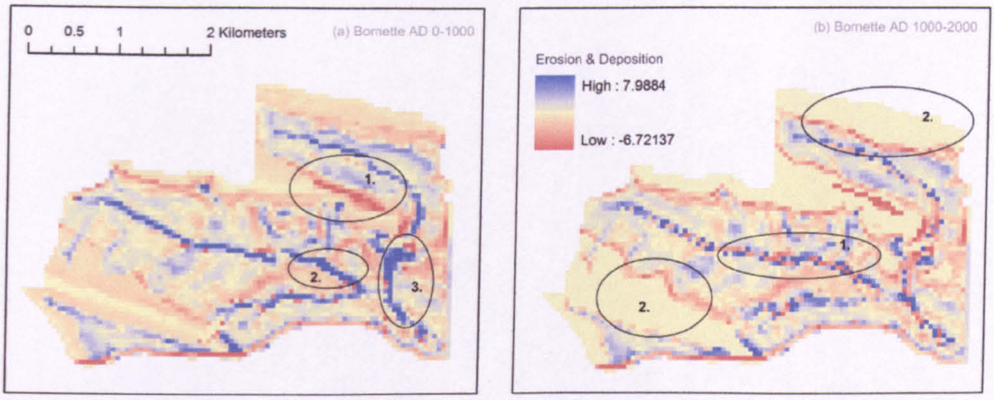

Figure 10.8 *(a) Net erosion and deposition in Bornette between AD 0-1000, (b) Net erosion and deposition in Bornette AD 1000-2000*

The first millennia in the Bornette is characterised by erosion of up to 6m on the steepest slopes (Area 1:Fig.10.8a) and in channel storage of up to 7m (Area 2-3: Fig.10.8a). The storage within the channel may be a function of lower discharges in the Bornette, therefore the capacity cannot move the amount of material supplied from the hillslopes, and accumulation takes place. This corresponds to field evidence in the catchment whereby sections of alluvium were at least 4m deep. The sediment discharge time series for this sub-catchment (Fig.10.3e) suggests that this sub-catchment is the most sensitive to changes in forest cover. Reductions in forest cover brings increased sediment supply from slopes and subsequently increased sediment delivery (Fig. 10.3e) out of the sub-catchment. It also suggests that the increased flood magnitudes under lower land use cross a threshold whereby some of the sediment can be moved out, as capacity exceeds supply. For the second millennia, there is much greater erosion in the main channel (Area 1: Fig.10.8b), which is likely to be a function of reduced sediment supply from the hillslopes (Area 2:Fig.10.8b), therefore capacity exceeds supply. There is still some in-channel storage, but it is more sporadically distributed than the first millennia whereby the entire channel was accumulating sediment.

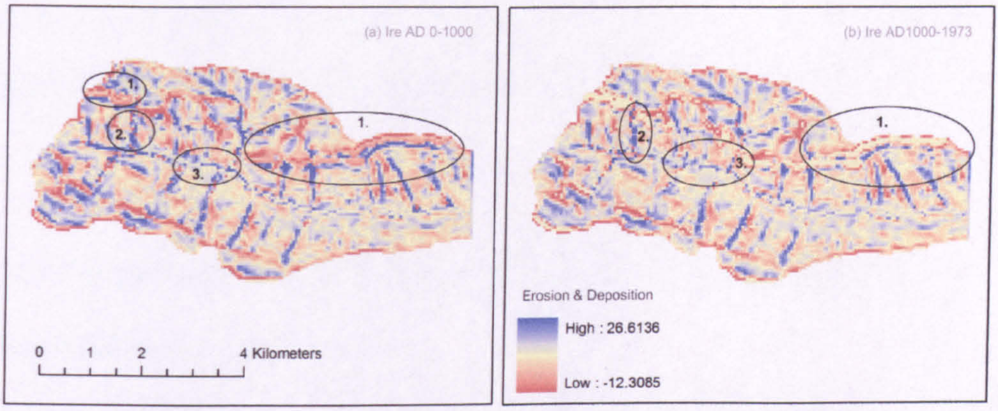

Figure 10.9 *(a) Net erosion and deposition in Ire between AD 0-1000, (b) Net erosion and deposition in Ire AD 1000-2000*

