Global Continental Changes: the Context of Palaeohydrology

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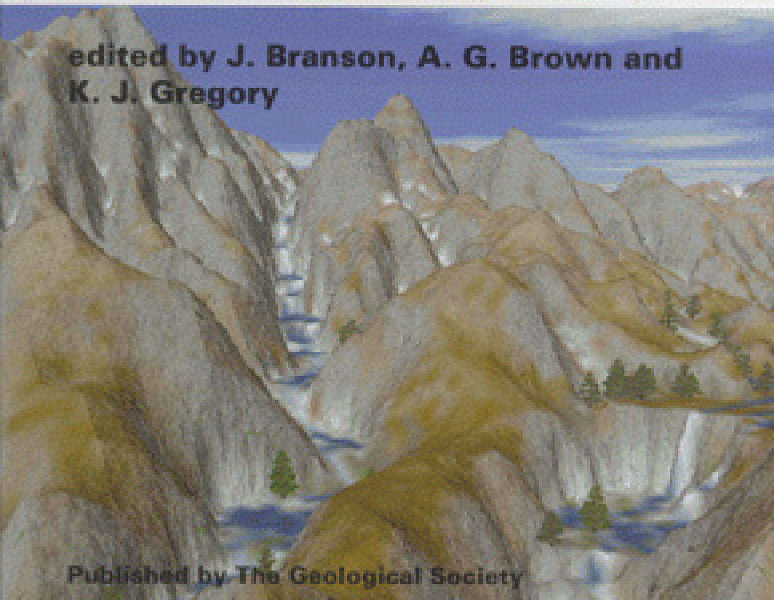

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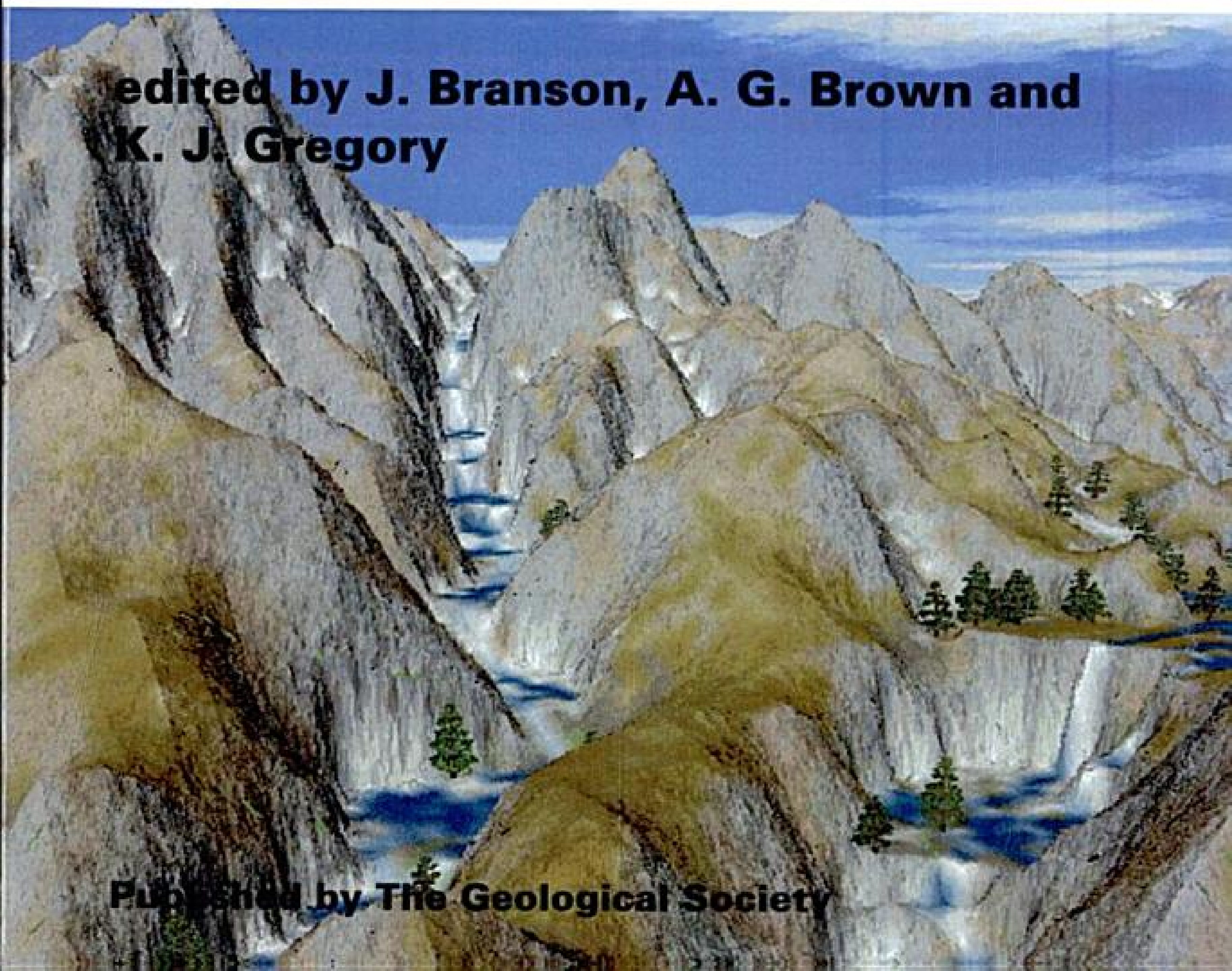

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Global Continental Changes: the Context of Palaeohydrology

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Contents

A FUTURE FOR PALAEOHYDROLOGY

Preface

This volume arises from two international meetings held at Chilworth Manor, Southampton, 9–12 September 1994 and at the Geological Society of London, 13 September 1994. The first was organized by the INQUA Commission on Global Continental Palaeohydrology (GLOCOPH) and the second by the Commission in conjunction with the British Geomorphological Research Group. The meetings each brought together researchers from a wide variety of disciplines to discuss research in global palaeohydrological change over the last 20 000 years. They were the first to be held as an integral part of the INQUA Commission.

The papers in this volume are organized into three sections. The first provides an overview of the context in which palaeohydrology is studied and researched, the second reflects the approaches to palaeohydrological analysis in a number of contrasting geographical regions and the third discusses a possible future for palaeohydrology.

We are very grateful to the many people who helped to organize and finance the meetings and to prepare the manuscripts, and we particularly acknowledge the assistance of those scientists who refereed the papers in this volume, and of Caroline Ensor at Goldsmiths' College. Palaeohydrology is now recognized as an important multidisciplinary area of research and we hope that this volume reflects the present achievements and provides the basis for more significant advances.

Julia Branson
Tony Brown
Ken Gregory
February 1996

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A FUTURE FOR PALAEOHYDROLOGY

From Branson, J., Brown, A. G. & Gregory, K. J. (eds), 1996, *Global Continental Changes: the Context of Palaeohydrology*, Geological Society Special Publication No. 115, pp. 1–8

Introduction

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Four aspects are outlined in this introduction to provide a background context for the volume. First, consideration of *developments* in palaeohydrology provides an opportunity to summarize the way in which the discipline has evolved. This is succeeded by a synopsis of seven *prevailing themes* which have been evident in the development of palaeohydrology and which have produced a number of *outstanding questions*. Finally the *current approach to global continental palaeohydrology* is explained.

Developments

The first international meeting of GLOCOPH (the INQUA Commission on Global Continental Palaeohydrology) held in Southampton in September 1994 occurred 50 years after the first explicit definition of palaeohydrology was made by Leopold & Miller in 1954. Over that 50 years, interest in palaeohydrology was at first somewhat slow to develop, but Schumm in 1965 produced a major paper on Quaternary palaeohydrology in which he suggested that palaeohydrology offered an innovative approach which was capable of further exploration. The fundamental proposal of Schumm's paper (1965) was that global relationships between run-off and precipitation and also between run-off and sediment yield could be employed to indicate, for different temperature conditions, how changes might occur under different climatic conditions. This was a major breakthrough because it gave a mechanism for indicating the types of change that were possible prior to the availability of estimates of hydrological change obtained from sophisticated computer models. Further developments that occurred in the 1960s and 1970s were eventually summarized in the book by S. A. Schumm (1977) entitled *The Fluvial System*. That book, in addition to summarizing an innovative approach to the Quaternary, particularly through palaeohydrology, also explained an approach to river metamorphosis which previously had been developed in a number of papers (e.g. Schumm 1969) founded upon the relationship first enumerated by the hydraulic engineer, E. W. Lane (1955). A basis of analysis in quantitative stream morphology was proposed by Lane (1955) using the approximations $Q_s d \approx Q_w S$ where Q_s is the quantity of sediment, d the particle size or size of sediment, Q_w the water discharge, and S the slope of the stream. The approximation was used by Lane to demonstrate six classes of change which were used to interpret the effects of engineering works upon river channels. Gregory (1983) indicated how the foundation provided by Lane (1955) was developed in a series of important papers by S. A. Schumm culminating in his book in 1977 (Schumm 1977).

Apart from these antecedents, a major development in palaeohydrology occurred through the IGCP (International Geological Correlation Programme) Project 158 which was organized from 1978 to 1989. That project followed a programme of research organized as two sub-projects. Sub-project A, concerned with fluvial systems, was headed by Leszek Starkel and Sub-project B on bogs and mires, was led by Bjorn Berglund. The research focused upon areas in the temperate zone and was very effective in three major ways. First, it co-ordinated the results from individual palaeohydrology research investigations that were already taking place in a number of separate countries. Secondly, through the organization of regular annual meetings, often with field excursions, it was possible to effectively compare results, techniques and data to facilitate the better understanding of palaeohydrology of the temperate zone. The third successful outcome was that the project produced publications which not only advanced the techniques appropriate for palaeohydrology but also provided reports for individual countries (e.g. Starkel 1990; Gregory *et al.* 1987) and gave correlated results from the overall project. The final volume, entitled *Temperate Palaeohydrology*, was published in 1991 (Starkel *et al.* 1991) and brought together results from 14 countries in the temperate zone.

Prevailing themes

Palaeohydrology may be defined as the science of the waters of the earth, their composition, distribution and movement on ancient landscapes from the beginning of the first rainfall to the beginning of continuous hydrological records (Schumm 1977; Gregory 1983). As research has progressed, it has become apparent that a number of separate themes has been pursued in palaeohydrology and that particular schools of approaches have been fostered. Perhaps as many as seven approaches might be identified from the research from 1965 to 1991, although inevitably there is overlap between each of the themes.

First, there have been studies of cut and fill sequences and of terrace and valley floor development. Analyses of valley in-fills have allowed the interpretation of the palaeohydrologic conditions under which different stages of valley floor development occurred. This was the *raison d'être* for the study by Leopold & Miller (1954) that led to the first formal definition of palaeohydrology. More recently other models of flood-plain and valley floor development have been developed such as the alternation of episodes of vertical accretion and catastrophic stripping proposed by Nanson (1986). Whereas that general approach was emphasised initially in semi-arid areas such as the south west of the United States, a second approach was essentially based upon the water balance (Schumm 1965) and was capable of application to a range of environments and particularly to temperate areas that had experienced periglacial conditions during the Quaternary. This approach embraced modelling techniques as linked to climatic change (e.g. Lockwood 1983) although the inclusion of palaeohydrology as an integral part of global change programmes has not been readily accomplished (Gregory 1995). A third approach was initiated by Dury (1964*a*, *b*, 1965) when he analysed former large meanders based upon valley meanders, concluded that extensive stream shrinkage had occurred, and deduced that one of the major reasons for such shrinkage was change of climate. Studies of such underfit streams were conducted by Dury and later developed by other researchers in the context of particular areas (e.g. Rotnicki 1983, 1991; Maizels & Aitken 1991). These investigations led to the conclusions that, not only had river discharges been significantly greater in the past, but also morphometric analysis of contemporary meanders and comparison with valley meanders could facilitate estimation of the palaeodischarges that had occurred.

Allied to this approach were studies based upon palaeohydraulic methods which utilized analyses of sediment characteristics to furnish information about palaeodischarges and palaeohydrologic conditions (Gregory & Maizels 1991; Williams 1983). The diversity of approaches and the equations available were stressed by Williams (1984, 1988). An approach to river metamorphosis can be identified as a fourth theme and was developed in a series of research investigations (e.g. Schumm 1969, 1977) which served to demonstrate the extent to which, in the Holocene, there had been a significant number of major changes of river channel patterns instigated by climatic changes and by human activity, with the influence of the two causes often interwoven in a complex way. This complexity has been demonstrated in studies of the upper Mississippi (Knox 1984), in Europe in general and in Poland in particular (Starkel 1991). Results obtained from the investigations of bogs and mires, particularly using palaeoecological techniques, provide a fifth strand of palaeohydrologic research and not only was this a theme in IGCP Project 158, Sub-project B but it has been the basis for a considerable number of research investigations (Berglund 1979) which have led to conclusions about the water balance under previous climatic conditions which in turn have been the basis for estimating the characteristics of the palaeohydrology (Berglund 1977, 1983; Lang & Schluchter 1988). In some areas, lake fluctuations have provided a significant indicator of hydrological changes and so lake databases (e.g. Harrison 1988; Street-Perrott *et al.* 1985) have been employed in a sixth approach to give information about the characteristics of palaeohydrological environments.

These six approaches have been used to varying degrees in studies of particular basins in the temperate zone to enable reconstruction of palaeohydrological changes during the last 15 000 years as part of IGCP Project 158. Thus has arisen a seventh approach which has endeavoured to integrate individual approaches and which includes impressive studies undertaken in the Vistula basin (Starkel 1995) and in other basins investigated as part of the project and described in *Temperate Palaeohydrology* (Starkel *et al.* 1991).

The seven types of approach were evident by the early 1980s and as research has progressed it has been aided by the application of new techniques including palaeomagnetism (Oldfield 1983), the use of trace minerals and mining debris (Lewin & Macklin 1986), and also by the investigation of palaeostage indicators (PSI) which have been employed in a variety of areas to give information on former catastrophic floods. This work was initially set in the context of flood hydrology (Baker *et al.* 1988) and a series of impressive research investigations has been achieved utilizing evidence from slack water deposits and from their situation and sequence (Baker 1987). In addition, other developments have occurred especially in relation to groundwater changes, in particular the effects of glaciers on groundwater (Boulton & Spring 1986; Boulton & Dobbie 1993; Boulton *et al.* 1993) and changes in arid regions (Issar *et al.* 1984; Issar 1985).

Outstanding questions

The general achievement of research over the last 50 years has enabled the improved understanding of palaeohydrological environments in the light of knowledge of contemporary hydrological processes. Thus, as the understanding of hydrology and of fluvial processes has advanced since the 1960s, particularly with the development of the variable source area concept and run-off modelling (e.g. Gregory & Walling 1973), the investigation of palaeoenvironments has now benefited enormously from an enhanced understanding of contemporary environmental processes. This awareness of global environmental process is reflected in recent summary texts (e.g. Mannion 1991; Roberts 1994) although the potential contribution of palaeohydrological research in the study of global environments is not always as clearly acknowledged as it might be (Gregory 1995). Although global change research programmes have developed significantly over the decade since 1980, and many now acknowledge the contribution that palaeohydrology can make, nevertheless palaeohydrology has yet to attract the level of funding that has been associated with general circulation models, with atmospheric modelling or with analyses of major oceanographic systems. This is perhaps slightly paradoxical because the impact of global change is often reflected through the river system so that the knowledge of the magnitude of palaeohydrological river system changes could have significant implications for understanding river flows in relation to human activity and hazards.

However, not all areas of the world have been studied in equal detail using palaeohydrological methods of any kind. Research investigations included in a database (Branson *et al.* 1995), which when analysed according to world distribution (Fig. 1), show that there has been a preponderance of research in European temperate areas and in some parts of North America. In just the same way that Graf (1984) related the location of geomorphological research investigations to the

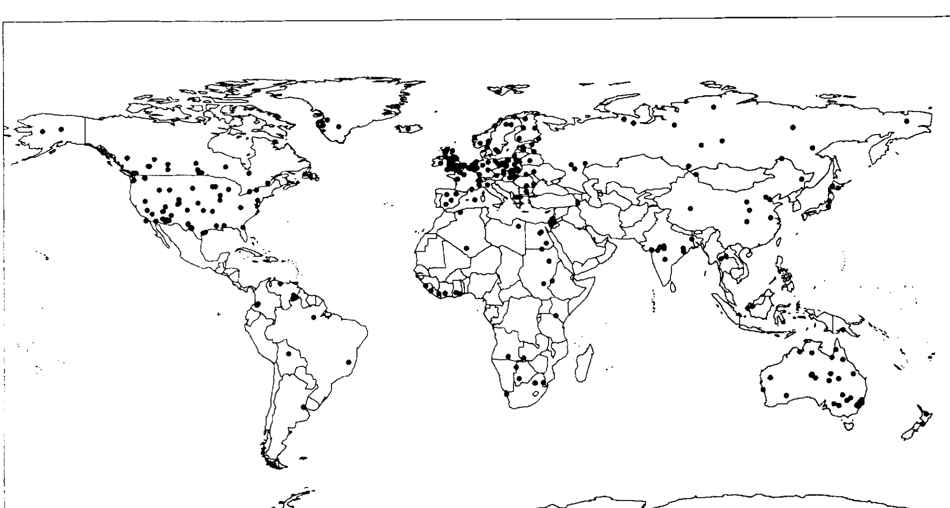

Fig. 1. Location of areas which are the subject of published research (English language) on palaeohydrology to 1994. The distribution was produced from a database compiled by Julia Branson (see pp. 235) funded by a Leverhulme Research Grant to K. J. Gregory.

different types of environment in North America, so it is possible to see how the themes that have prevailed in the development of palaeohydrology research have been associated with particular areas of the world. Despite the concentration of research in western and central Europe and to a lesser degree in the United States (Fig. 1), it is encouraging to see how some investigations have been undertaken in other continents but there is an outstanding need for further studies to increase the understanding of palaeohydrological change; the chapters in this volume and research work as part of GLOCOPH should fill some of the gaps.

A further outstanding requirement is to place the results of palaeohydrological investigations in a functional drainage basin framework. Contemporary hydrological models have to be developed for drainage basins, and to further advance modelling in palaeohydrology it is desirable to analyse the research results within the drainage basin framework. Although this was originally advocated by Schumm (1977) it remains to be fully developed; research of this kind is a theme which needs to be kept in mind with the further development of palaeohydrological research. Allied to progress in this direction will be the application of significant recent results in interpreting the geomorphic effectiveness of different kinds of floods (e.g. Costa & O'Connor 1995) and the spatial patterns of erosion and deposition (e.g. Miller 1995) to palaeohydrological sequences as anticipated by Thornes & Gregory (1991) and recognized by Petts (1995) and Gardiner (1995). This approach may enable us to add further to present understanding of changes of river channels during the Holocene (Starkel 1995) and provide what Baker (1991) called a bright future for old flows.

An approach to global continental palaeohydrology

Because of the variations in the intensity of palaeohydrological research (Fig. 1) and also the need to correlate results from one area to another, it was desirable to establish a collaborative research programme that synthesized results, from a variety of world areas, obtained by utilizing a range of techniques. At INQUA, in Beijing in 1991, a new commission was established concerned with global continental palaeohydrology (GLOCOPH) with Leszek Starkel as President and V. R. Baker and K. J. Gregory as Vice Presidents. It was intended that this new commission would build upon the work already achieved by the IGCP project, facilitate the investigation of palaeohydrology on a global basis and place the results in a context related to global water balance changes (e.g. Waylen 1995). Whereas the IGCP project had raised the profile of palaeohydrology and increased the awareness of the contribution that this interdisciplinary field could make, it was also timely to relate studies of contemporary processes to past environments and to use the increased understanding gained for the elucidation of future environments. The Global Continental Palaeohydrology Commission of INQUA was established in 1991 potentially using the involvement of some 90 researchers in more than 30 countries. The first international meeting of the Commission held in Southampton in 1994 involved the presentation of a series of important papers and this was followed by a day of lectures at the Geological Society of London when several invited contributions demonstrated the context for palaeohydrology. It is from those papers presented in London on 13 September and in Southampton between 10 and 12 September 1994 that this volume has been compiled. As palaeohydrology can seek to improve our understanding of past environments using a knowledge of contemporary processes, the research undertaken in this way is able to complement knowledge gained from the period of instrumented records and it may in turn provide insight into environments that could be analogues for future situations.

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Palaeohydrological reconstruction: advantages and disadvantages

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Abstract: The study of continental palaeohydrology is a young branch of science which has drawn on knowledge of the mechanisms, and observations, of present-day processes to reconstruct past hydrological conditions. Patterns of annual precipitation and evaporation, for example, can be determined from the distribution of plant communities or by dendrochronological techniques, lake-level variations can provide an indication of water storage fluctuations, reconstructions of runoff, particularly bankfull discharge and mean annual discharge can be based on palaeochannel form and sedimentology, and extreme floods can be studied using slack-water deposits. All of these techniques rely on the acceptance of assumptions which may not always be valid. It is therefore important that several different methods are used to provide a collaborative approach both to validating methods and in determining the driving force of hydrological changes.

In many palaeogeographic and palaeoclimatic reconstructions we touch upon the components of the hydrological cycle, mainly precipitation and an index of aridity. The water balance equation is the basis of these reconstructions, the elements of which may be retrodicted on the basis of geological evidence (stratigraphy, forms, fossil flora and fauna) as well as with models (Fig. 1).

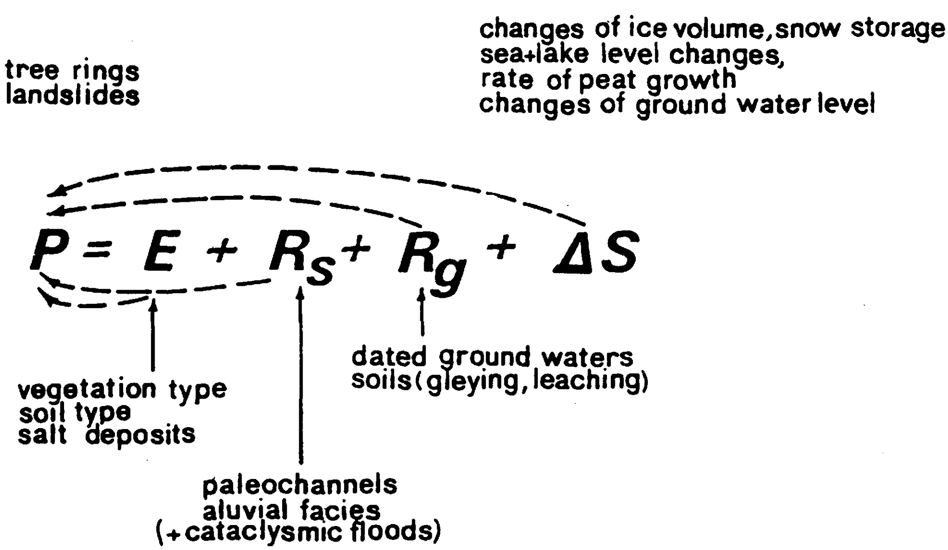

Fig. 1. Equation of the water balance. Main sources of information on the paleohydrological parameters.

At the outset the validity of the information of the water balance elements used to reconstruct past water circulation should be questioned (Starkel 1993). The reconstructions make the picture of changes more rational. The amount of water stored in the oceans and lakes are the most appropriate factors for quantification. However, it should be recognized that calculations of the present water balance differ significantly (e.g. Keller 1962; Lvovitch 1974; Shiklomanov 1980), and errors in the calculation of total water resources, runoff and atmospheric precipitation may be very large indeed.

Sources of information about hydrological parameters

If the water requirements of plant species are known, then on the basis of the uniformitarian approach it is possible to reconstruct the spatial and temporal patterns of precipitation and evapotranspiration (Webb & Bryson 1972; Grichuk *et al.* 1984; Klimanov 1990), and indirectly runoff (Georgiadi 1992) (Figs 2 & 3).

Fig. 2. Annual mean runoff during the Last Interglacial climatic optimum (deviations from present day values) (after Georgiadi 1992).

A problem with these reconstructions, however, is that the zonal concept is often over-simplified; ecotonal zones have shifted and varied in time, and the statistical approximations may be insufficient to accurately describe the ecotonal response to water deficit or permafrost conditions (Webb 1983). Under these environmental conditions the ecotonal zones may be extremely wide and complicated (cf. Frenzel *et al.* 1992), and the locations where organic material has deposited are rare and site-specific and therefore not representative of large areas. Thus, in the case of permafrost regions, the calculations of annual precipitation totals can be over-estimated.

Variations in lake levels have provided the best indicators of long-term trends in mean annual rainfall and water storage. Changes in lake level are reflected in coastal forms, facies differentiation and faunal spectra of diatoms and cladocera (e.g. Street-Perrott & Harrison 1985; Digerfeldt 1986). The most widely accepted reconstructions have been obtained for closed depressions of tectonic or similar origin (e.g. the Caspian Sea; Klige & Myagkov 1992).

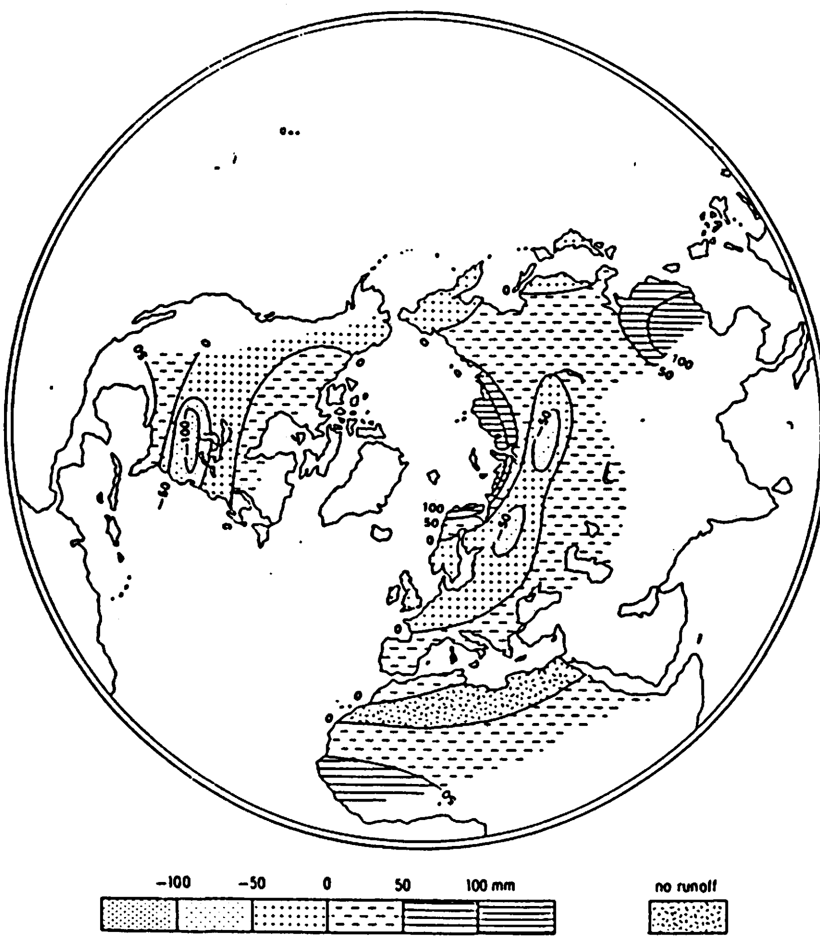

Fig. 3. Annual mean runoff during the Holocene climatic optimum (deviations from present day values) (after Georgiadi 1992).

Fluctuations of lake level and water input have been reconstructed for the North Crimean Lake Saki based on the thickness and lithology of the annual laminae. These fluctuations have been correlated with the discharge of the Dnieper River for the last 2500 years (Schwetz 1978). In the peat cores breaks in peat stratigraphy indicate phases of desiccation or drought, e.g. of distinct lowerings of the groundwater level (Casparie 1972; Aaby 1976). Many of these droughts occurred only locally, however, and do not provide a regional picture of hydrological change.

In deglaciated areas or thermokarstic depressions, the water level may be controlled by long-term trends in lake drainage or overtopping and thus the influence of climatic factors may only be indirect. Calculations of the percentage of randomly investigated lakes with higher or lower water levels in particular time slices may thus lead to incorrect conclusions (Harrison *et al.* 1993).

Reconstructions of runoff, particularly of bankfull discharge and mean annual discharge, are based on parameters derived from palaeochannel shape and sediments. Unfortunately, however, identification of the majority of these parameters is problematic (Gregory & Maizels 1991; Soja 1994). There are many equations that can be used to calculate bankfull discharges and mean annual discharges from measurements of meandering channels (Dury 1977; Rotnicki 1991). Most of these equations are based on the assumptions presented below.

In the case of palaeomeanders the existence of a single thread channel is assumed to be the normal equilibrium planform. It is difficult to determine the bankfull discharge or the channel depth and the base of the coarse (armoured) horizon from stratigraphical evidence: there may be several such layers and the upper one may have been formed after the channel was cut off. Mature, tortuous free meanders were probably created during long, relatively stable periods (perhaps lasting centuries or millennia), but conversely, the reconstructed discharge is related to the moment just preceding the beginning of filling dated by radiocarbon techniques.

In the case of braided palaeochannels the errors may be even greater that those involving single-thread channels, and may reach several hundred percent (Maizels 1983), as it is difficult to determine which branches of the braided system were active at any one point. Moreover, the grain size composition of the channel facies may provide ambiguous information on the flood discharges (Church 1978).

Extreme floods are studied by the analysis of coarse channel debris and slackwater deposits (e.g. Baker 1987). Using these techniques the occurrence of cataclysmic floods from a variety of origins has been identified on different continents (Baker *et al.* 1988; Rudoy & Baker 1993). When reconstructing the frequency of extreme events on this basis, however, the preliminary assumptions should be recalled. In one sequence of slackwater deposits several floods may be identified but each subsequent event must exceed the previous one in order to be registered in the stratigraphic sequence. Even in the case of the famous Missoula flood it is not known whether it was a single or several events, close or distant in time (Baker & Bunker 1985). Additionally, it may have been assumed that the channel floor was cut into the bedrock, and was therefore stable. Great floods may have deposited several metres of debris on the rocky bottom which were removed by a later flood event (cf. the Tista river in the Sikkim Himalayas; Froelich & Starkel 1987). New methods of dating historical alluvia by the use of heavy metals and radioactive elements (Zn, Pb, Cs etc.) helps to identify single floods and evaluate the more precise time intervals between them (Knox 1983; Macklin & Dowsett 1989).

In the temperate zone a clustering of contemporary black oaks is considered as an indicator of frequent flooding (Becker 1982). The use of such indicators requires caution since these trunks, as well as the sediments accompanying them, may be redeposited depending upon the local conditions (Kalicki & Krapiec 1995).

Extreme rainfalls (heavy downpours, continuous rains and rainy seasons; cf. Starkel 1976) are often accompanied by mass movements (e.g. debris flows and earthflows, shallow- and deep-seated landslides). Their distribution in time may be essentially random and therefore correlation with sequences of other events is needed (Starkel 1985, in press). Most mass movements are dated using organic deposits which fill the depressions over the sliding masses; such dating methods may underestimate the age of the landslide, however, due to the lag in peat formation or lacustrine deposition. The most reliable datings originate from sediments overridden by sliding masses or from reservoirs dammed by landslides (Alexandrowicz in press).

In the cold regions changes in the radiation/evaporation balance and snowfall are reflected by fluctuations of glaciers. A careful examination of the advance and retreat of Alpine glaciers indicates that parallel tendencies can be observed. An individual glacier, however, reacts to variations in temperature and precipitation with different intensities and response times (Patzelt 1985). Therefore fluctuations of the ice fronts can only adequately inform about the centennial changes.

Datings of fossil reservoirs, particularly in sedimentary rocks in the arid zone can provide information regarding periods of high groundwater storage (Geyh 1972; Sonntag *et al.* 1981). Again, interpretation must be cautious, since the reservoirs may be of a relict character or subject to manifold or continuous mixing.

Reconstruction of the water budget

The methods described above for the reconstruction of components of the water cycle or water budget serve as tests of retrodictions of the whole water balance. Closed lacustrine catchments are the most suitable environment to undertake this type of analysis. It is possible to compare geological records with the retrodiction based on modelling (Kutzbach 1980, 1992; Swain *et al.* 1983; Klige & Myagkov 1992). More difficult and uncertain are the tests of balance calculations based only on the water requirements of vegetation (Georgiadi 1992) or on the runoff characteristics reconstructed from the channel parameters (Rotnicki 1991). Different seasonal distributions of rainfall and type of flooding are connected with snow storage and existence of permafrost, and if these factors are not taken into consideration for the Late Glacial period the retrodicted precipitation and evaporation may be overestimated (Klimanov 1990; Rotnicki 1991).

The changes reconstructed from palaeohydrological analysis may be induced by climatic change and/or by modification of the water cycle by human activity. The relationship between flood frequency and rise of groundwater level with forest clearance, for example, are generally well known. The evidence gathered from valley floors in Poland (Starkel 1991*b*; Kalicki 1991) and in Britain (Needham & Macklin 1992), for example, indicate that the youngest observed phases of aggradation or higher flood frequency were caused by the coincidence of the climatic phases and human intervention (e.g. the late Roman phase, termination of the Medieval period).

In undertaking palaeohydrological reconstructions a special role is played by the interpretation of radiocarbon datings. The dated organic samples may either be assigned to the sediments *in situ* or to the re-deposited sediment. The histograms of frequency of the organic samples, used as the indicators of a higher water level or higher precipitation, may lead to incorrect conclusions. Especially doubtful are the reconstructions for the periods of time which correspond to rapid changes in vegetation and climate ^{14}C method is not sufficiently precise. A deviation of 100 or 200 years in the time intervals of 6000 or 5000 BP may result in the correlation of different phases or events. In such cases the different response times of different geoecosystems to long-term trends or extreme events should be considered.

Discussion and conclusions

When palaeogeographic reconstructions, including palaeohydrological ones, are made for specific time intervals it is assumed that the geoecosystems are in equilibrium with the radiation balance. This assumption, however, is erroneous at least for the beginning of the Holocene and for the Atlantic–Subboreal transition. There are also other causes of diachronity, for example, the stable ecosystems and fluvial systems of lowland areas demand much higher threshold values to exceed equilibrium conditions than more unstable mountain geoecosystems (Starkel 1991*a*).

The critical evaluation of the evidence and methods presented above does not imply that palaeohydrological reconstructions are unachievable. On the contrary, there is a strong need for the careful comparison of hydrological changes registered in sequences of sediments, in order to explain the causes and mechanisms of change. Several recent approaches (e.g. Harrison *et al.* 1993) have shown that the results obtained by different methods are difficult to correlate. This implies that either the methods used are incorrect or incomparable or the derived conclusions may be wrong (due to equivocal assumptions).

Two kinds of records of hydrological change in the past should be distinguished. First, continuous records of change in relatively stable environments such as laminated lacustrine sediments, cave calcareous precipitation, peat bogs, soil profiles, tree rings or ice cores. All of these register undisturbed long-term trends of changes in the water budget. Second, records which register extreme changes caused by fluctuations in precipitation and runoff such as catastrophic rainfall, flooding and drought. To retrodict a hydrological regime it is necessary to infer from both kinds of records (Klige 1990) which may simultaneously register different causes of change. In the temperate zone of Eurasia there are, for example, long-term changes in water storage which reflect changes in the air mass circulation pattern. This is observed in the N–S transect across Europe, which shows diachronous changes of lake levels and glacial advances during the Holocene (Karlen 1991; Harrison & Digerfeldt 1993). On the other hand, phases with a high frequency of extreme events have much wider coverage and are evidenced in tree ring patterns, flood deposits etc. connected with random rhythmicity of different global or hemispheric origins (e.g. phases of high volcanic activity; Nesje & Johannessen 1992).

The correlation between phases of different humidity and water storage with periods of extreme hydrological events indicate the existence of regions where both kinds of rhythmicity are superimposed or in phase with each other, e.g. wet phases coincide with phases of higher flood frequency. During the Holocene such coincidence occurred in Central Europe. In other parts of the globe these two types of rhythmicity were probably out-of-phase.

In many cases it is difficult to separate the long-term changes in the water budget from single extreme events. In closed lake basins a similar effect in the form of a rise in a lake level in areas of water deficit may be caused either by a secular rising trend of precipitation and runoff (or decline of evaporation) or by rapid catastrophic rain (cf. heavy rain at Sambhar Lake in Rajasthan; Starkel 1972).

The study of continental palaeohydrology is a young branch of science which has drawn on knowledge of the mechanisms and observations of present-day processes. The methods of reconstruction require further testing and continuous parameterisation, and their imperfections should be verified by cross-correlation of records obtained by alternative methods.

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Palaeohydrology and future climate change

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Since the mid-1980s there has been growing concern that increasing concentrations of so-called greenhouse gases will lead to global warming, changes to regional climates, and hence impacts on the environment, society and economy. The United Nations Environment Program and the World Meteorological Organization in 1987 set up the Intergovernmental Panel on Climate Change (IPCC), which has reported on climate change predictions (IPCC 1990*a*, 1992*a*), the possible impacts of climate change (IPCC 1990*b*, 1992*b*) and strategies for mitigating the effects of climate change (IPCC 1991). The IPCC Second Assessment is scheduled for publication in 1996.

Climate change science is essentially predictive: it is trying to predict conditions during the twenty-first century. By far the most common approach is based around the use of global climate models (GCMs). These numerical simulation models are used to predict the global and regional climatic effects of changing greenhouse gas concentrations, and many climate change impact studies use scenarios based in some way on GCM simulations. These scenarios are then used to perturb current climatic time series, and fed through a catchment hydrological model to simulate river flows and other hydrological properties (e.g. Bultot *et al*, 1988; Lettenmaier & Gan 1990; Arnell & Reynard 1993, and many others). The two major problems with this approach lie in the definition of credible catchment-scale scenarios from GCM simulations, and the development of realistic hydrological simulation models. The latter problem is fundamental to hydrological simulation, whilst the former is peculiar to climate change impact assessments and arises for two reasons. First, GCMs do not at present represent all the climatic processes in a realistic manner, particularly those relating to the development of clouds and the interactions between the atmosphere and the land surface, and second GCMs operate at a very coarse spatial resolution. Some important atmospheric processes, such as the development of meso-scale circulation patterns and convective storms, are therefore not simulated particularly well, and whilst GCMs can simulate large-scale atmospheric features well, regional and local climates are often not well reproduced. The coarse spatial resolution also means that GCM output has to be interpolated down to the catchment scale. A variety of techniques of varying degrees of sophistication have been used or proposed, ranging from simple statistical interpolation through empirical relationships between large-scale and local climate to the use of nested regional climate simulation models, but all rely ultimately on the reliability of the GCM simulations of large-scale climatic features (Arnell 1995*a*).

The other popular approach to climate impact assessment uses the past as an analogue for the future. One variant uses the instrumental period, in practice the last century, another uses historical data, and a third uses palaeoclimatic reconstructions.

This paper considers the potential contribution of palaeohydrology to the prediction of future climates and hydrogeomorphological characteristics, focusing on two issues: the contribution of palaeohydrological reconstructions to climate model validation, and the use of palaeohydrological analogues.

Climate modelling and palaeohydrological reconstructions

Global climate models (GCMs) are numerical simulation models, based on the laws of physics and applying the principles of conservation of mass and energy (for a comprehensive review see Henderson-Sellers 1994). They operate on a grid network, with the highest resolution global-scale models currently working on a grid resolution of the order of 250×250 km. Many significant atmospheric processes, including cloud formation at the micro-scale and the development of depressions at the meso-scale, operate at a finer spatial resolution than can be resolved by current GCMs, so are modelled through simplified parameterized representations.

GCMs are used for climate prediction and the development of climate change scenarios, and the credibility of such predictions and subsequent impact assessments is dependent on the credibility of the climate model. It is of course not possible to validate predictions of future climate change, but it is possible to validate simulations of current climate. However, a model that simulates current climate well does not necessarily produce a good simulation of future climate. Fortunately, it is possible to assess the ability of a GCM to simulate changed climates using palaeoclimatic and palaeohydrologic data (Street-Perrott & Roberts 1994).

The CLIMAP and COHMAP projects assembled comprehensive palaeoclimatic and palaeo-vegetation data sets for the period since the last glacial maximum (18 ka BP: CLIMAP Project Members 1976; COHMAP Members 1988), and these data have been used both to provide the boundary conditions for GCM experiments and to validate model simulations (IPCC 1990*a*). There is little information on temperature and precipitation for these early periods, so model validation has largely been based on a comparison of palaeo-vegetation (reconstructed from pollen data) and 'synthetic' vegetation, predicted using current relationships between indices of vegetation and temperature and precipitation. In general, the palaeoclimatic modelling studies have found good agreement between simulations and observations at continental scales (IPCC 1990*a*), although there are regional differences.

The Palaeoclimate Modelling Intercomparison Project (PMIP), part of the IGBP PAGES (Past Global Changes) project, is currently attempting to validate climate model simulations for three periods – 6 ka BP, 18 ka BP and 115 ka BP – using a variety of palaeoclimatic data. The time periods being used were selected both because they represented periods with different boundary conditions (insolation changes at 6 ka BP and the presence of ice sheets at 18 ka BP, for example) and because data were available. The US National Science Foundation project TEMPO (Testing Earth System Models with Palaeoenvironmental Observations) is contributing to PMIP.

In principle, palaeohydrological data can be used for the quantitative evaluation of GCM simulations, although there are a number of complications with using river runoff data for model validation and very few attempts have been made using *current* data (Arnell 1995*b*). One approach that has been adopted is 'top-down', using data from large river basins covering several climate model grid cells (Russell

& Miller 1990). A 'bottom-up' approach would try to build up information characterizing each individual climate model grid cell from small and medium-sized catchments within that cell (Arnell 1995*b*). Palaeohydrological reconstructions have the potential to assist in climate model validation *if it is possible to reconstruct seasonal or annual runoff totals.* Reconstructed lake levels, particularly for lakes in closed basins, are another source of quantitative data on the balance between precipitation, evaporation and runoff (Kutzbach & Street-Perrot 1985; Fontes & Gasse 1991; Harrison & Digerfeldt 1993), and been widely used in climate model validation.

Palaeohydrologic reconstructions as analogues for future climate

In principle, information from the geological past could provide useful insights into the dynamics of the natural environment and links between climate and hydrogeomorphological response, and could provide analogues for future climatic conditions.

Three palaeoclimatic analogues have been identified by the Russian school (IPCC 1990*a*):

(1) the Holocene climatic optimum (6.2–5.3 ka BP), representing an increase in temperature of 1°C;
(2) the last interglacial (125 ka BP), representing a 2°C increase in temperature;
(3) the Pliocene (3 to 4 Ma BP), representing a rise of 4°C.

However, there are a number of theoretical and practical problems with using such analogues.

There are three major theoretical assumptions. First, it is assumed that the relationships between form and process operating today are the same as those operating in the past – the uniformitarian assumption – and that it is possible to infer past process from past form. Second, it is assumed that the data from the past represent equilibrium conditions and this may not be true in practice; much palaeohydrological evidence reflects periods of adjustment to altered conditions. Third, and most importantly, it is assumed that the local effects of a change in climate are independent of the causes of the change in climate. Variations in global climate over geological time scales, however, are a result of changes in the Earth's orbit, and hence changes in the spatial pattern of receipt of solar radiation. Mitchell (1990) showed, for example, that the changes in net radiative forcing during the Holocene optimum were very different to the changes expected under a rising concentration of greenhouse gases; the spatial variability in forcing during the Holocene was considerably greater. Not only are such forcing conditions different, but boundary conditions, especially the extent of ice cover, have varied over geological time, affecting global and regional climates. The effects of such variations in forcing factors and boundary conditions over time may be rather different to the effects of a greenhouse gas-induced forcing, and palaeoclimates do not provide good analogues for future climate change (IPCC 1990*a*).

The major practical problem with the use of palaeohydrological data as an analogue lies in the derivation of quantitative information at relevant time and space scales: in other words, inferring process from form. Different techniques are used for different components of the water balance, and many are illustrated in this volume. Precipitation can be estimated from past vegetation patterns, derived through the analysis of pollen extracted from dated lake and peat bog cores, and from tree ring analysis. Multivariate relationships between vegetation pattern, rainfall and temperature, for example, have been calibrated using current data and applied to palaeo-vegetation patterns to infer both precipitation and temperature (COHMAP Project Members 1988). There are many examples of statistical relationships between tree ring width and precipitation, again calibrated on current climatic data (e.g Till & Guiot 1990; Cleveland & Duvick 1992), although these tend to extend back in time for at most 1000 years.