The first 1000 years in the Ire sub-catchment is characterised by high sediment supply from eroding steep slopes (Area 1: Fig.10.9a). The channel is largely eroding and depositing material equally, yet adjacent to the channel, there is a large amount of overbank deposition of up to ~3-4m which is available for the tightly constrained channel to re-mobilise during large flood events (Area 3: Fig. 10.9a). The gully systems were also well coupled to the channel and had accumulated up to 3-6m of sediment (Area 2: Fig. 10.9a), providing the system with a constant supply of sediment through the 1000 year period. The combination of sediment supply and well coupled gully systems to the main channel go some way to explain why there is a high background sediment discharge in the Ire (Fig.10.3c). The time series suggests that there is a reduction in background sediment supply in the latter 973 years most likely due to a reduction in supply of sediment from the steepest slopes (Area 1: Fig.10.9b), and less material(~1-2m) available for re-mobilisation adjacent to the channel due to channel incision (Area 3:Fig.10.9b) and in gullies (1-2m) (Area

2: Fig.10.9b), suggesting that these depocentres may have been a source of sediment supply through the latter millennia via re-mobilisation of sediments.

10.4 Comparison of CAESAR modelled data to lake sediment proxies

10.4.1 Sediment accumulation rate

A sediment accumulation rate curve was calculated for the modelled sediment discharge data by summing the total decadal sediments from each sub-catchment and dividing by a focussed abyssal area of the lake ($2km^2$) (assuming no flood plain storage). LA-13 sediment core contains up to 2% of total organic carbon and 20-60% biogenic calcium carbonate which increases the sediment accumulation in the lake by this amount (Fig. B). As CAESAR modelled sediments are a detrital signal, the biogenic calcium carbonate and total organic carbon from the LA-13 core have been removed to provide a best estimate of mineral matter from the lake sediments for comparison with the CAESAR modelled data.

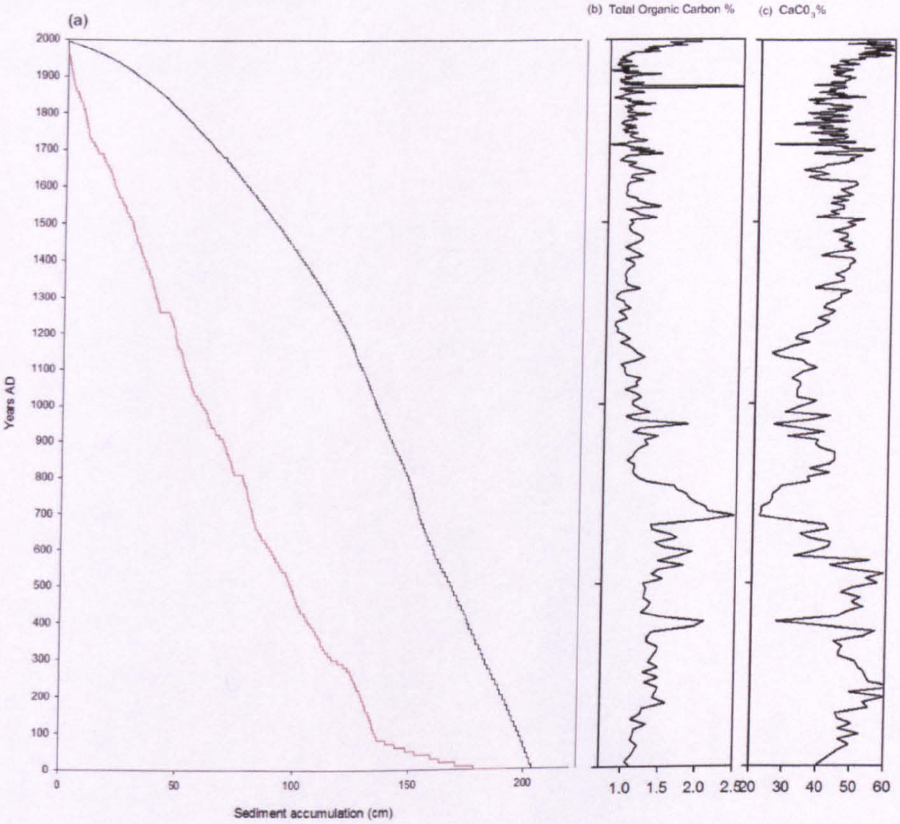

Figure 10.10 *Comparison of CAESAR modelled sediment accumulation in a focussed area of the lake (red) and sediment accumulation rate of mineral matter from LA13 core*

Figure 10.10a shows a comparison of the sediment accumulation rate from total CAESAR sediments (195cm) and LA-13 sediment accumulation (203cm) (corrected for ToC% and $CaC0_{3\%}$). The difference between the two records is ~8cm over 2000 years. The style of accumulation is similar for both, the CAESAR modelled data is slightly more stepped than the LA-13 data but both lie within the same order of magnitude as each other.