River discharge can be inferred from palaeochannel form and sediment analysis. Inference from form is based on the use of hydraulic equations, predicting flows from channel dimensions and planform characteristics (Maizels 1983; Gregory & Maizels 1991; Rotnicki 1991). These equations, however, require information which cannot readily be determined from geological evidence such as channel roughness, and it is difficult to draw inferences about hydrological regimes just from data about channel-forming discharges. Sedimentary data are most useful for reconstructing peak flows, again based on current empirical relationships between discharge and sediment deposition patterns. Knox (1993), for example, reconstructed a 7000-year flood chronology for upper Mississippi river tributaries from dated floodplain gravel deposits. He showed that a small change in climate could result in a large change in flood magnitudes. River flows have been estimated from tree rings (e.g. Earle 1993; Loaiciga *et al.* 1993; Young 1994), but this relies on intermediate relationships with precipitation and, as with reconstructions of precipitation, it is difficult to go back even as far as 1000 years.

Records of lake storage are rather easier to construct from both relict erosional features and from shoreline deposits (e.g. Magny 1992), and lake-level changes can give information on changes in the balance of precipitation and evaporation. Groundwater storage at different periods in the past can be estimated through the dating of groundwater (Love *et al.* 1994), and also in some cases by geological and geochemical evidence left behind by higher water tables. It is also possible to model changes in groundwater storage, using reconstructed climate data, validating the model results against dated groundwater or groundwater chemical characteristics (Hiscock & Lloyd 1992).

Whilst it may be possible to derive some quantitative information on past changes in hydrological regime, it is difficult to see how this can be used as an analogue for future climates; the factors causing change are different, both the forcing factors and boundary conditions are different, and the temporal resolution of the data will be low. Many of the events recorded in the geological record reflect changes from glacial to inter-glacial conditions, not changes from one non-glacial state to another.

Understanding landscape and climate dynamics: the contribution of palaeohydrological research to future climate change studies

Although palaeohydrological research does not contribute directly to research into future climate change, it has a major potential role in the development of improved understanding of both the dynamics of climate over different time scales and of the response of the fluvial system to change. The Past Global Changes (PAGES) project of the IGBP is aimed at providing a quantitative understanding of the earth's past environments, as a guide to defining the envelope of natural variability (Eddy 1992), and palaeohydrological research is clearly highly relevant to this aim.

Over the last few years it has become increasingly accepted that climate time series in many parts of the world show clear non-random behaviour, and that there are strong correlations in behaviour, or teleconnections, between different parts of the world. The El Niño/Southern Oscillation phenomenon, for example, affects circulation patterns, rainfall and river runoff every few years across much of the southern hemisphere (Kuhnel *et al.* 1990; Mechoso & Iribarren 1992; Simpson *et al.* 1993), and its signal can be seen in river flows in the western United States (Aguado *et al.* 1992; Kahya & Dracup, 1994). The sub-discipline of hydroclimatology is developing strongly, seeking explanations for such patterns and linkages and attempting to use these explanations to refine understanding of climate dynamics; with an improved understanding of such dynamics and correlations it should be possible to make more credible predictions for the future. Within Europe, one of the aims of the FRIEND project (Flow Regimes from International Experimental and Network Data: Gustard 1993; Arnell 1994) is to investigate links between atmospheric behaviour and large-scale hydrological anomalies, using just the observed river flow record. Palaeoclimatic and palaeohydrological reconstructions can provide valuable long-term time series for hydroclimatological studies; Wells (1990), for example, has reconstructed the history of the El Niño phenomenon in Peru through the Holocene from flood sediments.

Palaeohydrological reconstructions provide information on how the fluvial system responded to change. Even though these changes, such as de-glaciation and recovery from a glacial period, are not directly relevant to future global warming due to increasing greenhouse gas concentrations, the information provided can be used to test models of change and hypotheses about landscape dynamics. Validated models of change are essential for credible predictions of the future.

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Changes in moisture balance between glacial and interglacial conditions: influence on carbon cycle processes

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Abstract: During the arid Late Glacial and Last Glacial Maximum (between approximately 30 000 and 13 000 calendar years ago), vegetation cover retreated and large areas of the continents were occupied by desert and semi-desert vegetation. The result of this general decrease in biological activity would have been a decrease in the size of the land carbon reservoir, and a decrease in the rate of chemical rock weathering. By contrast, during the early–mid-Holocene, conditions in many areas seem to have been moister than today due to a more active hydrological cycle. All of these processes would have affected the global carbon cycle and altered the amplitude and timing of the climate fluctuations themselves. In effect, the climatic shift between glacial and interglacial conditions creates a very large 'missing source' of carbon, perhaps amounting to thousands of gigatonnes, to account for the carbon uptake by the land system during the present interglacial, and thus carbon cycle models of the late Quaternary may need to be revised extensively.

Evidence from polar ice cores indicates that the levels of the greenhouse gases CO_2 and CH_4 in the Earth's atmosphere have closely paralleled the changing climate over at least the last 250 000 years. (e.g. Neftel *et al.* 1988; Barnola *et al.* 1989; Lorius *et al.* 1990; Chapellaz *et al.* 1993; Alley *et al.* 1993) (Fig. 1). There is currently widespread interest in the controls on these changes in atmospheric composition, and global hydrological processes seem likely to play a dominant role in explaining how and why the carbon cycle changed at this timescale. In this paper the influence of these hydrological processes on CO_2 fluctuations is discussed; this influence is considered to be quantitatively much more important than CH_4 in reinforcing glacial-interglacial climate differences.

There can be little doubt that on the glacial–interglacial timescale the dissolved ocean carbon reservoir dominates the system, because it contains many times more carbon than is present in land ecosystems (40 000 Gt as opposed to around 2000 Gt; Bolin *et al.* 1977). Understandably, therefore, effort has mainly concentrated on producing ocean models (Broecker & Peng 1993) to explain the sequestering of larger amounts of carbon into the glacial-age ocean. More recently, however, there have been various attempts to provide a more complete picture by estimating how land carbon storage might have changed between glacial and interglacial conditions

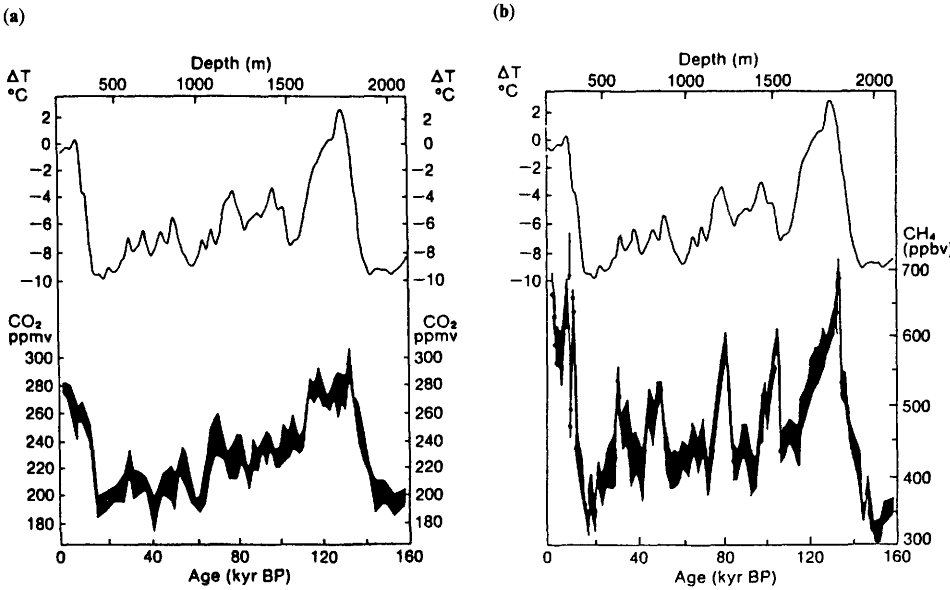

Fig. 1. (**a**) History of atmospheric carbon dioxide levels reconstructed from the Vostok Antarctic ice core record (Barnola *et al.* 1987). (**b**) History of atmospheric methane levels reconstructed from the Vostok Antarctic ice core record (Chapellaz *et al.* 1990).

(e.g. Faure 1989; Adams *et al.* 1990; van Campo *et al.* 1993; Peng 1994; Crowley 1995*a*, *b*; Adams 1995*a*, *b*). Such studies based on palaeoenvironmental reconstruction have suggested that the difference in water balance between full-glacial and full-interglacial conditions was greater than had previously been thought. With this large contrast in the hydrological balance of the continents, combined with major temperature changes, there would have been large changes in their role in the global carbon cycle and a feedback on the climate system itself. Here, the discussion concentrates on the picture suggested by recent mapping efforts to reconstruct the world's land vegetation cover based on fossil and palaeoclimatic evidence.

Evidence for changes in global water balance

Following the earlier view that glacial periods were always cooler and moister than at present, by the early 1950s it became apparent that the Last Glacial or oxygen Isotope Stage 2 (a period extending between about 30 000 and 13 000 years BP), and in particular the Last Glacial Maximum (LGM, about 20 000 years BP), was more arid than present throughout most of the world. As more evidence from the LGM has been presented, the extent and severity of this stage of aridity has become more strikingly evident. In contrast, evidence for moister-than-present conditions during the early–mid-Holocene has also been found for many parts of the world.

There have been a number of attempts to bring together evidence that relate, directly or indirectly, to changes in continental moisture balance over this timespan. The history of changes in water level in lake basins is a major source of data as a palaeohydrological indicator (Street-Perrott & Roberts 1983; Wright 1993).

Additionally, pollen databases such as that of Webb *et al.* (1995) use the plant fossil remains of past vegetation cover, which can be used to reconstruct the general climatic conditions of the past.

The baselines for general palaeoclimatic 'atlases' are being established (Wright 1993), but the process is in its early stages for many regions because of the shortage of data than can be readily translated into a quantitative form. There have been several recent attempts to reconstruct palaeovegetation on a global scale, using fossil and/or sedimentological evidence. These include maps by van Campo *et al.* (1993), Adams (1995*b*) and Crowley (1995*a*). Such palaeovegetation maps can be regarded as an indirect summary of the hydro-climatological regime of the past; for instance closed forest vegetation (defined in terms of a forest canopy cover formed by the tree crowns being close enough to touch one another) indicates moister conditions than desert or semi-desert. A recent attempt to collate the range of these and other sources of data into a coherent picture of biome distribution has been the Quaternary Environments Network (Adams 1995*a*, *b*; QEN 1995; Adams & Faure in press) review of palaeovegetation cover, based on a range of sources of plant fossil, sedimentological and zoological evidence which relate directly or indirectly to vegetation cover. The QEN biome map reconstructions for the LGM, the early Holocene and the present-natural illustrate the extent of change in the water balance, with the moist-climate tropical and temperate rainforest area being much reduced at the LGM (indicating aridity), and expanded during the early Holocene (due to more humid climates). Global deserts show the opposite pattern, being more extensive during the LGM and less extensive than present during the early–mid-Holocene. The Sahara desert almost disappeared under a vegetation cover during the early Holocene, indicating a moister climate than at present, and the same seems to have been true of the central Asian deserts.

GCM modelling experiments, using prescribed sea surface temperatures and ice distributions, also suggest that the LGM world was substantially drier than at present, and that the early–mid-Holocene would on average have been moister due to the Milankovitch effect of greater summer radiation on the northern continents (Prentice *et al.* 1993; Wright 1993). There is a disparity of evidence in that the computer model simulations for the LGM predict conditions substantially moister than the palaeoevidence would appear to suggest. However, such computer simulations are still at an early stage of development, with heat flux corrections continually necessary to make them conform to 'realistic' climate patterns (Crowley & North 1991; Kagan 1995). The boundary conditions against which the models are established and continually returned to may also turn out to be incorrect. For example, the prescribed sea surface temperatures (e.g. CLIMAP 1976) that published model studies have always used are now thought to be too warm for the LGM of the tropics (Crowley 1994; Broecker 1995). If LGM sea surface temperatures used in the models were reduced, the result would be even drier conditions over the continents, bringing the results more closely into line with palaeoevidence. However at present there is too much uncertainty regarding the reliability of GCM models to use them to reconstruct global changes in the past hydrological regime. For this reason we concentrate here on the picture gathered from more direct forms of palaeoevidence (Starkel 1989).

When discussing past vegetation cover and ecosystem processes in relation to past hydrological balance, it is necessary to recognise that the observed changes may not be the result of real climatic changes in rainfall or potential evapotranspiration. This is because the hydrological balance of plants and vegetation is also partly controlled by the ambient CO_2 concentration: the less CO_2 there is available, the more that plants need open their stomata (the pores in their leaves) to obtain it (Woodward 1987). The more that plants open their stomata, the more moisture they lose. Thus, lower CO_2 levels alone could mimic the effects of a 'drier' climate, causing vegetation zones to shift their positions along climatic gradients. However, the true importance of the relationship between CO_2 changes and vegetation ecology in the past remains highly uncertain, despite a great deal of speculation. Lake level evidence suggests that there was much lower rainfall at the LGM in most areas, relatively independent of vegetation processes (Street-Perrott & Roberts 1983; Wright 1993). Closed chamber experiments studying the effects of raised 'future' CO_2 levels on plants give confusing and sometimes contradictory results (McConnaughay *et al.* 1993; Koerner & Arnone 1993; Mooney & Koch 1994), and there is no clear evidence that the rapid 80–90 ppm CO_2 rise over the last 200 years has had any effect in increasing vegetation cover and productivity (Adams & Woodward 1992; Shiel & Philips 1995). It seems reasonable to conclude at present that most of the aridity of the LGM world was due to a reduction in rainfall and not 'apparent' aridity due to a direct effect of CO_2.

Changes in bulk carbon storage

For practical purposes of study, the world's vegetation cover of 'biomes' can be used to define around 35 'ecosystems', including not only the vegetation of each biome but also the litter, soils and peat layers underneath. Carbon storage in the components of these land ecosystems is closely related to local and regional hydrological conditions. Thus, in almost any given region of the Earth's surface, there is a striking broadscale relationship between annual precipitation and natural or semi-natural vegetation carbon storage. The relationships are even closer if differences in temperature are considered as a factor affecting moisture availability. Forest ecosystems, the richest in terms of *vegetation* carbon storage (Olson *et al.* 1983; Harmon & Hua 1992), require a greater moisture supply than dry grassland ecosystems, and if there is no moisture available, there is a desert with no vegetation carbon store. In a study of the Mediterranean region, Le Houerou & Hoste (1977) report an exponential relationship between vegetation biomass and annual precipitation. Even within individual biomes/ecosystems, such as the steppe biome, there is often a strong relationship between annual rainfall and vegetation carbon storage (Le Houerou & Hoste 1977; Olson *et al.* 1983).

Soil carbon shows a similar positive relationship with moisture availability (the balance between precipitation and evaporation), illustrated in the summary diagrams of Fig. 2 and Post *et al.* (1982). This is because the rate of input of dead plant parts from the vegetation to the soil tends to be greater under moist conditions, other factors being equal. Also, above a certain level of moisture content the soil is too waterlogged to allow efficient aerobic decay of plant material, thus promoting peat formation.

Given such simple, general relationships between moisture availability and ecosystem carbon storage, it is evident that a change in the moisture regime, whether due to real climatic changes or to a direct-CO_2 effect on plant water balance, would affect global carbon storage on land. With respect to carbon storage, attention

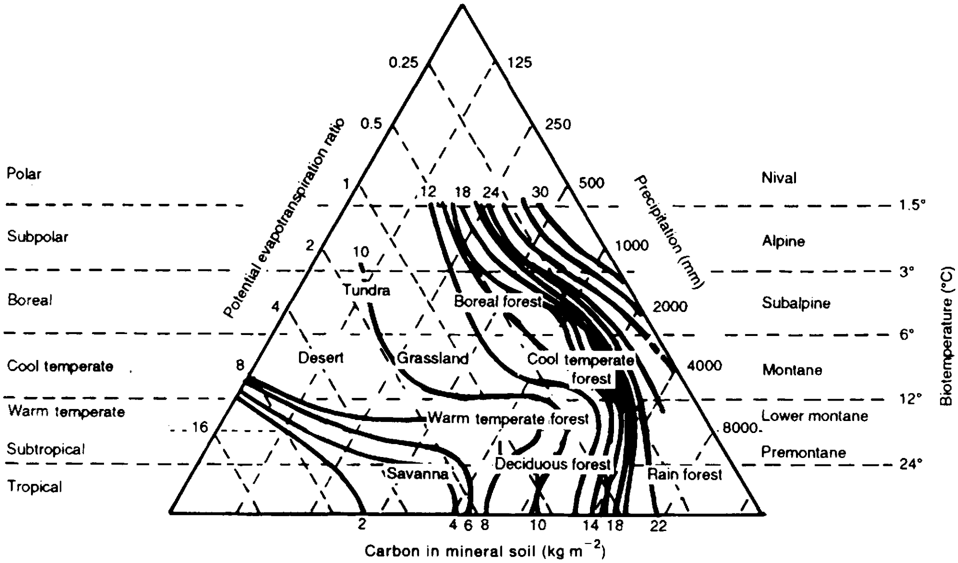

Fig. 2. Summary of the relationship between climate and soil carbon storage found in the database of Zinke *et al.* (1984). Reprinted with permission from *Nature*. (Post *et al.*, vol. 298, p. 156). Copyright (1982) Macmillan Magazines Ltd.

has focused on the LGM as the most 'extreme' part of the Stage 2 period in terms of aridity, but the same general picture is also relevant for much of the time period, particularly after *c.*25 000 BP. There are a number of methods to estimate the change in carbon storage between the opposite climatic extremes of the LGM and the early–mid-Holocene. These methods are in three different categories; (i) reconstruction of palaeovegetation/ecosystem cover for glacial and interglacial conditions (e.g, Adams *et al.* 1990; Crowley 1995*a*, *b*), (ii) use of general circulation models (GCMs) to 'predict' past climates and using these data to 'predict' palaeovegetation cover and carbon storage (e.g. Prentice *et al.* 1993), (iii) taking the magnitude of the carbon isotope change in the oceans between glacial and interglacial conditions as an indicator of the size of the land carbon storage change (e.g. Shackleton 1977). All three methods suggest that there was a net increase in land carbon storage of *at least* several hundred gigatonnes between the LGM and the mid-Holocene (Fig. 3) (Maslin *et al.* 1995; Adams 1995*a*, *b*). The ocean-isotope method gives the lowest estimates (350–700 Gt) (Shackleton 1977; Currey *et al.* 1988; Broecker & Peng 1993) with palaeoevidence-based estimates tending to be larger at around 700–1700 Gt, with GCM-based studies giving intermediate values (Freidlingstein *et al.* 1992; Prentice *et al.* 1993).

Each method of estimating total land carbon storage has inherent sources of error (Maslin *et al.* 1995; Crowley 1995*a*, *b*), and it appears that the interdisciplinary palaeoevidence-based approach is likely to be the most robust as it offers the most direct indication of the changing nature of land ecosystems themselves. Of the palaeoevidence-based studies (Fig. 3), several have produced estimates of a relatively low shift in carbon storage (400–1000 Gt C) due to (i) them incorporating a narrow range of palaeoenvironmental evidence, (ii) the assignment of present-day low anthropogenic carbon storage values to essentially pre-anthropogenic Holocene forest

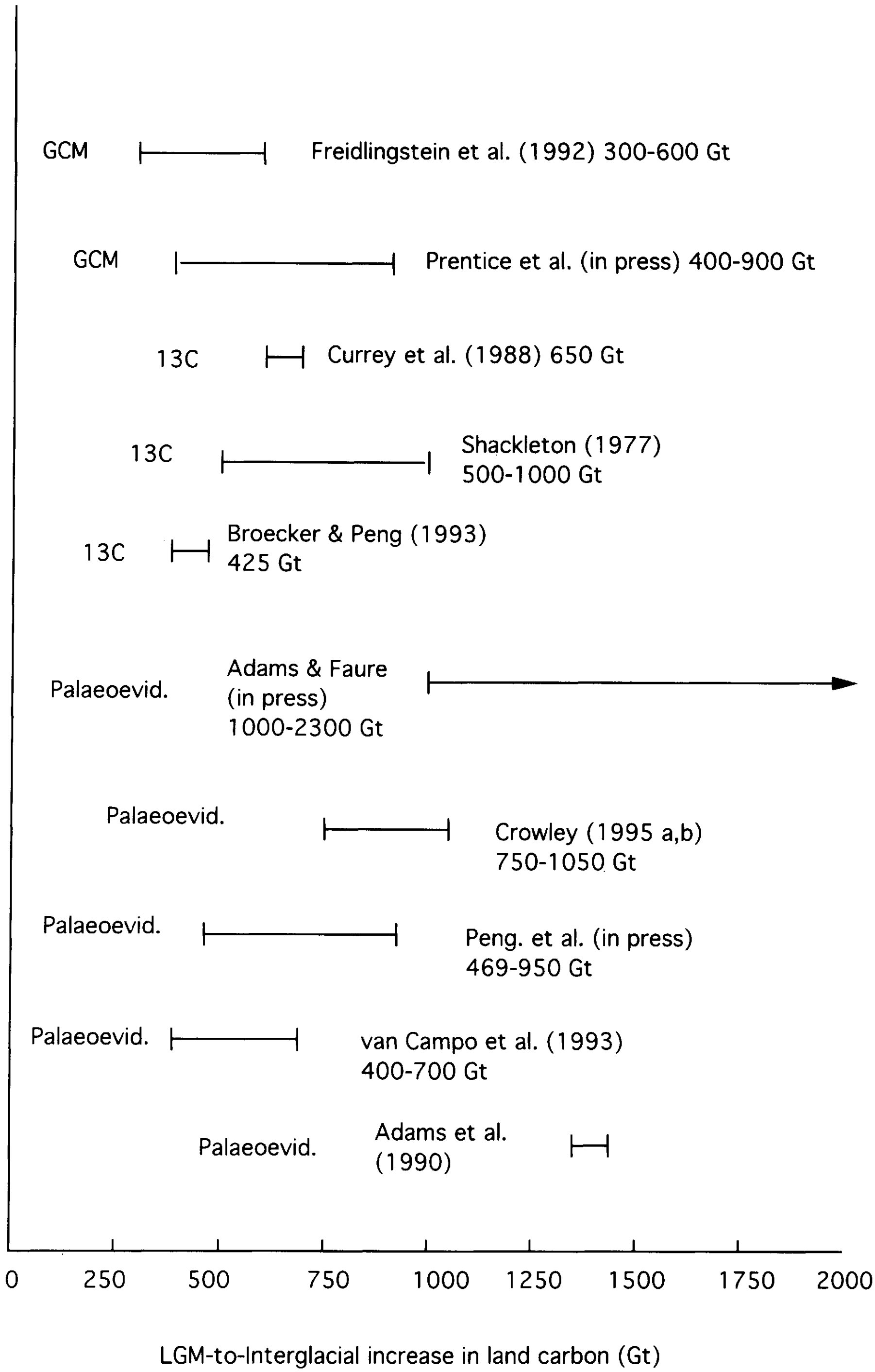

Fig. 3. Estimates for the LGM–Holocene increase in organic carbon in vegetation, soils and peatlands on land. After Maslin *et al.* (1995) Adams (1995*b*), gathered from sources not directly cited here. Palaeoevid, palaeoevidence-based reconstruction of ecosystems; GCM, climate model-based prediction of past ecosystems; ^{13}C, estimate based on shifts in the ocean $^{13}C/^{12}C$ ratio.

ecosystems, and (iii) the assumption that LGM desert and semi-desert ecosystems were as rich in carbon as present-day moist steppe. Given the bulk of evidence of global LGM aridity and carbon-poor soils, it seems appropriate to concentrate here on scenarios in which a relatively large (approximately 1000–1700 Gt) shift in land carbon storage occurs, though without neglecting the scenarios in which a somewhat smaller shift (around 500 Gt) occurs.

If one takes a conservative value of around 1000 GtC as the amount of carbon released from land ecosystems as the Earth shifts from interglacial to full glacial conditions, this would release 476 ppm of CO_2 into the atmosphere (if 2.1 Gt C $\approx$ 1 ppm CO_2). On the timescale of thousands of years this CO_2 would distribute itself between the atmosphere and the ocean, with the largest fraction of the CO_2 being dissolved into the ocean water as CO_2 gas and as bicarbonate. According to the alkalinity buffer system (Siegenthaler 1989), around 16% of the extra CO_2 would stay in the atmosphere, adding to the total atmospheric CO_2 level by *c.* 66 ppm. For a shift of *c.* 500 Gt C, the increase of CO_2 during the glacial period would be *c.* 35 ppm, and for around 1700 Gt C it would be *c.* 115 ppm. If this were the only factor in the carbon cycle to vary between glacial and interglacial conditions, the lower glacial-age land carbon storage would give an atmospheric CO_2 level 44–115 ppm higher than during the interglacial. In fact, the CO_2 level was some 80 ppm *lower* during the LGM than during the Holocene. The reduction of CO_2 was probably caused by the oceans taking up more carbon under glacial conditions than during interglacial conditions.

In general, previous attempts to understand the global carbon cycle during the LGM have only recognized that there was a reduction of 80 ppm atmospheric CO_2 under interglacial conditions, requiring the removal of about 170 Gt C from the atmosphere into the ocean. Various scenarios of change in oceanic processes, such as an increase in the 'biological pump' or a slowing of the rate of deep ocean circulation, would allow this extra amount of carbon to be held in the oceans under LGM conditions (reviewed by Broecker & Peng 1993). Unfortunately, if the 'extra' carbon that would enter the land system during the glacial–interglacial transition is added, then there are problems with the ocean models. Adding a modest 60 GtC of land carbon (given a very low estimate of the LGM–Holocene land carbon increase of 425 Gt C, and after allowing for the alkalinity buffer), Broecker & Peng (1993) found that several ocean carbon uptake mechanisms working in parallel might have been able to remove this extra carbon. However if in a more realistic quantity such as 140 Gt C or more is added (allowing for an increase in land carbon by 1000 Gt C), existing models are unable to explain the sink for this carbon during the LGM. It was probably held within the oceans, but existing models do not adequately explain *how* it was held there. Thus, consideration of the land carbon storage budget illuminates the need for a better understanding of ocean carbon cycle processes during the last glacial.

Another result from the consideration of the varying size of the land carbon reservoir is its role in opposing the reduction of CO_2 by the oceans under LGM conditions. If the release of CO_2 from the land had not occurred during the onset of glacial conditions it can be surmised that the uptake of atmospheric carbon by the oceans would have occurred nonetheless. Without the land source to replenish it, atmospheric CO_2 would have been *at least* several tens of ppm lower than it actually was. Given that the 80 ppm lower CO_2 level is generally seen as crucial to the magnitude of cooling that occurred during the last glacial, an even lower level of CO_2 would have resulted in a greater extent of glaciation. Indeed, since the Earth appears to be continually on the periphery of an ice-up of its surface (Crowley & North 1991; Lovelock & Kump 1994; Kagan 1995), one might regard the extra carbon released by the land during glaciations as preventing a catastrophe caused by excessive ocean carbon uptake.

Surface weathering

There is a continual flux of carbon dioxide out of the atmosphere through chemical reactions involving minerals in rocks and soil. The two main types of weathering reaction which take up carbon dioxide are:

Silicate weathering, involving reactions such as the following

$$CaSiCO_3 + 2CO_2 + 2H_2O \Rightarrow Si(OH)_4 + Ca(HCO_3)_2$$

and carbonate weathering

$$CaCO_3 + CO_2 + H_2O \Rightarrow Ca^{2+} + 2HCO_3^-$$

The first reaction provides an irreversible sink for carbon dioxide on the glacial–interglacial timescale, whereas the second is much more easily reversible. Nevertheless, both reactions provide a considerable flux of carbon out of the atmosphere each year; only part of which is balanced by dissolution of bicarbonate back into carbonate, carbon dioxide and water. Probst (1992) has calculated that the annual flux of inorganic carbon into rivers from these weathering reactions is 0.39 Gt C/a^{-1}. Although part of this carbon is 'ancient', derived from the carbonate in limestones, for example, most of this is derived more recently from atmospheric CO_2 and would be enough to empty the atmosphere of carbon in only a few thousand years if it were not compensated by metamorphic and volcanic output.

If the weathering sink varies on a glacial-interglacial timescale this too could have a significant effect on the whole global carbon cycle. Present-day weathering rates are strongly dependent on moisture regime, temperature and rock type, and vegetation and the soil biota catalyse the reactions (Velbel 1993; Ludwig *et al.* 1995). Given that conditions were generally drier and less vegetated during glacial periods, weathering rates would probably have been lower than during the moist, warm interglacial optima (Adams 1995*a*). Weathering rate also shows a strong temperature relationship if rock type and water availability are considered (Velbel 1993; Volk 1993), and the globally cooler conditions of the Isotope Stage 2 period (and the previous cool stages 5e–3 together spanning the last *c.* 100 000 years) would have suppressed weathering even further. On the basis of the weathered carbon export in rivers estimated by Probst (1992), Adams (1995*b*) has estimated the LGM global weathering rate, based on past distributions of biomes according to the QEN Atlas, and suggests that in contrast to the 0.39 Gt C/a^{-1} flux at present, LGM weathering uptake would have been about 0.15 Gt C/a^{-1} (Tables 1 & 2). Although the riverine carbon export rate overestimates total carbon uptake by weathering (and at least some carbon may be released back to the atmosphere by carbonate precipitation on reaching the oceans), the general point holds true. The estimate of the change in area of moist montane environments remains speculative, but a

Table 1. *Regrouped areas of global vegetation types ($\times 10^6$ km^2)*

	18 000 years ago	8000 years ago	5000 years ago
1. Mesic taiga	–	7.575	6.701
2. Temperate humid	0.762	4.905	4.975
3. Tundra, mid-Taiga	4.235	17.308	17.865
4. Temperate mesic	0.025	16.947	18.273
5. Temperate submesic	8.558	9.947	9.002
6. Sub-desert/dry grassland	34.672	22.640	26.683
7. Absolute desert, lake swamp, ice	82.197	21.421	20.036
8. Tropical seasonal	24.547	24.647	23.404
9. Tropical humid	7.502	23.229	22.615
10. Tropical moist montane	0.543	1.206	1.105
Total areas	163.000	150.000	150.000

After Adams (1995*b*) & QEN data base maps.

difference of this general order remains a realistic scenario because the world's mountain ranges and plateaus were generally much drier and colder during the Isotope Stage 2 period and indeed for most of the 70 000 years preceding this (Li *et al.* 1995; Crowley & North 1991). Considering that even a 0.1 Gt C/a^{-1} increase in Holocene weathering consumption (relative to the Last Glacial) would add up to 1050 GtC over the timespan of about 10 500 calendar years since the early Holocene, this factor has the potential to create some major 'missing sinks' and 'missing

Table 2. *Calculated globalweathering rates based on river bicarbonate transport*

	HCO_3 km$^{-2}a^{-1}$ (t)	18 000 years ago	8000 years ago	5000 years ago
1. Mesic taiga	12.7	–	96.2	72.402
2. Temperate humid*	72.3	55.092	354.000	359.692
3. Tundra, Mid-taiga	12.4	52.512	214.61	221.526
4. Temperate mesic	31	0.775	525.357	566.463
5. Temperate submesic	24.4	208.815	242.706	219.648
6. Sub-desert/dry grassland†	2.5	86.672	66.500	76.707
7. Absolute desert, lake swamp, ice	0	0	0	0
8. Tropical seasonal	5.35	131.326	131.861	125.211
9. Tropical humid	16.5	123.783	379.500	370.147
10. Tropical moist montane	69.9	37.955	84.299	77.239
Total HCO_3 export in rivers		696.930	2095.033	2089.032
$\times$ 0.196 (mole fraction of C) $\times$ 0.001 to convert into Gt				
Inorganic C export in rivers		0.135 Gt	0.410 Gt	0.409 Gt

* Includes temperate sub-humid.
† Includes both temperate and tropical .
After Adams 1995(b) & Probst 1992.

sources' in the global carbon cycle. If metamorphic/volcanic input remained relatively constant throughout this period, then the source of the 'extra' hundreds, or even thousands, of gigatonnes of carbon to sustain atmospheric CO_2 levels during interglacials, or the carbon sink during cold, dry glacial periods instead of being taken up in weathering and accumulating as CO_2 in the atmosphere, should be identified. It is possible that the late-interglacial pattern of decline in CO_2 observed in ice-core data from the Eemian is related to 'exhaustion' of the supply of CO_2 by a sustained bout of weathering under interglacial conditions.

The problem of balancing the carbon budget on the glacial–interglacial timescale is made even more difficult because when a glacial phase ends, the land surface is 'charged' with weatherable minerals that are unstable under the new, moister and warmer climate regime (Adams 1995*a*). For example, these minerals would have been in the form of finely divided silicate and carbonate minerals in glacial outwash deposits, desert surfaces and loess deposits, and also soil carbonate in the formerly semi-arid regions that occupied much of the world during the last glacial (Fairchild *et al.* 1995). Although difficult to quantify, the global uptake of CO_2 by weathering during the early stages of moist interglacial conditions could have been much higher than present-day observations on soils would suggest. Furthermore, the reduction of weathering CO_2 uptake during glacial conditions would allow the atmospheric CO_2 level to remain higher than would otherwise be the case, thereby providing an additional factor which would raise LGM CO_2 levels in opposition to oceanic uptake of carbon.

Lake organic sediments

Large areas of the Earth's surface are at present covered by freshwater lakes, particularly in the boreal zones of the Northern Hemisphere. Most of the millions of small lakes in the high-latitudes of North America and Siberia have formed since the onset of moist Holocene conditions. Highly organic-rich oozes often reach many metres in thickness, generally having accumulated since the beginning of the Holocene. A large number of small oxbow lakes and layers of organic-rich floodplain sediments also exist in tropical river valleys in the rainforest zones; again these seem to be generally a product of the moist Holocene conditions that replaced the drier Glacial phase.

Freshwater lakes did exist during the LGM–Late Glacial period, but they were less abundant. Large areas of the high latitudes were dominated by arid conditions and aeolian erosion (Spasskaya 1992; Velichko & Spasskaya 1992). Notable exceptions were the large proglacial lakes contained by the Laurentide, Fennoscandinavian and Uralic ice sheets (Goncharov 1989; Velichko & Spasskaya 1992). In contrast with most of their present-day boreal counterparts, these lakes were generally characterized by grey or yellow organic-poor sediments due to the combination of very low water temperatures and aridity in their catchments (Goncharov 1989; Spasskaya 1992; Velichko & Spasskaya 1992).

If an average organic carbon content of $30\,kg/m^{-2}$ is assigned to the areas of lake sediments which accumulated since the start of the present interglacial alone, for a global lake area of $2.5 \times 10^6\,km^2$ (Olson *et al.* 1983; Adams 1995*b*) one obtains a newly formed reservoir of 75 Gt C. This must be offset against the rate of lake organic sedimentation that occurred during the last glacial period, and those glacial lakes which have emptied and released their organic carbon back into the general carbon cycle. Even without being able to completely compensate for this it is evident that there would still have been a greater rate of organic lake sedimentation during the Holocene as opposed to the LGM–Late Glacial. This increase represents a drain on the ocean–atmosphere carbon reservoirs, which nevertheless has kept the CO_2 level higher during Holocene conditions. Further 'missing reservoirs' during glacial conditions are needed to explain the source of these tens of gigatonnes of carbon.

Sea-level fall

The land system did not lose water during glacial conditions, it gained it. The extra water was stored in the ice sheets that covered *c.* $37 \times 10^6\,km^2$ of the Earth's surface, which held an extra *c.* $40 \times 10^6\,km^3$ of water (Lorius 1991), and in consequence, the sea level fell by 100–120 m; a 3% reduction in ocean volume relative to the present. The freshwater of which ice sheets are composed contains very little carbon and therefore the ocean must have retained all of the carbon that was previously present in this 3% of its volume, representing a net gain of *c.* 1200 Gt C ($40\,000 \times 0.03$) by the ocean waters that remained. This represents another, very considerable, 'missing' amount of carbon whose disappearance needs to be accounted for if the carbon cycle is to be fully understood, but it is surprisingly not allowed for in any ocean models that seek to explain the carbon cycle during the last glacial period.

An overview: implications of the land system for the carbon cycle

In response to large-scale changes in hydrological and temperature regime, the Holocene land system appears to have 'gained' a vast amount of carbon from some unknown source or sources, possibly the ocean (Fig. 4). The bulk organic carbon storage reservoir in vegetation, soils and peats probably gained at least 1000 Gt between LGM and Holocene conditions. The lake-bed and groundwater carbon reservoirs likewise probably accounted for at least several tens of gigatonnes of the Holocene carbon uptake by the land system. Greatly increased weathering rates during moist Holocene conditions seem likely to have taken in hundreds or perhaps thousands of gigatonnes extra carbon, relative to the cumulative average weathering rate during the last glacial period.

It is important to note that these changes in land reservoirs and sinks did not operate strictly simultaneously. For example, the lack of a strong (interglacial level) weathering sink during the last Glacial (Isotope Stage 2) period was cumulative and thus reached its peak importance towards the end of this arid period, whereas the land organic carbon reservoir in vegetation and soils was at its low point at the LGM and was already increasing by the end of the last Glacial. Nevertheless, in general these sinks and reservoirs can be seen as combining to highlight the existence of a major 'missing source' of carbon that was not translated into higher glacial CO_2 levels. Furthermore, the decrease in ocean volume that was associated with the growth of ice sheets on land would have pushed a further burden of more than 1000 Gt C into the remaining bulk of the oceans.

The closer that the carbon cycle on the glacial–interglacial timescale is investigated, the less it seems that anyone truly understands it. The entry of this large amount of carbon into the land system under Holocene conditions must be explicable in terms of one or more the following three hypotheses.

(i) The operation of unknown mechanisms within the oceans for storing carbon under glacial conditions, in addition to the extra source from the reduction in ocean volume. This mechanism appears the most plausible of the three possibilities at present.

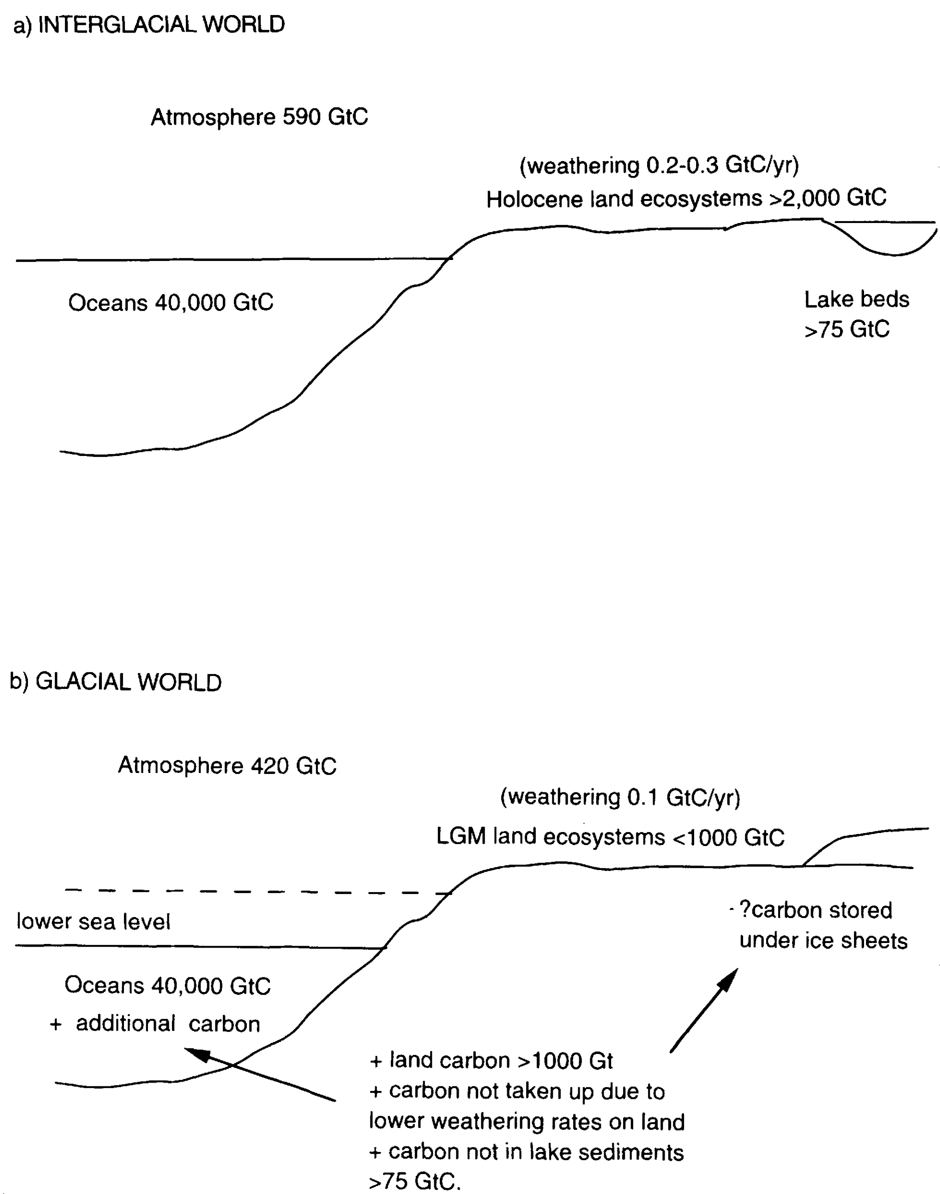

Fig. 4. Box diagram showing some of the major changes in the carbon cycle between glacial and interglacial conditions. During the arid glacial stage, a large amount of interglacial carbon is 'missing' and is presumably held either in the oceans or under ice sheets.

(ii) Retention of organic carbon from earlier (e.g. Eemian Interglacial, or interstadial) ecosystem cover underneath ice sheets, which was released under during deglaciation, as suggested by Franzen (1994). However, although buried pre-glacial carbon does occur in some localities, the lack of organic content in most moraine deposits from the last glacial (e.g. West 1978; Williams *et al.* 1993) contradict this hypothesis. It would appear that most land ecosystem carbon re-enters the broader carbon cycle at the onset of glaciation, rather than being covered by ice or held in permafrost.

(iii) An increase in volcanism since the beginning of the Holocene, which would supply extra carbon of which a large part was taken up by the land system. There is some evidence for this in the Mediterranean region (mentioned briefly by Caldiera 1992), but its true importance remains unknown.

Implications for the global climate system

The hydrological cycle seems to exert a major influence on glacial–interglacial changes in carbon reservoirs through its impact on a mixture of biologically linked and non-biological processes. It appears that the combined effect of reduced land carbon reservoirs, reduced ground water and lake bed reservoirs and weathering rate, would have allowed the CO_2 level at the LGM and during the Last Glacial to remain substantially higher than would have been the case if these continental reservoirs and processes had continued in their previous interglacial mode. Thus it seems that there is a feedback between the global climate and the terrestrial carbon cycle; as climate enters its full glacial mode of cold and aridity, the terrestrial system acts as a damper and releases carbon to prevent climate conditions becoming more extreme. Likewise, when climate is in its full interglacial 'optimum' state, the moist, warm conditions favour carbon uptake into the land system, thereby damping the warming. Viewed in this sense, the changes in the terrestrial carbon cycle that occur between glacial and interglacial conditions act as negative feedbacks (damping loops) to stabilize the Earth's climate and atmospheric composition against the positive feedbacks (amplifying loops) which involve ocean carbon uptake, and also the ice sheet albedo on land and sea ice albedo in the oceans. However, it should be considered that land ecosystems also amplify the broad climate oscillations in their own positive feedback loops involving climate. These include their effects on surface albedo, methane and dust contribution to the atmosphere, although the quantitative significance of these effects remains uncertain.

The overall picture that one gains from considering the terrestrial system in its various capacities is that even a single component, such as vegetation cover, can simultaneously exert a damping influence (e.g. by taking up CO_2 and promoting rock weathering as conditions become moister and warmer) and an amplifying influence (e.g. by increasing its surface coverage and thereby reducing dust fluxes to the atmosphere as global conditions get moister and warmer) on the glacial-interglacial climate cycle. There is much that is still not understood about the whole system, but what is already evident is that land ecosystem processes, and the terrestrial carbon cycle in particular, must play a major role in producing and/or modifying the observed pattern of glacial–interglacial climate oscillations.

Further research is needed on the processes by which palaeoclimatic variations may have been damped or amplified by their effects on biotic processes if an improved understanding of earth history is to be reached. It is possible that positive and negative feedbacks between the carbon cycle and global palaeohydrology have been important in the evolution of climate on timescales both shorter and longer than the glacial-interglacial cycle considered here. Such processes may also prove to be important in the near future due to anthropogenic modification of climate and ecosystems.

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Erosion and sediment yield in a changing environment

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Abstract: Existing assessments of patterns of global sediment flux from the land to the oceans and of global patterns of sediment yield have tended to treat the global denudational system as a static system. There is, however, increasing evidence that the sediment loads of the world's rivers have changed significantly as a result of human activity and interference. As well as being of scientific interest, such changes have important environmental and economic implications and should be included in current concerns for the impact of global change on the Earth's environment. Lack of reliable long-term records precludes a detailed assessment of longer-term and recent changes in suspended sediment transport by world rivers, but the available information can be supplemented by evidence from several other sources, which include the long-term geological perspective, lake sediment records, catchment experiments and space-time substitution; all of which provide valuable information concerning the sensitivity of river sediment loads to environmental change and the magnitude of the changes involved. There is evidence of significant increases and decreases in sediment yields in many areas of the world. Any attempt to relate such changes in sediment loads to environmental change within the upstream drainage basin must, however, take account of the complexity of the sediment delivery system.