Magnetic parameters can be useful proxies of environmental change through time. Welsh et al. (2009) made use of χpara as a proxy for detrital sediment supply from a catchment, particularly as it avoids complications from autocthonous biogenic activity within the lake such as bacterial magnetosomes and iron sulphates dominating the sediment magnetic record (Dearing, 1999; Dearing et al., 2001).

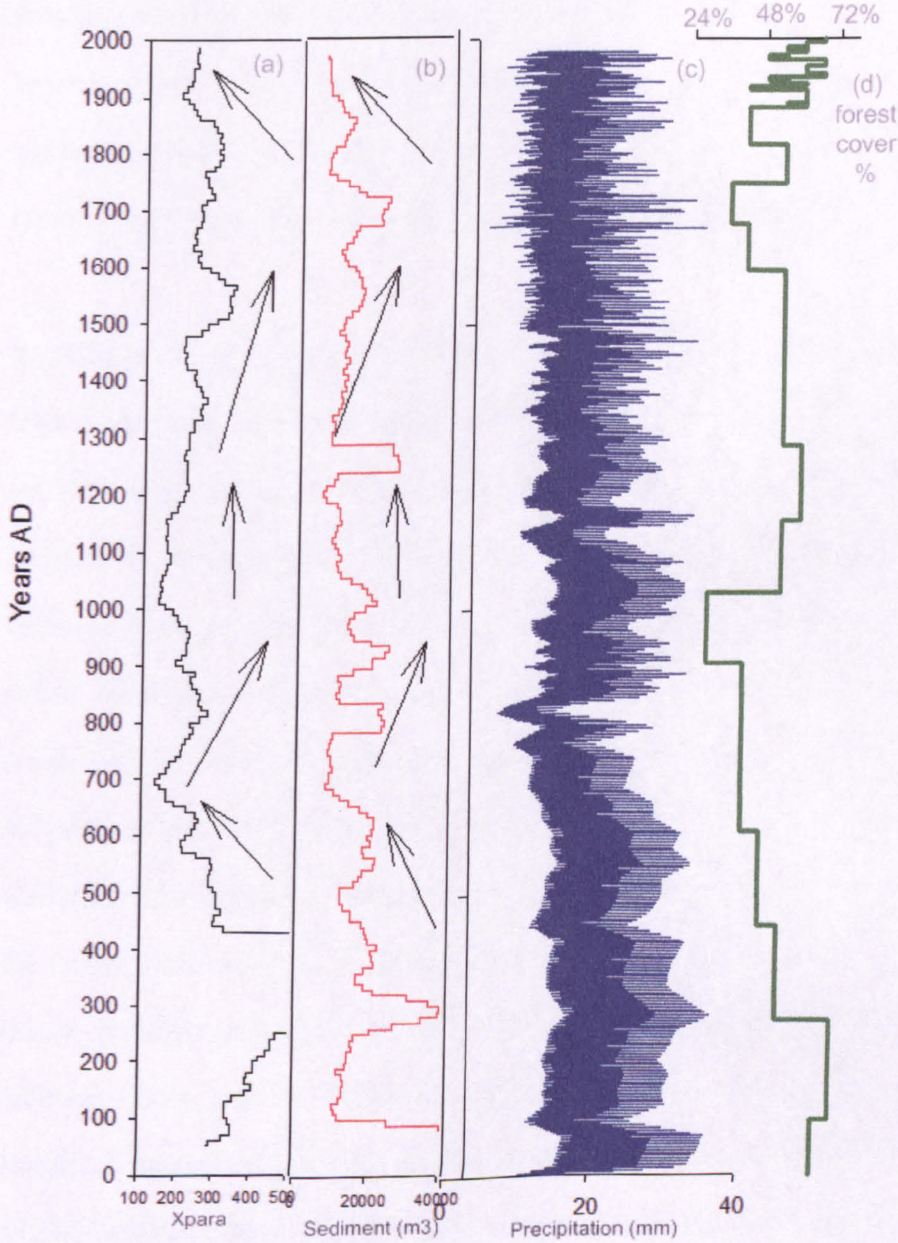

Figure 10.11*(a) χpara% five-point moving average, magnetic proxy from LA13 (Jones et al., in prep) (b) five point moving average of total sediment (four finest grain size fractions) from CAESAR (d)Annual maximum precipitation (d) forest cover percentage.*

Comparatively the overall behaviour and trends between CAESAR modelled sediments and χpara is good (Fig.10.11). In both records, the period AD 0-600 is characterised by the highest sediment flux throughout the 1973 record. Both curves then follow a downward trend to AD 800 before both CAESAR sediment and χpara increase slightly. The overall trajectory is reduced sediment delivery in CAESAR and a reduction in χpara until around AD 1200, followed by an increase in both curves to AD 1600. After AD 1600, the two curves are less comparable.