With the current concern for global change and the impact of both climate change and human activity on the global system, there is clearly a need to consider erosion and sediment yield as a key component of the system and to assess their sensitivity to environmental change. Knowledge concerning changing sediment fluxes is important to studies of rates of landform development, terrestrial inputs to the oceans and global element budgets, but such changes can also have important environmental, economic and social implications relating to land degradation and reduced crop productivity, increasing rates of reservoir sedimentation, destruction of aquatic

Table 1. *A comparison of soil erosion rates under natural undisturbed conditions and under cultivation in selected areas of the world*

Country	Natural ($kg\,m^{-2}\,a^{-1}$)	Cultivated ($kg\,m^{-2}\,a^{-1}$)
China	<0.20	15.00–20.00
USA	0.003–0.30	0.50–17.00
Ivory Coast	0.003–0.02	0.01–9.00
Nigeria	0.05–0.10	0.01–3.50
India	0.05–0.10	0.03–2.00
Belgium	0.01–0.05	0.30–3.00
UK	0.01–0.05	0.01–0.30

Based on Morgan (1986).

habitats, river management and adverse effects on river water quality (Clark *et al.* 1985; Mahmood 1987; Braune & Looser 1989).

There is widespread evidence that soil erosion rates on cultivated land are frequently up to several orders of magnitude greater than those under natural undisturbed land (Table 1), and the global implications of such trends are clear when it is recognized that the area of the Earth's surface given over to crop production and livestock grazing has increased more than five-fold over the past 200 years (Buringh & Dudal 1987) and that the recent ISRIC/UNEP global survey of human-induced soil degradation (Oldeman *et al.* 1991) has shown that nearly 10% of the total land surface of the globe is currently adversely affected by water erosion. A substantial proportion of the sediment mobilized by accelerated erosion is likely to find its way into rivers and there are reports of increased sediment yields in many areas of the

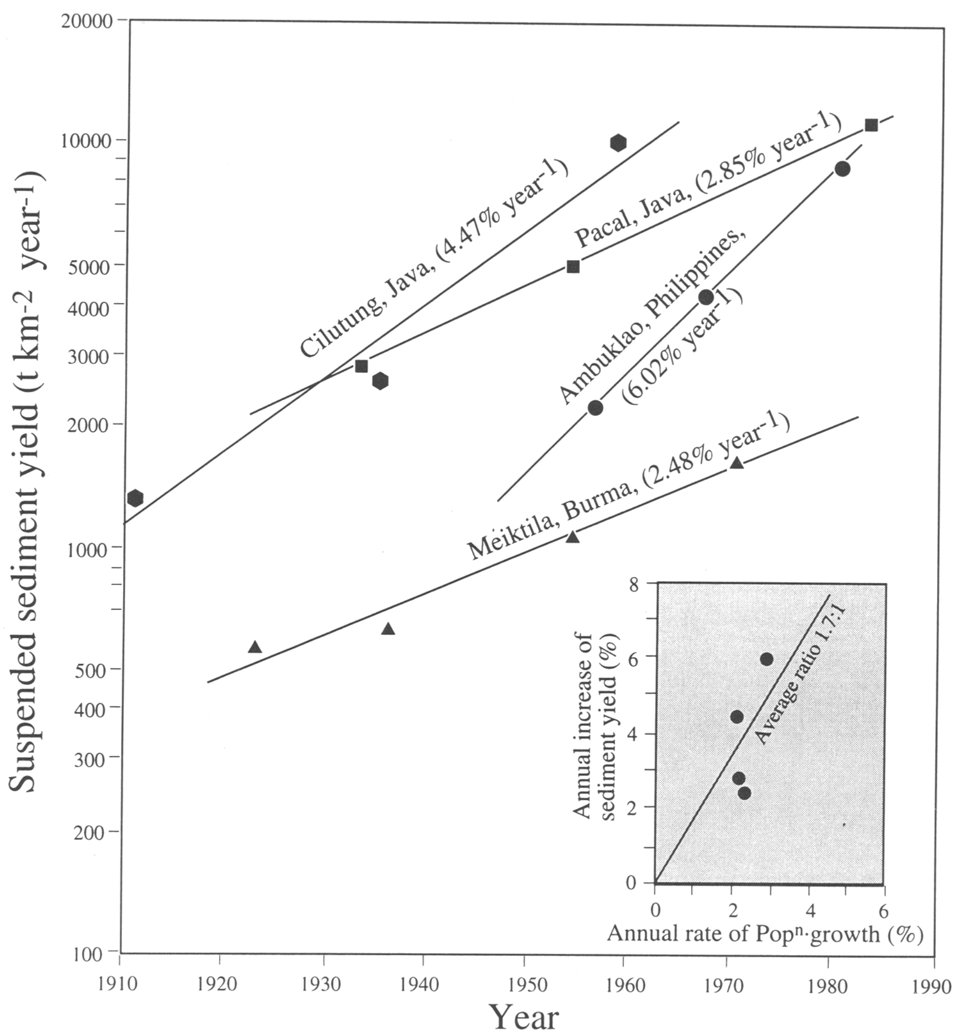

Fig. 1. Trends of increasing sediment yields in selected reservoir catchments in Southeast Asia (based on Abernethy 1990).

world, and particularly in developing countries (Douglas 1967). Figure 1 based on the work of Abernethy (1990), demonstrates the magnitude of the increase documented for several small reservoir catchments in Southeast Asia which have experienced substantial land clearance and land use intensification during the present century. In these catchments, sediment yields have shown annual increases of between 2.5% and 6.0%. Abernethy (1990) suggested that these increases closely paralleled the rate of population growth in the areas concerned (Fig. 1 inset), although he found that the ratio of the rate of increase of sediment yield to that of population was greater than unity. Based on this evidence, he suggested that annual suspended sediment yields in many developing countries were currently increasing at a rate equivalent to 1.6 times the rate of population increase, and could therefore be expected to double in about 20 years. Contrasting situations will, however, exist in other areas of the world where reservoirs now trap a major proportion of the sediment formerly transported by the rivers. Thus, for example, in the classic case of the River Nile, the construction of the Aswan High Dam has caused the annual suspended sediment yield at the mouth of this major river to decrease from *c.* 100 million tonnes to near zero.

It is clear that river basin sediment yields can be highly responsive to changes in erosion rates within their upstream catchments consequent upon changing land use or climate change and may also be impacted by the construction of reservoirs and other river engineering works. Any attempt to provide a more general assessment of the sensitivity of the sediment yields of the world's rivers to such environmental change and of the magnitude of the changes involved is, however, hampered by the general lack of reliable long-term records of sediment yield in most areas of the world. Few, if any, records extend back beyond the present century and, in view of the many problems associated with obtaining accurate assessments of river loads (Walling & Webb 1981, 1987, 1988), the reliability of early records is frequently open to question. Evidence based on available river records must therefore be supplemented by additional sources of information. These sources involve a variety of timescales ranging from a long-term geological perspective, through the evidence provided by lake sediments which extends back over timescales of 10^2 to 10^3 years, to recent catchment experiments and the potential for using contemporary data to provide a longer-term perspective by means of space-time substitution. Some of the evidence provided by these various sources of information can usefully be reviewed.

The long-term geological perspective

Information on the present-day distribution of sedimentary rocks on the Earth's surface can provide a basis for estimating sedimentation rates at different periods in the past (Gregor 1985). Since, in broad terms, such estimates of sedimentation rates can be equated with the global erosion rate, it is possible to derive estimates of global denudation rates in the past. This approach has been used by Tardy *et al.* (1989) to reconstruct the temporal pattern of global specific sediment yield over the past 500 million years presented in Fig. 2a. This indicates that global specific sediment yields have ranged between about $30\,t\,km^{-2}\,a^{-1}$ and $70\,t\,km^{-2}\,a^{-1}$ during the geological past. These variations reflect fluctuations in the global climate and particularly runoff amounts, as well as in relief, tectonic activity and vegetation cover. The Devonian period (*c.* 350–400 × 10^6 a BP) which is marked by relatively high sediment

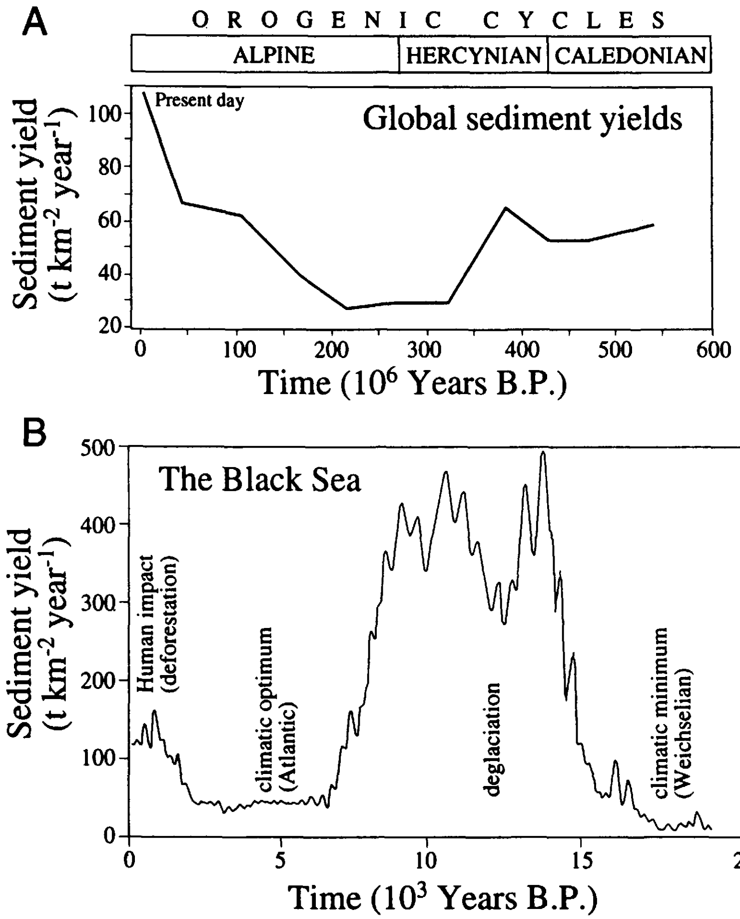

Fig. 2. Variations of sediment yield over geological time. Based (A) on Tardy *et al.* (1989) and (B) on Degens *et al.* (1991).

yields, is known to have been a particularly wet period with high runoff rates. The value of contemporary sediment yield (i.e. $108\,t\,km^{-2}\,a^{-1}$) depicted on Fig. 2a is based on river load data and is therefore not strictly comparable to those based on the mass of accumulated sediments. However, the overall trend of an increase towards the present would appear to be real and can be accounted for in terms of the increasing tectonic activity since the end of the Jurassic period (*c.* 130×10^6 a BP) and, more importantly, the impact of human activity including forest clearance and the development of agriculture.

Although involving a more restricted spatial and temporal perspective and a more detailed interpretation of the sedimentary record, Degens *et al.* (1976, 1991) have used a similar approach to reconstruct sediment inputs to the Black Sea over the past 20 000 years (Fig. 2b). The results indicate that sediment inputs were relatively low during the Weichselian glaciation, but increased dramatically during the subsequent period of deglaciation in response to the increased runoff and abundant supply of sediment. They then declined towards the Atlantic optimum, when a dense protective cover of vegetation would have existed. The subsequent increase in sediment yield during the past 2000 years has been related to the impact of land disturbance by human activity. The evidence provided by the long-term global perspective presented in Fig. 2a and by the sedimentary record from the Black Sea in Fig. 2b, underscores the importance of human activity in giving rise to increased sediment yields, but it also emphasizes the 'natural' variability of sediment yields in response to long-term environmental change and more particularly changes in climate and tectonic activity. The Black Sea evidence demonstrates that natural climatic variations, particularly those associated with periods of glaciation and deglaciation, may cause even greater changes than those caused by human activity.

Evidence from lake sediments

Where lakes occur at the outlet of a drainage basin and a substantial proportion of the sediment output is trapped, detailed analysis of the sedimentary record can provide valuable evidence concerning fluctuations in sediment loads during the recent past (e.g. 10^2–10^3 years). Where the catchment draining to the lake is relatively small, it may also be possible to relate the reconstructed sediment yield record to documentary evidence regarding land use changes and other human activities within the catchment (cf. Dearing *et al.* 1990). A classic example of the potential value of evidence of this type is provided by the work of Davis (1976) on Frains Lake in Michigan, USA. This lake is located at the outlet of a small 0.18 km^2 drainage basin and a detailed analysis of the sediment deposits enabled the sediment yield record for the past 200 years to be reconstructed (Fig. 3). This evidences low rates of sediment in pre-settlement times, rising by a factor of up to 70 with the onset of settlement and land clearance after 1830, and stabilizing after 1900 at a rate about ten times the pre-settlement level. In this case it is therefore possible to distinguish both the immediate

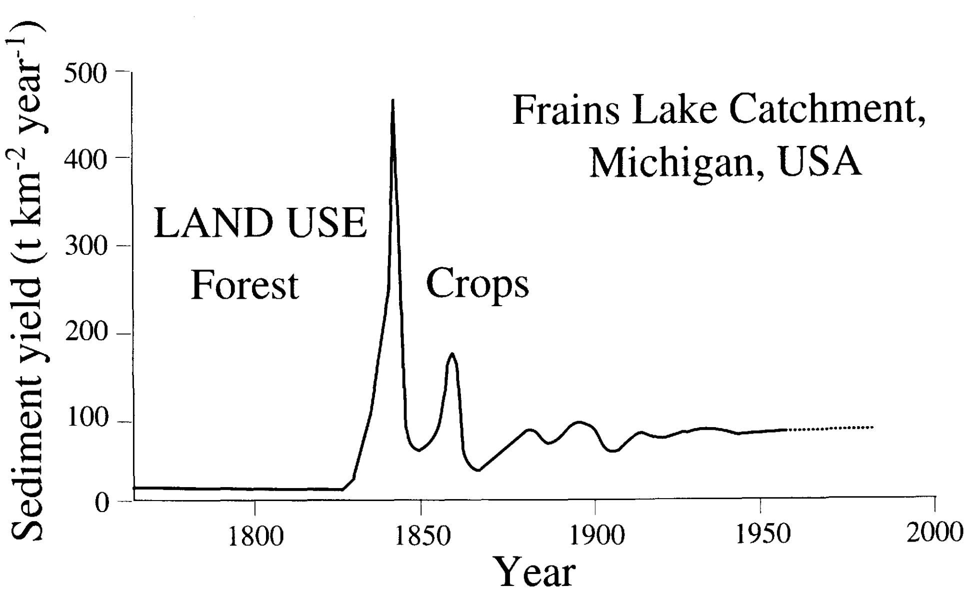

Fig. 3. Lake sediment-based evidence of historical trends in sediment yields to Frains Lake, Michigan, USA (based on Davis 1975.

Table 2. *Lake sediment-based evidence of increases in sediment flux due to catchment disturbance by human activity from tropical environments*

Lake	Location	Documented increase in sediment flux	Source
Lake Patzcuaro	Mexico	×7	O'Hara *et al.* (1993)
Lake Sacnao	Mexico	×35	Deevey *et al.* (1979)
Lake Ipea	Papua, New Guinea	×10	Oldfield *et al.* (1980)
Lac Azigza	Morocco	×5	Flower *et al.* (1989)

short-term impact of land clearance and the longer-term sustained increase in sediment yields. The findings of several other studies where lake sediments have been used to estimate the general increases in sediment yield from catchments in tropical regions associated with land disturbance are listed in Table 2. The trends evidenced in Fig. 3 and Table 2 and by the results of many other similar studies, again serve to emphasize the sensitivity of sediment yields to land use change in terms of both the magnitude of the increases involved, which may frequently exceed an order of magnitude or more, and the variations reflecting particular phases of human activity.

The evidence from long-term records

The general lack of reliable long-term records of sediment transport by the world's rivers and the absence of any concerted attempt to collate and synthesize such information, preclude a detailed consideration of the evidence provided by such records. However, two examples may be usefully introduced to demonstrate the nature of some of the evidence, the first from the former USSR and the second from the USA. Figure 4 based on data provided by Bobrovitskaya (pers. comm.), presents information on recent increases in the suspended sediment loads of four rivers in different regions of the former Soviet Union. The Dema river is a tributary of the Volga, the Kolyma river is located in western Siberia, the Dnestr river flows through the Ukraine to the Black Sea and the Yazgulem river is a tributary of the Amu Darya river in Kazakhstan. Each of the rivers shows evidence of an increase in suspended sediment yield of between 1.4 and 1.8 times since the 1960s. In the case of the Dema river, this increase can be related to the expansion of agricultural activity, whilst in the Dnestr river it can be linked to forest clearance. In the Kolyma river the increase reflects the effects of increased mining activity, and in the Yazgulem river it can be related to the expansion of irrigated agriculture. It is, however, important to take account of the possible impact of climatic change as well as the effects of changes in catchment condition, because both the Dema and the Dnestr rivers evidence a trend of increasing annual runoff over the period of record. In the Kolyma river, the annual runoff totals have remained stationary over this period, whilst in the Yazgulem river there is evidence of a slight decrease. It would therefore seem likely that the increased annual runoff in the Dema and Dnestr rivers has also contributed to the increased sediment yields associated with these rivers.

Whilst there are equivalent reports of increasing sediment loads in US rivers (Uri 1991, there are also many instances where the available records point to a marked

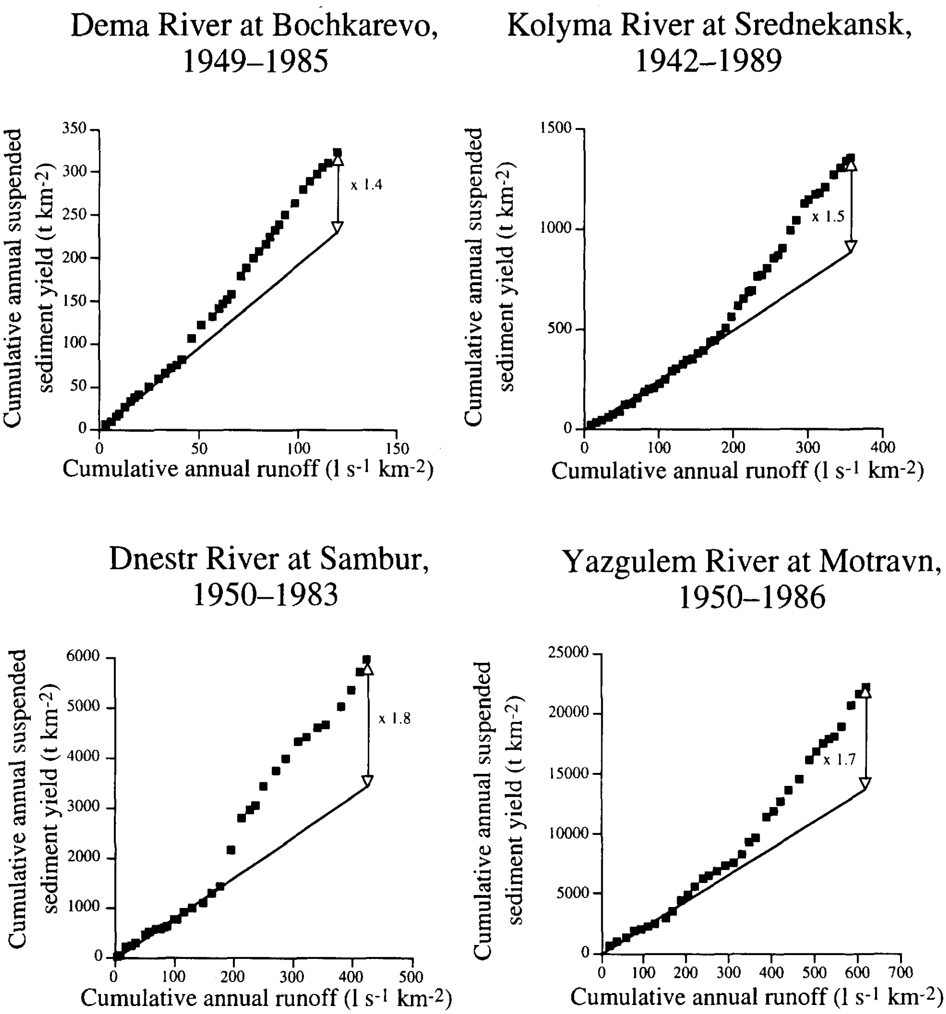

Fig. 4. Longer-term trends in suspended sediment transport by four rivers in the Former Soviet Union. (Basesd on data supplied by N. Bobrovitskaya, State Hydrological Institute, St Petersburg.)

reduction in sediment loads as a result of damming for reservoir development. Meade & Parker (1985) have, for example, documented how the construction of five major dams on the Missouri River between 1953 and 1963 caused a marked reduction in the sediment loads transported by this river, such that the load entering the Mississippi at its confluence was reduced to only about 25% of its former value. Since the Missouri River formerly represented the major supply of sediment to the Mississippi, the sediment load of that river has also declined, and the load at its mouth in 1984 was less than one-half of the value before 1953.

Evidence from catchment experiments

Most catchment experiments are, by design, concerned with relatively small areas and involve specific treatments or land use changes applied to the entire catchment

Table 3. *Some results from experimental basin studies of the impact of land use change on sediment yield*

Region	Land use change	Change in sediment yield	Reference
Increase			
Westland, New Zealand	Clearfelling	×8	O'Loughlin *et al.* (1980)
Oregon, USA	Clearfelling	×39	Fredriksen (1970)
Northern England	Afforestation (ditching & ploughing)	×100	Painter *et al.* (1974)
Texas, USA	Forest clearance and cultivation	×310	Chang *et al.* (1982)
Maryland, USA	Building construction	×126–375	Wolman & Schick (1967)
Decrease			
Loess Region, China	Soil and water conservation	97%	Mou (1991)
Loess Region, China	Soil and water conservation	93%	Li (1992)

area in order to monitor their impact (Walling 1979). The results obtained cannot generally be directly extrapolated to larger heterogeneous drainage basins representative of local or regional conditions, in order to examine longer-term trends in sediment yields. The results generated do, nevertheless, provide valuable information on the likely magnitude of the changes in sediment response associated with particular types of catchment disturbance or treatment and specific land use practices. The examples of *increases* in sediment yield documented by experimental catchment studies listed in Table 3 provide information of this type, and again emphasize the sensitivity of sediment yields to catchment disturbance. In these case studies increases of up to more than two orders of magnitude have been documented. Instances of *decreases* in sediment yield consequent upon the implementation of soil conservation measures are also listed in Table 3. In this case sediment yields have been reduced by *c.* 90% or more, but it is important to recognize that the catchment areas involved were small and that there is a general lack of definitive evidence regarding the potential of such measures to effect a significant reduction in the sediment yield of much larger river basins.

Application of space–time substitution

In the absence of reliable long-term sediment yield records providing evidence of trends associated with specific environmental changes, space–time substitution, or the ergodic hypothesis, can be used to demonstrate the likely trends involved. The basis of this approach is the assembling of information on spatial variations in sediment yield in response to variations in catchment condition (Fig. 5) and the use of this information to infer the likely magnitude of changes in sediment yield associated with known past changes in catchment condition. A classic example of the potential of this approach for reconstructing the long-term record of sediment yield for an area is provided by the study undertaken by Wolman (1967) in the Piedmont

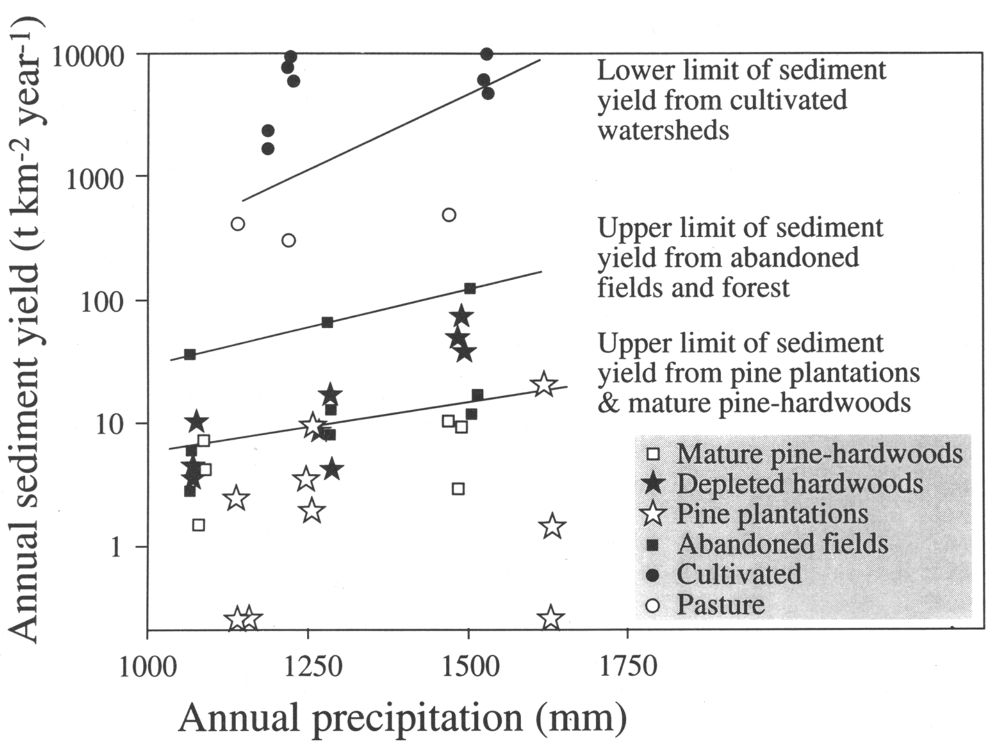

Fig. 5. The spatial variation of annual suspended sediment yields in Northern Mississippi, USA in response to land use activities (based on Ursic & Dendy 1965).

region of Maryland, USA, which synthesized the record of temporal changes in sediment yield depicted in Fig. 6. A somewhat different example of the application of this approach, in this case at the global scale, is provided by the work of the Soviet scientists Dedkov & Mozzherin (1984). They collated sediment yield data from more than 3600 world rivers and classified these into three groups according to the degree of disturbance by agricultural activity. By comparing the average sediment yield of the group of basins experiencing minimal disturbance with that of the most disturbed group, they suggested that the sediment yields of lowland river basins which had been heavily disturbed by agricultural activity have increased by *c.* 10 times, whereas for mountain rivers the equivalent value is *c.* 3 times.

Some problems of interpretation

In the preceding discussion of the various sources of evidence for the sensitivity of the suspended sediment loads of rivers to environmental change, a close and synchronous link between changing sediment loads and the forcing variables has been implicitly assumed. It is, nevertheless, important to recognize the complexity of the drainage basin sediment delivery system which may attenuate the link between changing erosion rates and sediment output (Robinson 1977; Walling 1983, 1989). An excellent example of this complexity is provided by the classic work of Trimble

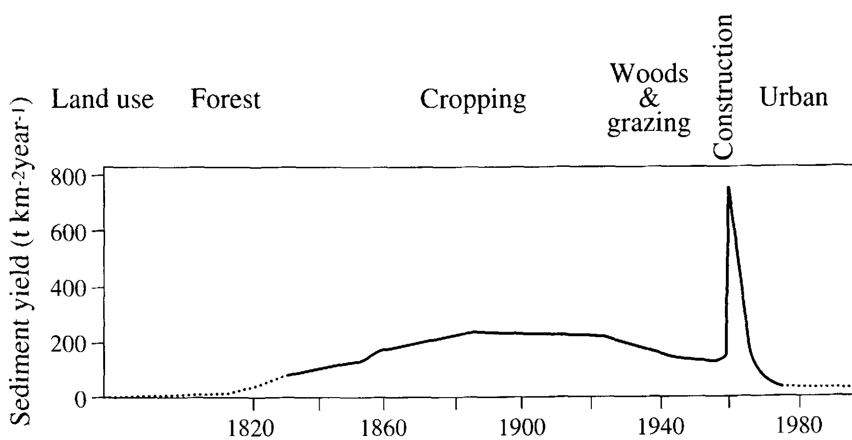

Fig. 6. The reconstruction of the historical record of suspended sediment yields in the Piedmont region of Maryland, USA, proposed by Wolman (1967) on the basis of space–time substitution.

(1976, 1981, 1983) who investigated the long-term response of the sediment budget of Coon Creek in the Driftless Area of Wisconsin, USA, to land use change during the period 1853–1975. Through this work he was able to reconstruct the sediment budget of the 360 km^2 basin for two periods, namely 1853–1938 and 1938–1975. The first represented a period of poor land management, which resulted in severe erosion, whereas the second was characterized by the introduction of conservation measures and substantially reduced erosion rates. The sediment yields estimated for the two periods were, however, almost identical at *c.* 110 and 115 $t\,km^{-2}a^{-1}$ respectively. During the first period large volumes of sediment were eroded from the slopes of the basin, but most of this was stored on the lower slopes and in the valley floors, and only a small proportion (*c.* 5%) was transported out of the basin. During the latter period, erosion rates in the upland areas were substantially reduced through the introduction of conservation measures, but the sediment yield at the basin outlet remained essentially the same due to remobilization of sediment stored in the valley floors.

When interpreting changes in yield from catchments it is also important to recognize the potential interactions between land use change and changes in climate, both in terms of isolating their respective impacts and possible synergistic effects. A useful example of the problems involved in separating the effects of catchment management and climate change on sediment yields is provided by the work of Zhao *et al.* (1992) in the 4161 km^2 basin of the Sanchuanhe River in the gullied-hilly loess region of the Middle Yellow River basin in China. Extensive soil and water conservation measures were implemented in this basin during the 1980s and a comparison of the mean annual sediment loads for the periods 1957–1969 and 1980–1989 indicated that sediment yields in the latter period had been reduced to only about 25% of those in the former. However, the soil and water conservation works undertaken in the basin were only partly responsible for the reduction. It also reflected a shift to drier conditions, with the mean annual precipitation falling from 542 mm in the first period to 509 mm in the latter. By using a range of data analysis

Table 4. *The relative importance of climatic variation and watershed management in accounting for the recent reduction in the sediment yield of the Sanchuanhe River, China*

Annual sediment yield 1957–1969	36.81×10^4 t.
Annual sediment yield 1980–1989	9.63×10^4 t.
Total reduction	74%
Reduction due to reduced rainfall	33–38%
Reduction due to watershed management	36–41%

Based on Zhao *et al.* (1992).

techniques, Zhao *et al.* (1992) were able to demonstrate that *c.* 50% of the reduction in sediment yield documented for the period 1980–1989 could be attributed to the soil and water conservation measures and *c.* 50% to the reduced annual precipitation (Table 4). It is likely that in some situations, changes in catchment condition and changes in climate may operate synergistically to produce changing sediment yields. For example, where land use change causes general degradation within a catchment leading to instability, it may require a shift in climate towards wetter conditions to trigger a major change in erosion and sediment yield. The impact of climate change may thus be important both in its own right and in providing a crucial forcing function during more general catchment disturbance.

Perspective

The preceding review of sources of information regarding the response of catchment sediment yields to environmental change has emphasized the sensitivity of this parameter to such changes. Evidence from the geological past emphasizes that sediment yields may exhibit substantial temporal variations even under natural conditions, in response to changes in climate and variations in tectonic activity and that the 'natural' background should itself be seen as exhibiting significant variations. The various sources of evidence available for more recent times indicate that increases in sediment yield of several-fold have been widely documented in drainage basins impacted by land use change and other forms of human-induced disturbance and in some cases such increases have been further augmented by the effects of climate change. It has been suggested that at the global scale, the sediment yields of rivers in lowland regions that have been heavily impacted by agricultural activities could have increased by an order of magnitude over those existing under undisturbed conditions. Such increases are likely to have been particularly marked in many areas of Asia and Oceania (Fig. 1) which account for a major proportion of the total transport of sediment from the land to the oceans and therefore exert an important control on the overall magnitude of this flux (Milliman & Syvitski 1992). However, it is important also to recognize that there are many instances where the sediment loads of rivers have declined, or at least remained essentially unchanged, in the recent past. In the case of the Sanchuanhe river in China cited above, climate change has been associated with reduced annual precipitation and this in turn has resulted in reduced sediment yields. There are also many examples of rivers where reservoir construction has caused reduced sediment loads.

In view of the important economic and environmental implications of increased sediment loads in rivers, it is important that their sensitivity to environmental conditions should be seen as a key component of the global change which is currently attracting widespread attention. Lack of reliable long-term records has precluded detailed assessments of the overall trend involved but there is clearly scope to collate and analyse existing data in more detail and to couple this with other sources of information such as those reviewed in this contribution. The increasing rate of degradation of the global soil resource and changes in the flux of materials between the land and the oceans are clearly important implications of this aspect of global change which would also benefit from an improved assessment of the magnitude of the changes involved.

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Human dimensions of palaeohydrological change

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Abstract: Because floodplains are 'attractors' of human activities and settlement, Late Pleistocene and Holocene palaeohydrological changes are an important part of archaeology. Indeed there has been a tradition of palaeohydrological writing within archaeology centring around topics such as river transport, drought/floods and floodplain farming, rivers as physical and cultural barriers, riverine resources, and the flood hazard. An agenda for continuing research in this inter-disciplinary area in a more quantitative and rigorous manner than in the past has been developed. This recognizes that the potential of palaeohydrological research to be applied to the study of past and future environmental change revolves around three areas: (i) the estimation of past frequency distributions and their relationships to climate; (ii) the past and future representations of risks; such as flooding; (iii) the uses and limitations of analogies (e.g. by using the Medieval climatic optimum as an analogy for future global warming).

Floodplains and human activity

This paper highlights some of the human dimensions of palaeohydrological change. Although human hydrological impacts appear relatively frequently, especially in work on historical hydrological change, there is far less written concerning human responses to climate change, yet human responses at least partially condition the environmental impact of further hydrological change. It is also hoped to illustrate this and the multi-disciplinary nature of palaeohydrology in this paper. Of particular relevance are intellectual links between palaeohydrology and archaeology, which have some longevity (Raikes 1967).

Hydrological change is change in both a resource and hazard, both in the past and in the future. One aspect of this change, namely floods, can be preserved in the geological–geomorphological record and so are particularly suited to investigation over a variety of timescales. Even in countries which have been collecting meterological–hydrological data for a century or more, it remains difficult to objectively assess flood risk even under the current climate (Bevan 1993). This is especially true of extreme floods which result from a combination of unusual storm characteristics and catchment conditions.

It is traditionally assumed that the occurrence of an event conforms to a probability distribution of some sort (Zelenhasic 1970); this is necessary in order to assign a frequency to any event, and normally involves the assumption that this distribution has not changed (stationarity). Palaeohydrological studies show this to be untrue. The period 1920–1960 in NW Europe was probably atypical with low inter-annual variation and the warmest mean temperatures for the last half millennia (Flohn & Fantechi 1984). The Francis (1973) division of the Thames flood series shows how stationarity is not even the case within this century.

However, frequency analysis of floods cannot be dispensed with because it is the only way at present to account for damage (or benefit) in economic terms where the costs or returns are to be spread both spatially and temporally. There is also strong empirical evidence derived from palaeohydrology that extreme events are bounded and that there are upper limits to flood magnitudes (Georgiadi 1979; Enzel *et al.* 1993, this volume). This is fundamentally because there is a limit to the amount of water that can be rained-out per unit time and translated into flow in a basin of finite size and saturated conditions. The extreme events series will vary with changing climatic and catchment conditions. An engineering approach to the problem of predicting extreme floods is through the probable maximum flood (PMF) which can be defined as 'that magnitude of rainfall over a particular basin which will yield the flood flow of which there is virtually no risk of being exceeded' (Myers 1969) and which is in theory governed by the structure of storms and antecedent catchment conditions. Calculation involves several steps:

(1) identification of maximum recorded rainfall for a specific catchment and comparable areas;
(2) transposition to the catchment concerned;
(3) upward adjustment of transposed rainfall values (maximisation) on the basis of meteorological conditions over the catchment area.

Statistically the PMF can be related to any flow distribution e.g.:

$$\mathrm{PMF} = Q + K\sigma$$

where Q is the mean flow and σ is the standard deviation of flow (or coefficient of variation) (ICE 1978). A modern value of K for rainfall in parts of the USA is 15 (Ward 1978). Transposition can involve pooling from a homogeneous area, such as that produced as part of the FREND project (Gustard *et al.* 1989). The recurrence interval conventionally given to the PMF is 1:10000 years, a figure that has no hydrological basis but is used in economic analysis. Other approximations include $3.5 \times 1{:}150$ year flood (Upland Britain), 2.5 times the normal maximum flood and some multiplier of the best estimate of the mean annual flood (BESMAF, NERC 1975). In reality the relationship between extreme floods and the mean annual flood is both spatially and temporally variable, indeed palaeohydrology is partly concerned with defining this relationship through the retrodiction of past hydrological regimes. The multipliers vary systematically from west to east across the UK and into Europe (Fig. 1). That the processes responsible for extreme events will change over time is the axial assumption of the nested frequency concept (NFC; Brown 1991; Bevan 1993; Fig. 2). Each frequency distribution which is characteristic of different climatic conditions has a tail that is related to the PMF. This is similar to the recognition of periods of drought or flood dominated regimes (Erskine & Warner 1988) and depends upon the subdivision of the record of climate change into relatively homogeneous periods with discrete characteristics; examples in northwest Europe might include: the medieval warm period, the Little Ice Age (Neoglacial) and the eighteenth century reversals associated with volcanic activity. Changing catchments conditions are also important and in theory this can be accounted for by changing the unit hydrograph used in the conversion of probable maximum

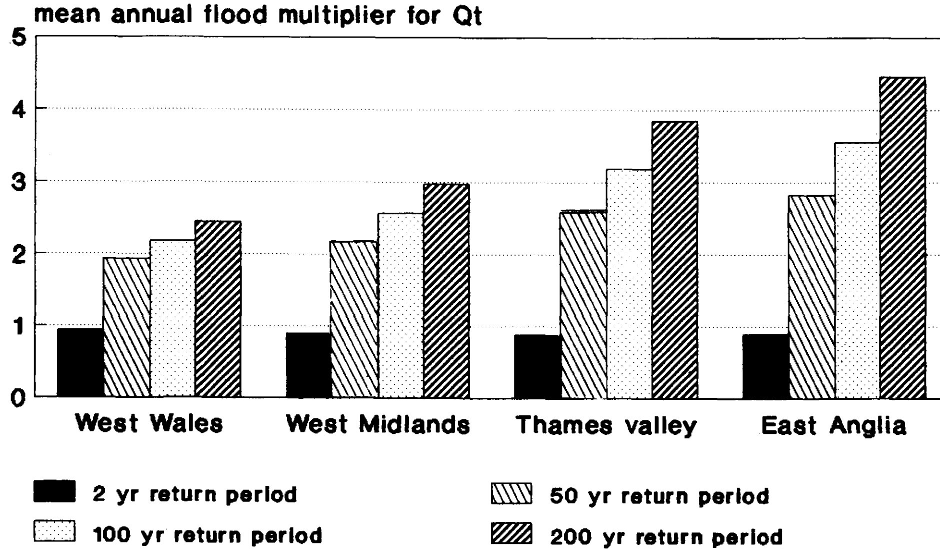

Fig. 1. The variation in flood multipliers from west to east across the UK. Data from IH (1987).

precipitation (PMP) to the PMF. This approach is a simplification of reality and in theoretical terms is something of a 'half-way house'. This is because in reality the distributions are constantly changing and overlapping, due to the multiplicity of causes of regional climate change. For example it would appear that the climate of NW Europe has over the last 1000 years been influenced by variations in solar output (solar forcing), changing patterns of typical synoptic conditions (particularly persistent periods of weakening and displacement of westerlies associated with temperature changes in the Northern Atlantic) and the influence of volcanic events. The first two factors may well be related but the lags and feedbacks involved mean that they may operate over different timescales and can be regarded as being quasi-independent at the regional scale. If these two factors are combined it becomes apparent that the division of the late Holocene, cold and dry, and warm and wet periods is unlikely and climate change may be both in phase and out of phase region to region (Table 1). When a truly independent factor is added the results are likely to be even more complex both decade to decade and spatially. The NFC provides one way of describing this complexity and translating it into human terms. The most recognizable regimes will be those most persistent and strongest; and probably those with the greatest seasonal contrasts. In a recent summary of the evidence for the medieval warm period (Hughes & Diaz 1994) illustrate how the existence of the medieval warm period and evidence of climatic cycles is regionally variable within NW Europe being out of phase with Fennoscandia and Southern Europe.

Palaeohydrology can help approximate these regimes/distributions through some estimate of mean annual flow and some estimate of the maximum flood series. Palaeomeander and palaeochannel analysis, and dendrohydrology (Jones *et al.* 1984) can be used for the former and palaeoflood analysis for the later. The most pertinent point is that one frequency distribution should not be used to assign a frequency to extreme events beyond the period of relative climatic stability, the alternative is to

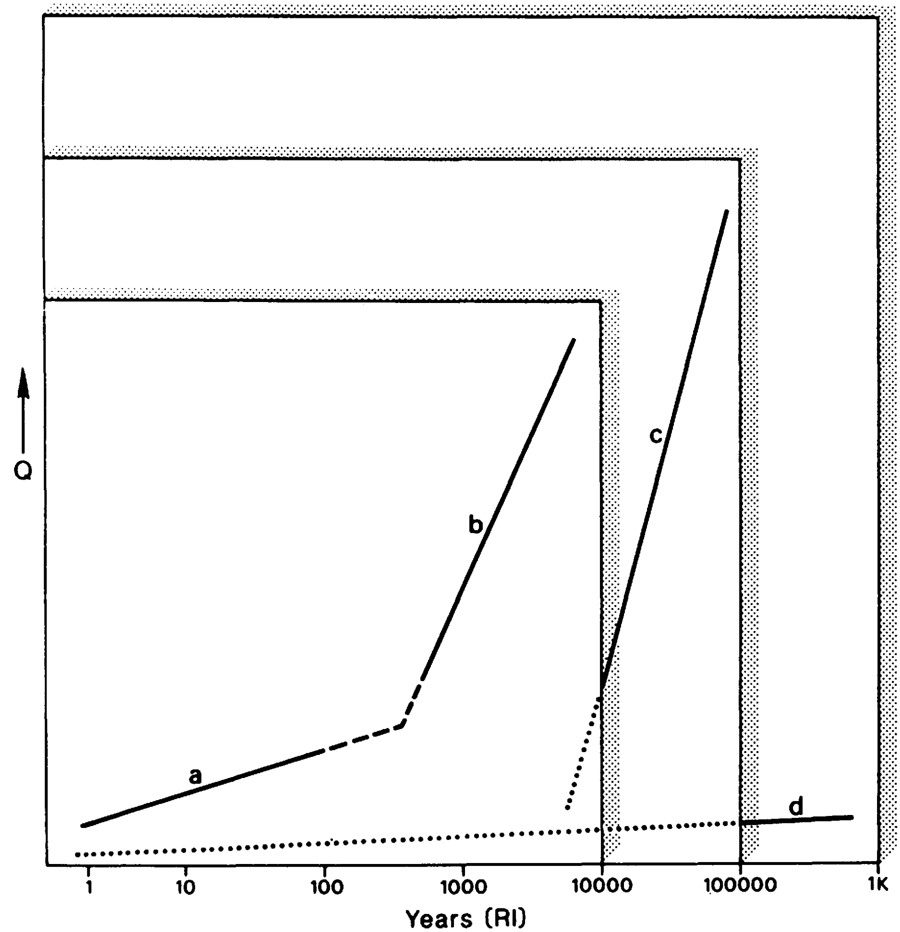

Fig. 2. The nested trequency concept. Slope a is the flood series produced by 'normal' interglacial maritime processes, e.g. westerly air masses crossing the UK, and floods are rainfall generated. Slope b is the flood series produced by a coincidence of rainfall and/or snowmelt (maximum controlled by maximum rainfall intensity + maximum rainfall rate). Some historical floods fall into this series, including the 1947 flood and probably floods that occurred in the 'Little Ice Age'. Slope c represents large floods caused by a sudden change in the thermal regime resulting in rapid melting of ice-fronts, glacial-lake bursts etc. Slope d represents rather low variance regimes produced by very cold continental conditions (unfavourable for surges or jokulahaups), low rainfall and limited summer melting. Reproduced with permission from John Wiley and Sons.

use palaeoflood series (Macklin *et al.* 1992) or palaeoflood estimates (Baker *et al.* 1990) which although highly satisfactory are limited to areas with suitable gorges and slack-water deposits.