In addition to the trajectories of change being similar, the overall behavioural response to changes in land use and climate are similar. Between AD 200 and AD 400 there is an increase in magnitude of peak precipitation events and a reduction of land use by 20%, at which time both, sediment flux recorded in CAESAR and the values of χpara increase and result in the largest sediment peaks of the 1973 year series. Reductions in forest cover at a similar magnitude (~20%) at AD 900-1000 combined with lower magnitude peak precipitation events than AD200-400 produce the peaks in sediment and increased χpara values but both are much lower than the earlier event, suggesting that the largest peaks in sediment are likely to occur under the lowest forest cover AND highest precipitation. Perhaps low forest cover may induce sediment peaks, but without high magnitude precipitation events, these sediment peaks may be reduced in magnitude. Consecutive years with high magnitude precipitation events may also determine peaks in sediment delivery – as AD200-400 has many consecutive years of high precipitation and higher magnitude sediment peaks whereas other phases that have low forest and high precipitation such as AD1600-1700 have high magnitude rainfall events in single years but slightly lower magnitude peaks. Other similar behaviour shows that after AD1150-1280, both χpara and CAESAR modelled sediment experience low values, this is likely to be a response to the increase in land use at this time. The overall behaviour of the CAESAR modelled sediments suggests that there is increased sediment delivery during periods of low land use; a characteristic of the short cores of lake sediments taken by Foster et al., (2003) whereby there is increased sediment during periods of lower forest cover.

10.5 Conclusions

The main conclusion to be drawn out of these results is that CAESAR can produce modelled sediment records that are comparable to lake sediments in terms of behaviour, trajectory, magnitude and apparently the timing of process-response features, like major peaks in sediment flux, over decadal-centennial timescales.. Errors in the radiocarbon chronology for the lake sediments renders any closer comparison between model outputs and the sediment record invalid. This gives us confidence that CAESAR can be robustly used as a tool to model future system behaviour and likely trajectories of change in response to changing climate and land use.

The modelled time series of total sediment delivery has allowed general behavioural trends to be identified for each sub-catchment; some of which is a characteristic of all sub-catchments. The results suggest that the Montmin sub-catchment is most sensitive to changes in precipitation and land-use, which suggests that the effects of future climate and land use changes may be most evident in catchments that have a high erosion rates on long, steep slopes that provide a rich supply of sediment to gullies that are well coupled to the main channel, and in channels where little/no in-channel deposition can be deposited. The responses may be most clear in these highly efficient systems, and may suggest likely future trajectories, which renders them the most predictable type of system from time series alone.

However, systems such as the Tamie or St Ruph that have large flat zones of accumulation, coupled with complex cycles of storage and release are the least predictable from time series alone. There may be a response such as increased hillslope erosion rates at the time of changes in climate or land use, but the increased sediment may be stored overbank, or trapped in a fan and may not be re-mobilised until flood magnitudes cross a threshold whereby the can be accessed once more. By using spatially distributed numerical models that provide time-stepped net change in erosion and deposition, these response can be detected on hillslopes or in channels and zones of accumulation can be identified. The model outputs suggest time-lags in the order of 100 years between significant accumulation in storage zones and release downstream as a result of a disproportionately low river discharge event. This is particularly true for the St Ruph. Some of the largest peaks in sediment re-mobilisation coincide with precipitation events.

During periods of high forest cover, there is both a reduced supply of sediment to the system and a reduction in flood magnitudes. The immediate response under these conditions is to store the available sediment due to lower channel capacity (reduced sediment discharge seen in time series). However, the longer term response suggests that over time (decades/centuries) the stored sediment is slowly eroded and re-mobilised by the channel during high magnitude precipitation events and the system can move to an eroding system i.e. when supply of accumulated material is depleted; the system erodes to provide fresh material for transport. At the same time, sediment supply is increased due to increased rates of hillslope erosion (wetter conditions) and water discharge magnitudes increase. The high magnitude precipitation events seem to allow flood magnitudes to cross a threshold and channels can re-mobilise previously stored sediments. This is true for both narrow catchments with little storage, or in low-gradient, wider catchments where there is greater opportunity for storage.