There are other approaches which may be theoretically superior but they tend to have high data requirements, an example being fractal Gaussian noise models (FGNs) where the current probability of an event depends upon the history of the process or frequency distribution (Mandlbrot & Wallis 1969) or the broken-line method Mejia *et al.* (1972).

Table 1. *A conjectural illustration of how just two quasi-independent factors may have affected the regional climate of the NW European seaboard*

	Low solar output	High solar output
Strong Gulf Stream: westerlies prevailing over continental high pressure conditions – maritime regimes	Pre-medieval cold period type regimes and transitional regimes	Twentieth century regimes and transitional/ intermediate regimes
Weak or displaced Gulf Stream: continental high pressure systems dominant over westerlies – continental regimes	Little Ice Age-type regimes	Medieval warm period-type regimes

The human dimension of this hydrological change is that due to changing flood magnitude–frequency curves the probability of a large flood at certain times in the past was not what we would currently estimate it to be on the basis of measured flows. So the probability of a high flood during the Little Ice Age in the UK was probably much greater than it is today (due to higher frequency of snowmelt contributions). Furthermore, human responses depend not on the 'real' frequency of events but on what the frequency is perceived to be through scientific (post enlightment formal statistical analysis) or some other way of estimating environmental risk. Different responses to different flood environments are discussed below.

Floods in the Medieval climatic optimum: the Hemington bridges

Evidence from mountain glaciers, Greenland ice cores and tree rings suggest that in Europe between AD 700 and AD 1300 significantly warmer conditions persisted in summer, and possibly winter although this is less certain (Bell & Walker 1992; Hughes & Diaz 1994). The exact timing seems to vary within Europe and the cause is not firmly established although reduced solar input and increased volcanic dust seem most likely factors. Studies of a sensitive raised bog in Cumbria, NW England also shows that this was accompanied by a peak in bog surface dryness (Barber 1981). Despite this palaeoenvironmental evidence and historical records (Lamb 1982) there is archaeological and sedimentary evidence of floods during the twelfth and thirteenth centuries AD. There are several possible explanations; first that these records come from the climatic deterioration at the end of the period and that this is diachronic, secondly that the severity of winters has been underestimated, thirdly that the increased temperatures produced more mesoscale convective systems and a fourth possibility is that it is a catchment effect. The subsequent century (fourteenth century) is also known for frequent storminess in the North Sea (Gotteschalke 1971–1977) which has been linked to change in settlement in marginal areas (Brown *et al.* in press).

Recent sedimentary evidence comes from a gravel pit that has been operating at Hemington in the English Midlands which lies exactly opposite to the Derwent–Trent junction and where there is a dense pattern of palaeochannels (Fig. 3). This pit has produced valuable data on the construction of the Trent floodplain and channel change in this reach. The Trent is long known for its large floods and frequent

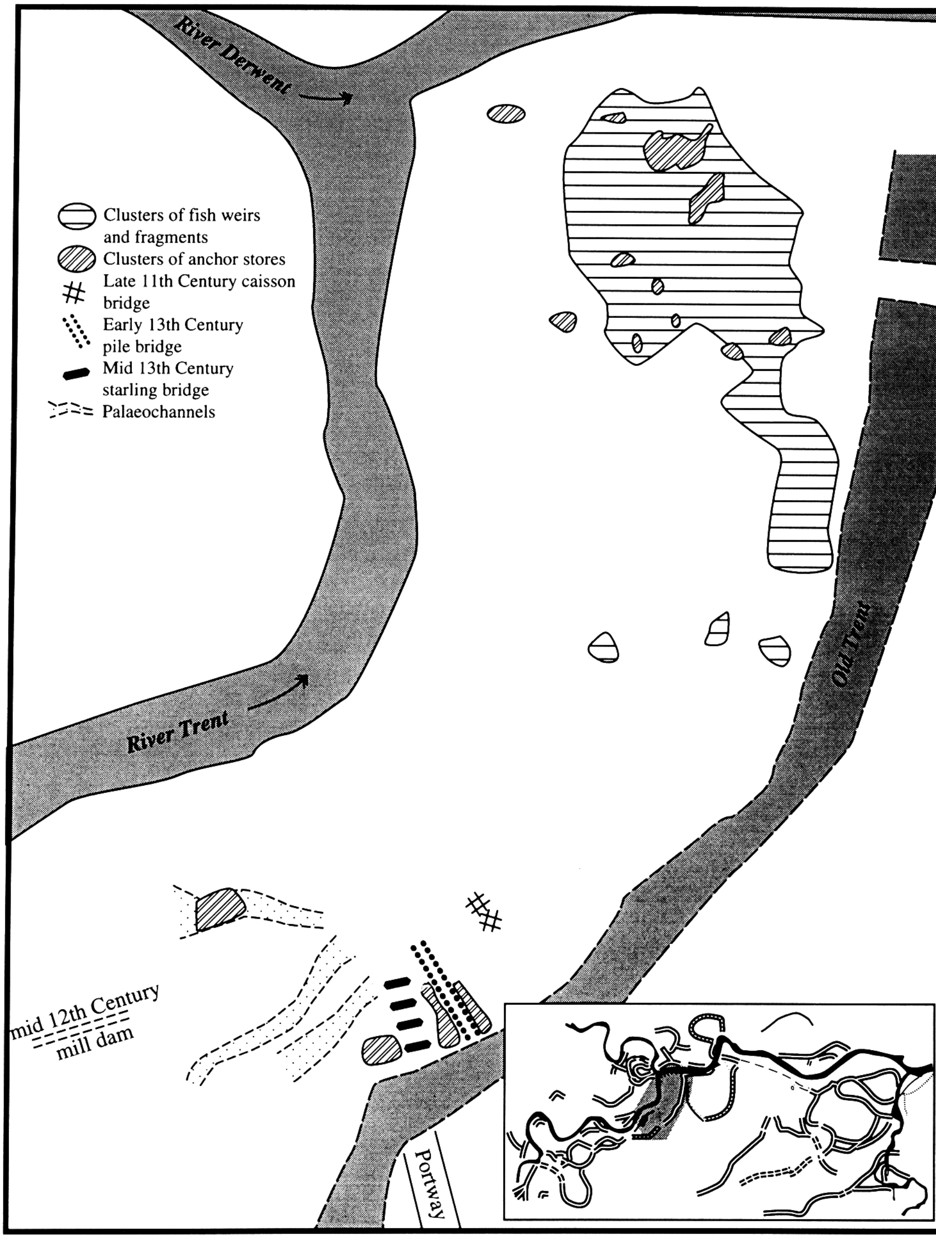

Fig. 3. The Hemington bridges. Plan with inset of the palaeochannels in the reach.

channel changes; in this respect it is unusual for lowland alluvial rivers in Britain. In the reach from the Derwent junction to downstream of Nottingham large areas of the floodplain have been reworked by meandering, braided and anastomosing channels during the late Holocene. At the classic site at Colwick black-oaks were deposited in a channel aligned with the flood flow (Salisbury *et al.* 1984), this is an unusual feature for lowland UK floodplains but common in this reach of the Trent and in Central Europe (Becker & Schirmer 1977; Kalicki & Krapiec 1991). Work by Salisbury over several years at Hemington has revealed the remains of a log boat, Neolithic wood and many fish weirs and anchor stones (Salisbury 1992). Excavations in 1993 revealed the remains of three bridges crossing a medieval palaeochannel. The bridges are of different designs, the earliest (eleventh century) being supported on diamond-shaped caissons and displaying typical Anglo-Saxon carpentry. The early thirteenth century bridge was supported by wooden piers and the mid thirteenth century bridge, which was *c.* 40 m upstream, was supported by stone piers. The eleventh century bridge seems to have failed due to scour around and between one the caissons which created a scour pool into which one of the caissons slipped. Whilst the date of this is not as yet known it must predate the early thirteenth century, likewise it is likely that the failure of the early thirteenth century bridge will predate the mid thirteenth century. The date of failure of the mid thirteenth century bridge is not known; however, there was a well-documented flood in 1403 which caused channel change isolating a pasture at Wilne. By 1740 the channel had reached the location of the Old Trent which was itself abandoned by the nineteenth century.

The stratigraphy of the site is complex with large tabular and shallow cross-bedded units of gravels underlying up to 2 m of silt–clay. The upper gravels are medieval in age, but because they are reworked Devensian gravels there is little or no sedimentological difference between these and the Lower Devensian gravels *in situ.* Recent radiocarbon dating has shown that these basal gravels date from the Younger Dryas between *c.* 11 700 years BP and 10 000 years BP. In practice the only way of distinguishing the Medieval from the Devensian is by colour (the red iron staining of the Devensian gravels is replaced by yellow-brown staining in the medieval gravel) and by the presence of artifacts. From the ninth century onwards all the evidence points to one, or more probably two, shallow and wide gravel-bedded channels prone to move and fill with gravel during floods. The relative sizes of the bridges also suggests a progressive shift of the flow from the western side of the floodplain to the eastern side between the tenth and fourteenth centuries. The lack of systematic preservation of earlier floodplain makes it difficult to tell if this was a distinctive phase or channel behaviour typical earlier in the Holocene.

The reason for the sensitivity and atypical response of this lowland river, in Western maritime terms, is its unusual subcatchment input. Within only 40 km it receives discharge from four large tributaries, the Tame, the Dove, the Derwent which drains the southern Peak District and the Soar which drains lowland Leicestershire. The Derwent which enters at Hemington is hydrologically quite unlike the Trent having a far greater flood yield per area and higher peak flood in relation to the mean annual flood (Table 2). It is a major contributor of discharge during floods and on average it contributes 35% of the combined discharge. However, in occasional floods the Derwent may contribute far more, a good example of this was the flood on 10 December 1965 when the Derwent contributed 60% to one of the largest floods on record. These events are very likely to have had significant geomorphological impacts at the junction and would have caused high overbank velocities and shear stresses in the Hemington reach.

The human response to the problems created by these large floods was technological involving different designs and different locations of bridges. Bridge collapse induced by pier scour associated with unusually large floods has been recognized by British Rail as a potential problem associated with climatic change.

Table 2. *Flood Discharge Estimates for Trent Sub-catchments*

Catchment/Station	n (yrs)	Area (km^2)	Max. recorded flood/area (m^3 s^{-1} km^2)	BESMAF/area (m^3 s^{-1} km^2)	Flood multipliers (mean Q to max. rec. Q)
Trent: Yoxall	10	1230	0.10	0.05	1.7
Trent: Drakelow Park	6	3070	0.08	0.05	1.4
Trent: Shardlow	14	4400	0.09	0.06	1.6
Derwent: Longbridge	33	1120	0.46	0.14	3.2
Soar: Zouch	–	1290	–	0.03	–
Trent: Trentbridge	82	7490	0.15	0.06	2.2
Trent: Colwick	11	7510	0.11	0.07	1.8
Thames: Teddington	39	9870	0.11	0.03	3.2
Severn: Montfort	31	2030	0.13	0.06	2.4
Severn: Bewdley	32	4330	0.21	0.11	1.7

Data from NERC (1975).

Floods in the late eighteenth century

The late eighteenth century is a pivotal period for palaeohydrological studies, as in many parts of the world this is when systematic recording of natural phenomena really began. It also lies at the end of the last neoglacial cycle ('the Little Ice Age') in the northern hemisphere. From this period we can start to make both inter-basin comparisons and quantitative reconstructions of flow regime and flood statistics from written records. Earlier flood lists which although they provide records of flood frequency can rarely be converted into discharge and normally only exist for just a few locations; examples are the excellent records from the Nile (Evans 1990), Florence (Losaco 1967) and Rome (D'Onofrio 1980). The late eighteenth century in

Table 3. *Recorded floods from the late eighteenth and early nineteenth centuries*

River	Date	Est. Q (m^3 s^{-1})	Comments
Trent, Nottingham	16.2.1795	1416	Largest in over 300 years
Severn; Shrewsbury	4.2.1795	–	Largest in 322 years, exceeded downstream of Worcester by the 18.11.1770 flood
Stour; Kidderminster	–.2.1795	–	Largest on record
Wye; Belmont	11.2.1775	1080	Largest recorded
Thames; Teddington	10.3.1774 27.1.1809	– –	Larger than highest in recorded series (1884), largest in eighteenth century, largest on record
Tyne/Tees/Wear	16.11.1771	*c.* 4000	Largest on record and probably largest in UK in the last 300 years
Exe; Stoodleigh	–	–	Largest on record

Data from several sources.

NW Europe was characterized by rising temperatures, but they were as spatially variable and erratic year to year as had been the most intense period of the Little Ice Age (Lamb 1982). There is considerable evidence that, in England, seasonality was more extreme than today with warmer summers and harsher (colder and wetter) winters and this would have altered the slope of the flood frequency curve. This reversal in the climatic recovery of the eighteenth and nineteenth centuries is generally ascribed to an extraordinary frequency of explosive volcanic eruptions (Lamb 1982). Many of the largest and earliest quantifiable floods in the UK took place between 1770 and 1800.

The winter of 1794–1795 is especially important in the UK as it was extremely severe and produced widespread flooding (Table 3). Rainfall in February produced

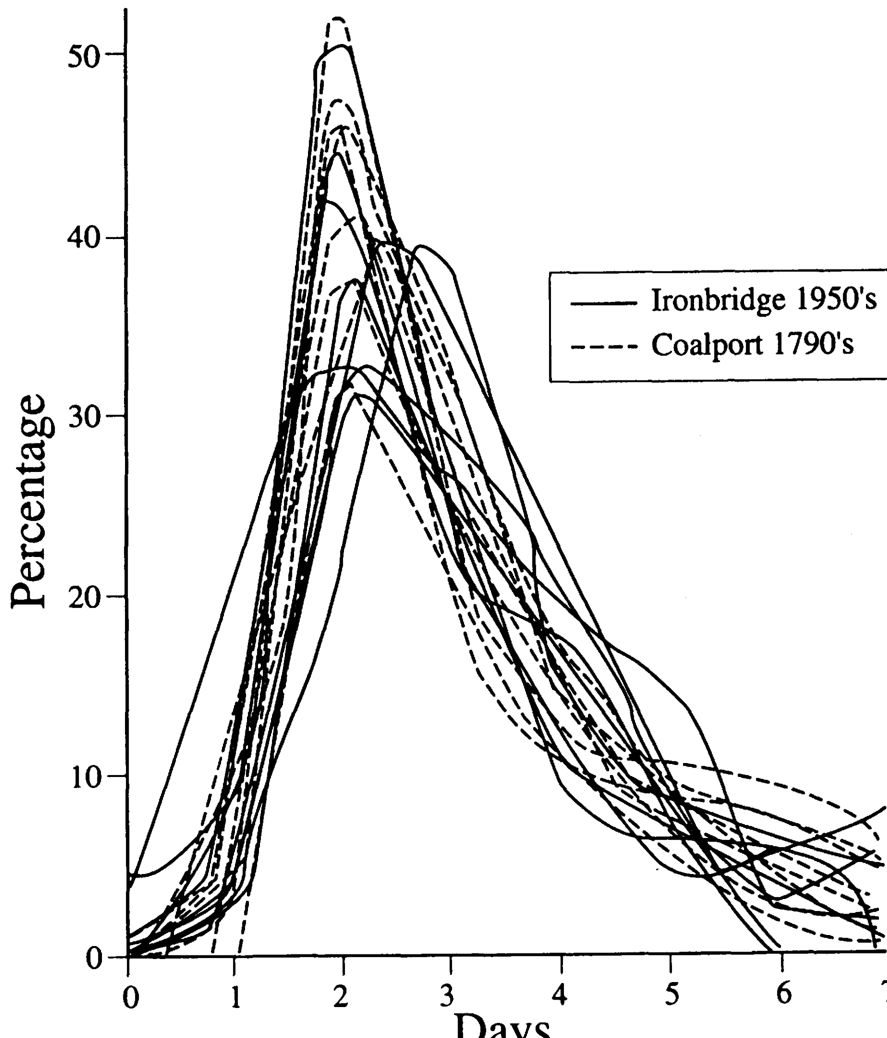

Fig. 4. Reconstructed late eighteenth century flood hydrographs from the River Severn at Coalport. Data collected by Telford and published in Plymley (1803).

the largest floods on record for the Trent, Upper Severn, Wye, Stour (Worcestershire) and Ouse (Yorkshire). Conventional analysis would give them a recurrence interval of between 199 and 350 years and yet in the late eighteenth century they were much more normal than this suggests and again in the late nineteenth century very similar flood levels were repeatedly reached and in some cases exceeded. The highest floods on record for most other UK catchments also occurred between 1770 and 1810.

One of the responses to these floods was the start of daily stage monitoring of the Severn at Coalport by Telford. Although the stage board has been lost the approximate location is known and as the section is in bedrock relatively little channel change can have occurred since the late eighteenth century allowing the comparison with more recent stage records from the nearest stage gauge at Ironbridge. The Coalport series shows smaller in-bank floods (Fig. 4) of similar dimensions and frequency to recent in-bank floods although less peaked (due to less drainage). If this is the case given the record of extreme events during the period then the flood frequency–magnitude curve must have been steeper than it was in the 1950s–1960s. The series also reveals that the Severn river froze over for an average of 11 days per year during this period. Another response to the floods in this area was to initiate a debate over the cause, with some arguing that land use change and the construction of embankments was responsible whilst other maintained they were simply part of natural variability. The systematic construction of engineering works to mitigate the riverine flood hazard date from this period.

Human response: a nineteenth century example, Gundagi, NSW

The abandonment of settlements, because of flooding in the historical period is extremely rare, even in the case of catastrophic losses of human life. A revealing exception is found during the colonization process, when the settlers have little or no experience of the environment in which they are now living. A rather vivid example of the mis-location of settlements by pioneers during colonization is the town of Gundagai, New South Wales, Australia. The town stands at the point where the floodplain of the Murrumbidgee widens to approximately 1 km and leaves the mountains. The floodplain was settled at this location, because it was the crossing point for explorers and travellers moving west into the Riverine Plain. Due to the fact that the river was frequently too high to be crossed, traders set-up on the banks providing services to the travellers' camps. Buildings were constructed on the floodplain, and the town was gazetted and surveyed in 1838. There are historical records (Crooks 1986) of frequent warnings given to the Europeans by the local aborigines that the location was not safe. The town was flooded to a depth of 1 m in 1844 and again in 1851, the response being to construct upper storeys presumed above any flood height. Then, on 25 June 1852, the town was devastated by a flood 4–5 m deep which killed 89 people (over 30% of the population) and destroyed 71 buildings. The buildings were smashed by the flow and particularly floating tree trunks. This remains Australia's worst flood disaster and it would have been even worse but for an Aboriginal named Yarri who saved many lives during the flood using a bark canoe. After an even greater flood in 1853, the present town was rebuilt on the valley slopes to the south and north of the floodplain. The lowest street of the

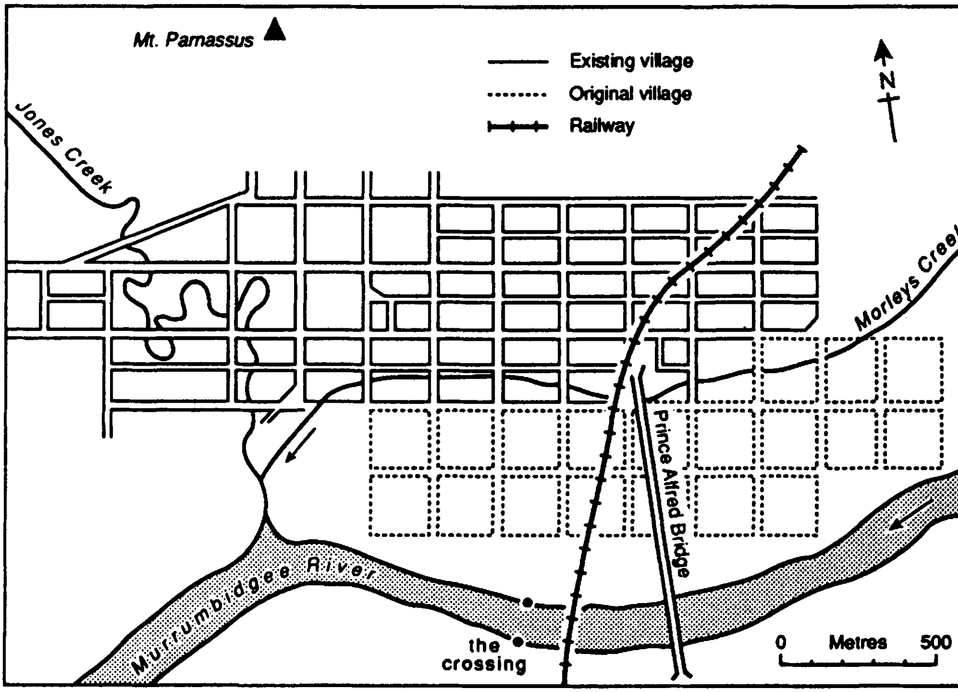

Fig. 5. The Gundagi town plan, before and after the 1852 flood. Adapted from Crooks (1986).

new north town was the highest street of the old town (Fig. 5). The floodplain where the original town was located is now crossed by a unique metal bridge completed in 1867 and a railway viaduct 921 m long built in 1901. The floodplain form does give some indication of the power of overbank flow in this environment, as it is uneven with several flood channels, scour depressions and flood levees. The name Murrumbidgee is also derived from the aboriginal word Morambeedja which means 'one big water' or 'mother of waters'.

This sort of flood disaster is unusual and related fundamentally to cultural factors, such as colonialism and unregulated development.

Integrating palaeohydrological change and risk: human response

In human terms hydrological risk consists of two components; the real likelihood and the perceived likelihood (the PMF has elements of both). The resource attractions of a riverine location can clearly outweigh perceived risks and/or the perceived risks were based upon a period with different hydrological conditions. So perceived risk will not necessarily change along with real risk. Humans display different risk behaviour based largely on socio-cultural factors (Royal Society 1992). The way humans have perceived risk has changed; if 'we' have a probability based perception (even if unformalised) others, even today, have a fatalistic perception. The result in terms of adaptation or other responses would vary markedly. A rational economic view would be that different risk strategies would be related to different levels of real risk and that therefore these would change as the risk changed

(Fig. 6). This can be used to compute pay-off matrices (Table 4) and identify areas sensitive to changes in flood magnitude and frequency (Arnell this volume). In this rational model floods are accepted and adapted to *in situ* (subject to technological constraints) as long as the combination of costs of moving elsewhere and opportunity returns of the riverside location exceed costs attributable to damage, rebuilding and flood defense. This can be refined by allowing for different utilities:

(a) expected utility, on the basis of all expected outcomes;
(b) subjective utility, on the basis of a subjective view of the probability of expected outcomes;
(c) bounded rational choice, on the basis of something less than the goal of maximizing returns.