During periods of low forest cover, there is both an increase in supply of sediment to the system and an increase in flood magnitudes. The response of the 'well-coupled' style sub-catchments (e.g. Ire, Bornette, Montmin) is that of increased sediment delivery out of the catchment. For the 'wider, lower gradient' sub-catchments, the response is different. There is some element of increased sediment delivery, but the capacity or stream power of the channel in these systems is too low to transport all of the material, therefore storage takes place. The stored sediments are often re-mobilised under high flood magnitudes after several decades which are caused by high precipitation magnitudes.

In summary, land use determines the supply of sediment and can increase/decrease channel capacity. Although land use does play some role in determining flood magnitude, the high magnitude precipitation events are the key factor in providing high discharges and re-mobilisation of sediment that has been previously stored. Catchment morphometry provides an equally large role in determining sediment delivery out of catchments.

Chapter 11

General Conclusions and Wider Implications

11.1 The alpine context

Projections of future climate change indicate that over the next 100 years global temperatures will increase by an average of at least 1.5°C with higher estimates ranging between 3°C and 5.8°C. Winter precipitation in the European Alps is likely to increase by 20% and there is likely to be increased frequency of extreme high magnitude precipitation events (IPCC, 2007). The impacts of the warming projections alone could alter the annual snow hydrology enormously leading to earlier onset of snowmelt, faster rates of snowmelt and increased frequency of high magnitude flood events throughout the year, particularly in autumn and winter months. These changes may be further compounded by a projected increase of 20% more winter rainfall. The impact of additional rainfall and less snow storage could lead to sizeable changes in annual hydrology, fluvial erosion rates, hillslope erosion rates and shifts in sediment regimes. Ultimately there is likely to be an increase in the frequency and magnitude of flooding, downstream aggradation of sediment leading to increased flooding and a higher incidence of slope failure due to more saturated ground.

Furthermore, human impact on landscapes has been well documented throughout the Holocene, particularly the impacts from deforestation and changing land use. Currently, there is a large degree of abandoned agricultural land in the European Alps, coupled with some management strategies that aim to keep the landscape open. Reducing levels of forest cover or changing pasture to agricultural land can have massive implications for catchments through changes in sediment supply and increased flood magnitudes.

Understandably, with so many potential impacts arising from two system drivers, tools with which to explore the controls and importantly the *interactions* that climate and land use exert at a local level would be useful to identify and potentially quantify the impacts. However, previous attempts to disentangle the relative impacts of climate and land use change through various methods have been hampered by the integrated nature of the signal in sedimentary archives, or fragmentary fluvial archives (e.g. Harvey et al., 1981; Macklin, 1999; Foster et al., 2003; Chiverrell et al., 2007). The system response is often non-linear and can result in disproportionate outcomes from seemingly small-scale interactions. These outcomes cannot be quantified by studying component parts of systems; in order to fully understand these systems a holistic approach (Holling, 2001) must be taken that explores the interactions and aims to identify key behaviours in systems that are a result of the combination of climate and land use changes.

11.2 Model development

Here, a numerical modelling framework has enabled climate and land use drivers to be isolated in order to identify how much of a role each of the drivers exerts on system behaviour and what the impacts of their combined interactions are. The CAESAR model (Coulthard, 1999; Coulthard et al., 2005, Van de Wiel, 2007) has been tested, developed and applied to the Petit lac d'Annecy catchment. The model developments have included the ability to imitate a snow melt and snow storage facility through the precipitation time series that has been moderated against temperature. Another fundamental development has been the improvement of hillslope process representation which has given more realistic erosion and deposition rates, more realistic modelling of landforms such as alluvial fans, and more realistic rates of sediment delivery.

Application of the model over different time scales, in different morphological settings and under different hypothetical scenarios has provided a much better understanding of system dynamics to extrinsic drivers (climate and land use). By approaching this research in a modelling framework, some degree of quantification can be made in when assessing specific changes such as the impacts of a 20% reduction in forest on sediment delivery to the lake, or the changes of sediment composition under hypothetical 'snow-free' conditions.

Each of the results chapters presented in this thesis have been discussed and conclusions have been drawn within the chapter, but the overriding themes that have wider implications for the future are discussed here.

11.3 Model performance

The principal aim of this thesis was to test, develop and apply CAESAR with the anticipation of using it as a robust tool for identifying local-scale impacts of future climate and land use projections. Data-model comparisons show an excellent match between short-term modelled water discharge (which includes a snow melt/store function) and instrumental data. Additionally, with the developed hillslope processes, CAESAR can produce short-term rates of erosion and deposition that are comparable to empirical evidence and field data for the catchment. Whilst these are useful, the focus of this thesis was to extend the length of data-model comparisons by using palaeorecords from lacustrine environments to make long term (10^2-10^4) comparisons to modelled sediment data. The overall long term modelled sediment flux, calculated sediment accumulation rate from modelled sediment discharge, temporal patterns of modelled sediment discharge and behaviours compare well to the proxy records of detrital sediment (for example, χpara) and lake sediment accumulation rates (Jones et al., in prep.). Errors inherent from chronological control render any high resolutions comparisons of sediment discharge invalid. Comparing emergent outputs from the model from the iterative calculation of values in cells based on simple process laws with real emergent forms in the landscape can be deemed a qualitative method of verification (Fonstad, 2006). Emergent properties in this research shows alluvial fan development, build up of fine grained materials followed by a swift reduction after a major increase in sediment flux and intrinsic storage and release cycles confirming that the model is capturing complex system behaviour. The temporal and spatial responses from CAESAR are realistic and compare well to multiple spatial and temporal records. This suggests that CAESAR is a robust tool for use with future projections of climate and land use to explore overall system behaviours and likely trajectories.

11.4 Catchment controls

The results presented within this thesis demonstrated that catchment morphology has arguably more of an impact on the non-linear system dynamics than changes in climate and land use. Whilst the drivers of the system play an important role in governing frequency and the magnitudes of sediment flux, the intrinsic factors within the catchment, namely storage-release cycles, supply-capacity relationships and hillslope-channel coupling also play a major role in governing timing, behaviour and partially the magnitude of sediment response. Whilst total sediment yield over 2000 years for each sub-catchment (chapter 10) were within an order of magnitude of each other, the decadal-centennial scale responses were largely dependent on the balance of the intrinsic factors. These suggestions are corroborated by Coulthard et al., (2007) who suggest that catchment morphology holds the greatest role on levels of non-linear behaviour. Reductions in land use can affect the balance of the sediment conveyor, resulting in lagged responses. Importantly, this research concludes that sediment legacy within a channel over decadal to centennial time scales plays a large role in determining the future sediment response. Dearing (2006b:8) states that "*a modern system is not separated easily from its past: we should expect that it has been conditioned or sensitised by past events or at least bears the legacy of past forcings and responses*". The results presented here demonstrate this idea well, in different sub-catchments under different land use and climate conditions. This is particularly important to take into consideration if future modelling is to suggest likely trajectories of sediment behaviour. Ideally, model simulations should start as far back into the past as possible to allow the system to be conditioned by the past sediment. Some of the sub-catchments are more sensitive than others, the Montmin sub-catchment was particularly sensitive to changes in both land use and climate. This style of catchment with high erosion rates due to steepness and length of slopes, well coupled gully-channel systems and little opportunity for storage in the channel is ideal for studying future impacts as the high magnitude changes are likely to be registered in these zones first.

Quantification of the impact of changes in climate and land use has been a key question through this research. In order to make specific quantifications such as "how much more sediment will be delivered to the lake under 20% increase in forest cover?", multiple scenario approaches work best as other factors can be held constant, therefore any quantifiable differences must be a function of a single parameter (e.g. forest). Trying to quantify differences in data through a single simulation is very difficult as many processes are variable (e.g. for individual sub-catchments in chapter 10). For example, comparing the changes in sediment discharge from a 25mm precipitation event under a cleared landscape with a 25mm precipitation event 20 years later under a 'dense' landscape; one is essentially looking at the differences between more than one variable (i.e. not just forest cover), as sediment legacy in the channel is important to determine future response and that alone may alter the magnitude of the sediment peak between the two precipitation events. General qualitative suggestions can be made, and results from chapter 10 suggest that a ~20% reduction in forest cover around AD 900 increased the amount of sediment delivery by at least one order of magnitude. This level of increased sediment supply could have massive implications for the capacity of channels to carry sediments out of the system, therefore more storage may occur resulting in more frequent high magnitude floods. Coupling these conditions with high magnitude precipitation events, would further increase the amount of sediment discharge.

11.5 Snow cover

The hypothetical snow/snow-free simulations presented here conclude that warmer temperatures will lead to changes in the annual hydrological cycle, with more frequent high magnitude flood events through the year rather than a long snow melt period. Under snow-free conditions, there would be around 1.3-1.4 times more sediment over the 2000 year period and both sub-catchments suggest a similar overall change. In terms of how this additional sediment would impact the lake, a catchment wide estimate suggests that there would have been an additional 1m of sediment accumulation in the lake under snow-free conditions. Although based on hypothetical past scenarios, the conclusions are valid for broad trajectories of change for the future under snow-free conditions. The results presented here for grain size changes suggest that as temperatures increase in the future, above average annual temperatures will control the rate of snow melt and ultimately coarse grain sediment discharge. The warmer the temperatures get, the faster the snow will melt, and it is more likely that increased coarse grained sediment discharge will occur. If we arrive at a point in the future where there is no snow, then the system will become altogether more coarse-grain dominant, and the results here suggest that under ~30% forest cover, the system can transport the 1000μm grain size class out of the catchment compared with ~ 250μ to 500μ at present (Jones et al., in prep.). Not only is the amount of sediment discharge important to quantify but the composition of that sediment is particularly crucial. The results here suggest that from now until the point of a 'snow-free' pre-Alpine environment, coarse grain sediment discharge will increase as temperatures increase. This has massive implications, particularly for the channel, as coarse grain sediment aggradation can lead to increased likelihood of braiding in streams and alter not only the flood frequency and sediment conveyor but would also impact on the diverse fauna that dwell in gravel-bedded rivers.

The results from the forest cover simulations show that a 20% reduction in forest cover over the whole catchment would increase the sediment accumulation in the lake by approximately 0.5-0.8m in 2000 years. Similarly, there would be an additional 1m of sediment accumulating in the lake if the entire catchment was snow-free. The implications of this amount of additional sediment within the system would be large-scale. There would be more disruption to the sediment conveyor due to a change in the balance between supply and capacity, which ultimately dictates the storage and release cycles. Aggradation of sediment in the channel would increase flooding, particularly in low lying areas. Areas that would be most affected, are downstream zones, predominantly in the flood plain areas. In the Petit lac catchment, the town of Faverges which is situated very close to the river would suffer from increased flooding if there was downstream sediment aggradation. Any level of flooding on the Eau Morte floodplain could damage crops, flood houses and land and fundamentally impact the local economy. The additional amount of sediment into the lake would have implications for aquatic macrophytes, which would ultimately reduce water quality, and cause disruption to benthic-dwelling fauna. In other areas, particularly those with reservoirs, increased sediment accumulation could lead to reduction in water yield capacity which would have implications for the already problematic water resource availability in mountain areas.

11.6 Forest cover

The results attempt to quantify the point at which forest cover is sufficiently low that the landscape is sensitised to changes in climate. The results point to forest cover over ~60% showing little signs of sensitivity to change in precipitation but do result in decreased flood magnitudes which may lead to more channel incision. This suggests there may be a tipping point around 60%. Within the 0-60% cover, there may be a further tipping point somewhere between 20% and 40% as linear reduction between the 'original' and 'dense' forest cover does not result in linear magnitudes of sediment discharges. The possible tipping points are important to note for land use management schemes, particularly given that the present day forest cover is around 55-65% in most of the sub-catchments in the Petit lac d'Annecy. In addition to this, all three simulations in both sub-catchments demonstrate that significant responses to changes in climate and land use occur when forest cover is less than ~55-60% suggesting this may be the tipping point at which slopes and channels may be de-coupled. Below 55-60% cover there is a response to change on the slopes (increased sediment supply due to increased rates) and a response to the change in the channel (higher magnitude flood events and increased transport capacity). Above ~60% there is a reduced supply from the hillslopes, reduced flood magnitudes so sediment cannot access over bank deposits and channels begin to incise.

11.7 Spatial representation

A final conclusion following this research is that the use of a spatially distributed numerical modelling approach cannot be overestimated. Temporal records provide part of the story, but to fully understand the internal system dynamics of storage and release, spatially distributed modelling is key. Many explanations can be hypothesised from time series, but spatial modelling or field study can corroborate these ideas so that a fuller understanding of the response can be made. One of the key examples presented here was the non-linear response of sediment discharge to increasing forest cover. Lane et al. (2008:246) state that "*localised reforestation appears to be capable of reducing delivery rates significantly*" but the results from the model challenge this idea, and suggest that if forest cover is too high, channel capacity is too low to reach any sediment stored in overbank deposits, thus incision of the channel occurs leading to an increase in sediment delivery rates. The benefit of using a spatially distributed model is that it is possible to investigate these hypotheses further and identify from maps of net erosion and deposition if this is actually present in the simulated geomorphology. Ideally, all temporal studies should use some level of spatial verification to identify if the story that the temporal pattern suggesting has actually taken place in the landscape. However, this is extremely difficult to do over long time periods as the processes have often obscured past evidence, leaving only fragmentary records of landforms such as fluvial terraces. Spatially distributed numerical modelling allows the full spatial record to be captured in maps that can be analysed to corroborate our hypothesis in the temporal story.

11.8 Modelling constraints

One of the biggest practical limitations to cellular automata modelling is computational processing capacity. Coulthard et al. (2000) outlined this limitation, and although there has been some improvement in computer processing capacity in the last ~10 years, each model simulation of 2000 years at 50m resolution (catchment size 7km^2) on an Intel Quad core processor (Q6600 @ 2.40Ghz) can take around eight weeks, depending on the size of the catchment. In order to model multiple scenarios, a suite of computers are required to undertake simultaneous simulations. Finer resolution spatial scales (10-25m) are desired to look in more detail at the geomorphology, but at catchment scale, the simulations would be exponential in their simulation time. Coulthard et al., (2007) suggest that CFD models may eventually supercede reduced complexity models, but at present, computer processing power is insufficient to apply CFD models over long time scales, and although the fluvial process representation is high resolution, there are many difficulties with coupling these processes to sediment transport (Kirkby, 1999). The opposite side of this argument is that hopefully computer processing power will enable reduced complexity models to run much faster, giving more scope for greater spatial and temporal exploration.

11.9 Further research

The possibilities for future development and application of CAESAR are endless. In the Petit lac d'Annecy catchment, asimulation of the Eau Morte floodplain is likely to be the next stage of research following this thesis. This will give greater insight into the role that intrinsic factors (e.g. storage/release and supply/capacity) play on sediment dynamics. In addition to this, multiple precipitation scenarios (e.g. IPCC) could be coupled with systematic changes in land use to attempt to quantify future sediment impacts from projected combinations of land use and climate.

In terms of model development, the next major step would be to incorporate more spatially distributed parameters. These developments could include the ability to assign different land uses to different parts of a catchment (e.g. one slope forested, one slope pasture, valley floor agriculture etc.) rather than homogeneously as in the present model. This would provide a more useful tool in future impact assessment studies and give a better idea of possible trajectories of change under more complex land use practices. One of the most important spatially distributed additions would be that of snow depth and accumulation as an additional grid within the model which would allow temperature lapse rates to vary with altitude and snow melt/store to vary across the catchment.

Further validation, particularly spatial validation could be sought using finer resolution digital elevation models to simulate more detailed catchment geomorphology. In the Petit lac d'Annecy there is the potential to drill palaeochannels across the Tamie fan using a Vibrocorer. They could then be dated and the Tamie fan could be modelled at high resolution to investigate whether the model could simulate incision the channels at the same rate as the rates in the landscape.

In terms of application, the advantage of CAESAR is that it is a generic model which can essentially be applied in many different environments. Recently fluvial geomorphologists have discussed the possibility of applying CAESAR in semi-arid environments such as South Africa or the Mediterranean region. Although some small developments would be necessary, the application of CAESAR under radically different environmental conditions could be useful for the exploration of impacts of system drivers in these regions. In addition to this, it is increasingly important to bring a human agent into landscape dynamics and immediate future research is likely to focus on working with agent-based modellers to provide some hypothetical land use scenarios for the future in different socio-economic environments to identify the scale of likely impacts in different socio-economic environments.

In practical terms of future application of the methods presented here, lake sediments do provide excellent continuous records of environmental change and are an extremely useful record with which to validate the model against. The best sites to use for future research are likely to be small catchments with rapid sediment accumulation rate (e.g. Brotherswater, UK (Schillereff (2009), unpublished Master's thesis) or those with limited biogenic activity to provide a palaeorecord that provides the best proxy for detrital sediment supply as possible. If this is not possible, then perhaps pre-lake deltas may potentially yield more information about detrital sediment supply than lakes. Alternatively, the correction approach used within this thesis (chapter 10) is the best approximation that can eliminate any biogenic activity from the lake sediment signal. The potential for application of CAESAR is unlimited, and studies over the last ten years have only begun to scratch the surface of its capabilities for future research.

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