Table 4 also illustrates how the outcome depends on access to information, the ability to process it, the values of the individual or society and upon social structure. The reasons for the failure of rational economic models to predict hazard response are illuminating for both explaining the past and predicting the future:

(a) information on the probability distribution of flooding is never complete or entirely accurate;

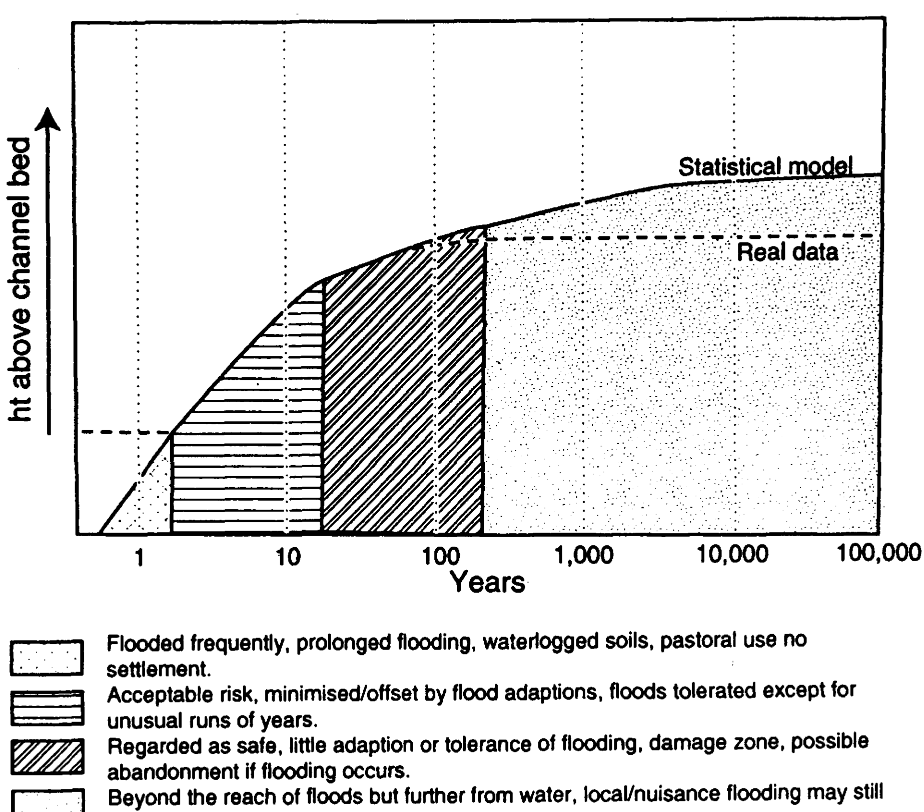

Fig. 6. Hypothetical floodplain risk in relation to flood magnitude and frequency.

Table 4. *Pay-off matrices for flood: expected utility and known probability*

	Flood breaches the levee	Flood fails to breach the levee	Expected utility
Evacuate	Pay for evacuation $0.4 \times 1 = 0.4$	Pay for evacuation $0.6 \times 1 = 0.6$	1.0
Remain	Lose all $0.4 \times 0 = 0$	No loss $0.6 \times 2 = 1.2$	1.2

Adapted from Burton *et al.* (1978) and Slovic *et al.* (1974).

(b) the opportunity costs of floodplain abandonment are not easily estimated and will change if some abandonment occurs;
(c) optimism, or extreme risk taking behaviour;
(d) transfer of the costs of the location to another body e.g. landlord or state;
(e) inability to move due to political control;
(f) fatalistic beliefs.

For these and/or other, perhaps individual reasons, it is bounded rationality that is most likely in most situations as studies of flood response show (Kates 1962). Contemporary observations and archaeological work, such as that in the Mississippi (Steponaitis 1978) and Midland England (Brown in press) suggests that coping with persistent flooding is relatively easy through adaptation (e.g. trackways, stilts, flood defenses, etc.) and it may be a catalyst for innovation. In theory it should be the margins that are sensitive the change, indeed this underlies the recent call for a sensitivity analysis of flood zonation maps in the UK to scenarios of greenhouse gas associated climate change (Arnell *et al.* 1994).

Questions for applied palaeohydrological research

This brief paper has attempted to outline how palaeohydrological research can, or could, be used in both studies of past and future human response to environmental change. In summary its potential revolves around three areas.

(1) The estimation of past frequency distributions and their relationship to climate

An alternative way of defining K is by using a temporally constrained sample. If we assume that the recurrence interval of an event cannot exceed the duration over which conditions that typically produce that event persist (e.g. a decade or 50 years) then either the recorded flood transposed from anywhere within the homogeneous area or an estimated probable maximum flood can be used to calculate the maximum K value. Changes in the value of K will then mark climatic change (and should correlate with forcing factors) and it is the distribution of K upon which design criteria should be based. This implies far greater attention to the periodicity of climatic fluctuations and to Hurst-type phenomena (Mandlebrot & Wallis 1968) and land–ocean interactions.

(2) The past and future representations of risks

There are two components to this. Firstly there needs to be a sensitivity analysis of the effect of likely changes in BESMAF, the mean flow (Arnell *et al.* 1994) or preferably changes in both mean flow and in the flood frequency curve over the areas of floodplain inundation. This may well result in the revision of floodplain inundation maps which are produced in the UK by the National Rivers Authority (now the Environment Agency) for planning authorities. Palaeohydrological estimates of these parameters as well as those predicted by hydrometeorological modelling should be used. Secondly the way flood (or drought) risks are communicated to the public needs to be re-evaluated.

(3) The use of analogies in palaeohydrology: the medieval climatic optimum – a possible analogy for global warming in NW Europe?

Whilst there can be no atmospheric analogies for greenhouse gas-induced warming the processes of discharge generation will not change. In temperature terms the Medieval warm period may superficially approximate the scenarios for global warming within the next 30 years, i.e. an increase in mean annual temperature of 1–2°C but due to a different combination of external factors (solar forcing, ocean–atmosphere conditions) it is unlikely that the seasonal contrasts will be the same and natural quasi-cyclic fluctuations will also modify anthropogenic forcing. The late Medieval was also a period in which land-use was not dissimilar to that today. By AD 1400 there was probably slightly more land under the plough in lowland England than today and arable agriculture was practised at higher elevations than today (Parry 1978). Although the period has been used as a climatic analogy it has not been used as a palaeohydrological analogy yet the data may be available to do this (Rumsby & Macklin, this volume). Of particular importance is the relationship between higher temperatures and convective storm activity.

Palaeohydrology allows the geological record to be the laboratory in order to generate and test models of environmental change and human response, which by necessity will often be part deterministic and part probableistic. Archaeological and historical data have a two-fold value here; first in providing detailed and well dated evidence of hydrogeomorphological changes and secondly by illustrating the complexity of societal response to changing environmental conditions. If palaeohydrological work can reveal unsuspected processes or response pathways then it will be of direct use in planning for future environmental change.

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Discovering Earth's future in its past: palaeohydrology and global environmental change

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Abstract: Palaeohydrology is the study of various past aspects of the hydrological cycle. It is accomplished by interpreting indices of past hydrological processes. By its focus on the effects of *realized* processes, palaeohydrology provides a critical scientific complement to studies that *idealize* hydrological systems in order to predict future change. Future habitability of the planet will surely require the guidance of scientific concepts, but it will also require a basis in the perception of change that compels people to act. Human perception is grounded in the concrete particulars of reality, not in the abstract idealizations of theoretical science. While we scientists may wish otherwise, say for a more enlightened public and political sector, it is pragmatic not to expect such on the short time scale of potential detrimental changes for Earth's habitability to humankind.

The greatest repository of scientific knowledge for stimulating human perception is the geological past. This experience of the Earth is the only real evidence of global environmental change available. It can function scientifically as a source of data with which to test models of change, confirming or falsifying theoretical extrapolations to past states (retrodictions). However, its more important role is as a source of discovery. In the geological record the scientist can discover previously unanticipated realizations of hydrological processes that will require new or revised models (hypotheses) for their explanation. It is our challenge as scientists to devote as much effort to exploring this real world of Earth experience as we currently apply to idealizing the abstract world of Earth systems.

Human civilization has long been at risk from environmental change. Droughts, floods, and soil degradation have all imperiled the infrastructure of past societies. Modern concerns about 'global change' have focused on issues such as greenhouse warming, ozone depletion, desertification, and deforestation. International programmes for 'Earth-system science' and 'geosphere–biosphere' study are being actively promoted for achieving solutions to these problems. In the dozen years since their inception (Perry 1991) these programmes have moved to the forefront of concern by the international science community. For the United States research strategies have been established to achieve an understanding of global change (Committee on Global Change 1990). The primary strategy of 'integrated modeling of the Earth system' is being implemented into policy for the funding of US science. The goal of the 1994 US Global Change Research Program (Committee on Earth Sciences 1993) is 'to produce predictive understanding of the Earth system to support national and international policy-making activities across a broad spectrum of global, national, and regional environmental issues'.

The prospects of global environmental change adverse to human habitability of the planet pose an immense challenge to science. Not only must good science be done, but that science must effectively translate into public policy. The latter involves a process in which the analytical procedures of science necessarily play a limited role (Lindblom & Woodhouse 1993). Much as scientists may wish otherwise, say for a more enlightened public and political sector, action on important societal issues is most commonly driven by immediate perceptions. There is only a limited role for long-term scientific study, for which inadequate resources of time and financial support are available.

Palaeohydrology is the study of past aspects of the hydrological cycle (e.g. Baker 1991). Clearly water plays a critical role in environmental change. However, the change that concerns humanity is in the future, not the past. Palaeohydrologists interpret various indices of past hydrological processes. How can such information help in coping with an uncertain, potentially perilous future? Should not the relevant science for society be focused on that future? The style of science advocated for studying global environmental change necessarily defines a role for studies of the past. That role is stated explicitly in the 'Earth System History and Modeling' section of the US strategy document (Committee on Global Change 1990). The past provides (1) a *data set* for validating the response of climate models, or it provides (2) *information* on the Earth system in terms of the coupling of its components or the responses of its subsystems. Such a role, while central to the mathematically predictive scientific philosophy being promoted for global environmental research, is inimical to the spirit of the naturalistic/historical scientific philosophy out of which sciences of Earth's past derive (Baker 1994*a*).

Two styles of environmental science

The Earth-system science of global environmental change seeks a predictive understanding of the future. In seeking this path it makes a value choice, favouring a mathematical/predictive approach over a naturalistic/historical one. Any such value choice will result in a kind of hierarchy in the sciences. For this value choice predictive and experimental sciences occupy top positions in the hierarchy while the mathematically less sophisticated historical and descriptive sciences fill in the low positions (e.g. Alvarez 1991). Of course, such classifications are arbitrary and nontestable; they are philosophical, not scientific. Nevertheless, they are assumed by many scientists, so they become very important in the practice of science. The assumption of philosophical ascendancy for the mathematical/predictive approach results in a rather specific manner by which palaeohydrology becomes incorporated into predictive earth-system science of global environmental change. For this reason the comparison of scientific approaches requires careful attention (Table 1).

Mathematical/predictive sciences are best exemplified in the experimental/theoretical methodology of classical physics. Indeed much of philosophy of science is written as though the words 'physics' and 'science' are interchangeable. Physics is 'the science devoted to discovering, developing and refining those aspects of reality that are amenable to mathematical analysis' (Ziman 1978). Its approach is conceptual, seeking universal classes of phenomena that can be generalized by means of the underlying physical laws presumed to govern nature. Its abstract laws, theories, and relationships must be objectively verified or tested against measured reality through controlled experimentation. As Sir Francis Bacon noted, scientific experiments are

Table 1. *Comparison of scientific styles for studying gobal environmental change*

	Mathematical/predictive (experimental/theoretical)	Naturalistic/historical (experiential/observational)
Basis	Define elements of nature (systems) capable of controlled study	Take the world (nature) 'as it is'
Goal	Develop theories that explain the world	Develop understanding of the world
Emphasis	Idealizations: general principles presumed to apply at all times to all places	Real phenomena: concrete particular happenings, especially in the past
Method	Controlled experimentation (in the lab) to justify knowledge	Observation (in the field) to stimulate hypotheses
Tools for study	Facts	Theories
Role of 'data'	Verification (validation) of models (predictions); theory confirmation	Signs that provide a 'conversation with the Earth'
Inference	Deductive analysis (rigorous and elegant) and inductive synthesis (for theory confirmation)	Retroductive (abductive) synthesis followed by reality-based deduction and induction
Policy relevance	Conceptual guidance for societal action	Perceptual basis for initiating potentially productive action

questions put to nature. However, the need for objective control means that questioning occurs as an interrogation (Keller 1985) in which the facts of nature are expressed through numerical measures comparable to those generated by the theoretical representation of its essential underlying laws. In this context palaeohydrology furnishes factual data that exemplify environmental change. Comparison of these realized results (completed 'natural experiments') to the theoretical predictions results in scientific validation of theories that are expressed through predictive models.

The naturalistic/historical sciences do not focus on idealized theories verified in experimental laboratories. Rather, their prime concern is with realized phenomena observed in the natural world, uncontrolled by artificial constraints. By not limiting herself to the world amenable to mathematical analysis, the naturalistic scientist takes the world as it is. Rather than general principles of universal application, it is concrete particulars that are the focus of attention. The richest source of such reality is the various evidence of happenings in the past. Hence, the naturalistic sciences merge with the historical. Data do not serve the interrogative function of experiments designed to verify. Instead, the observations of past phenomena are revealed as signs, providing a language for what Cloos (1953) termed a 'conversation with the Earth.' The observations do not serve primarily in the menial function of model validation. Rather, they provide inspiration for hypotheses, as classically described by Gilbert (1886, 1896). Hypothetical reasoning, therefore, provides a kind of logic, or inference, through which one can distinguish the naturalistic/historical sciences (e.g. Chamberlin 1890) from the mathematical/predictive.

Palaeohydrological inference

The reasoning process of palaeohydrology has been characterized as a kind of 'think back' in which specific phenomena (e.g. river sediments, palaeochannel dimensions) are viewed as the results of causative processes (e.g. river mechanics and hydrology), which, in turn, derive from the general controls of climate, lithology, structure, and tectonics (Baker 1974, 1978*a*, 1983; see also Allen 1977; Gregory 1987; Richards 1987). Reasoning from general to the specific is often labelled 'deductive', while the converse is considered 'inductive'. Thus, inferences made from specific observations, including the results of palaeohydrological phenomena, commonly receive the label 'inductive'. Recent work on inference in geology (Von Engelhard & Zimmermann 1988) reveals this characterization to be overly simplistic. Induction infers general principles of theoretical concepts from specific events or occurrences. The inference of cause from effect, or 'consequent' from 'antecedent' as Gilbert (1886) described it, is properly not induction. This mode of inference is central to all geology and is the principal logic applied in palaeohydrology. It has been accorded the formal labels 'retroduction' and 'abduction' by the American logician Charles Peirce, who traced the latter term to use by Aristotle and specifically associated it with geology (Baker in press). While retroduction/abduction is clearly important in geological geomorphology (Baker & Twidale 1991; Rhoads & Thorn 1993), nevertheless, it might appear to be an issue of obscure logic. The following example will illustrate the relevance of this and other reasoning modes to understanding hydrological phenomena.

Cataclysmic flooding from the failure of ice-dammed lakes is now recognized to be a hydrological process of global significance. Such flooding was extensively associated with the late Pleistocene deglaciation of northern continents (Baker 1994*b*). However, the first recognition of this flooding (Bretz 1923) was received with immense skepticism because of its outrageous character relative to prevailing theories of river behaviour (Baker 1978*b*). The path to eventual hypothesis acceptance involved several decades during which the compelling reality of field evidence eventually overcame the restrictive influence of prevailing theories upon the scientific community (Baker & Bunker 1985).

Suppose one wishes to study cataclysmic glacial flooding via a deductive approach. Never mind, for the moment, the long history whereby this idea even

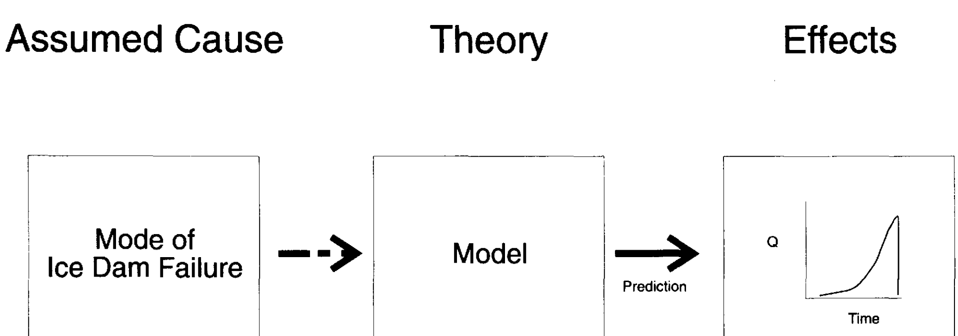

Fig. 1. Deductive inference in which a theoretical model, employing first principles, predicts various effects (glacial lake outburst discharges). Note that causative phenomena must be assumed.

became respectable to pursue as science (Baker 1981). In deduction one derives (predicts) results or effects from first principles, the best theory of nature codified into a model. Somewhat hidden in this procedure, however, is the need to assume the causative reality to which that model applies. For the late Pleistocene Lake Missoula flooding (Baker 1973) the assumption must be made as to the mode of ice-dam failure (Fig. 1). It is reasonable to assume failure modes commonly observed in modern ice-dammed lakes. Using such reasoning, Clarke *et al.* (1984) predict reasonable peak discharges for Lake Missoula flooding, with best estimates on the order of $3 \times 10^6\,m^3\,s^{-1}$.

Another approach is inductive. Here multiple observations of effects are generalized into a theory that can be extrapolated to the range of unobserved phenomena. In this manner Beget (1986) compiled historical information on volumes of ice-dammed lakes and the associated peak flood discharges generated by release of those volumes in failures. The resulting relationship (Fig. 2) can be extrapolated from the observed population to the volume of ancient Lake Missoula, yielding an approximate peak discharge of $2 \times 10^6\,m^3\,s^{-1}$. This would appear to be a confirmation of the model prediction. The logic by which it was achieved is essentially that advocated for

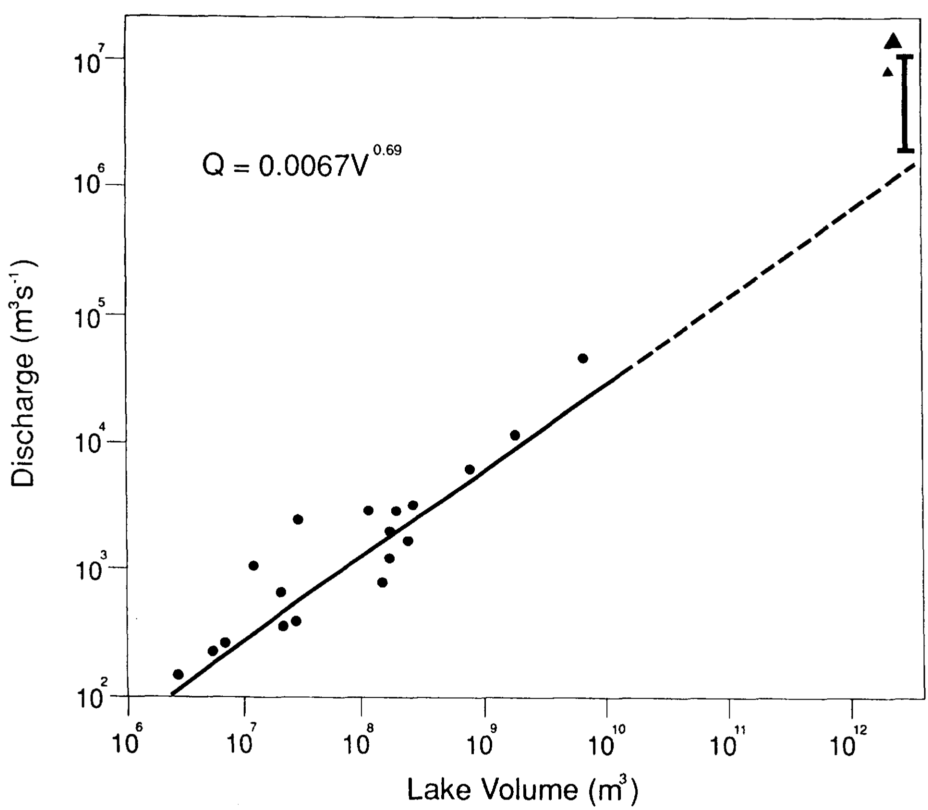

Fig. 2. Relationship of peak glacial lake outburst flood discharge, Q, to lake volume, V (Beget 1986). The relationship is extrapolated to the volume of glacial Lake Missoula ($2.5 \times 10^{12}\,m^3$). Modern data indicated by dots. Lake Missoula flood peak estimates of Baker (1973) indicated by triangles.

mathematical/predictive global change research: models are developed to predict future effects and measured effects are used as data to test the model results.

Again there is a hidden assumption. Induction, like deduction, must assume the nature of the causative phenomenon. In the simple example above, the lake volumes are assumed to be the only significant cause according to the relationship established for the relatively small modern lakes that were studied (Fig. 3). The extrapolation involves the assumption that Lake Missoula, two orders of magnitude larger than any lake in Beget's (1986) sample, also failed in a manner similar to those in the sample. Given the presumed confirmation of the model, however, these assumptions might not be questioned unless some other reasoning process is brought to bear.

If Beget's (1986) sampling of lake failures had been part of a carefully controlled experiment in physics, there would be no need for worry that extraneous 'causes' might creep in to invalidate the results. Palaeohydrology lacks this luxury, so it must escape the resulting limitations by employing the third form of inference: retroduction. Retroduction involves a kind of inverse reasoning, but it is not inverse deduction since theories are not generated from effects. In the cataclysmic flood example it is the causative process elements (flood discharges) that are generated by the observed effects, e.g. flood high-water marks and palaeochannel geometries (Baker 1973). The generation is possible because experience has been codified into an acceptable form for this purpose, in this case hydraulic principles that describe the physics of flow (Fig. 4). The theoretical principles are not being sought; they are being applied. It is the causative process that is being sought. The remarkable thing is that this quest discovers new things. O'Connor & Baker (1992) applied improved hydraulic theory to determine the Lake Missoula peak outflow at about $17 \times 10^6\,m^3\,s^{-1}$. This figure is nearly an order of magnitude larger than Beget's (1986) generalized prediction that might (unwisely) have been used as confirmation of the Clarke *et al.* (1984) model prediction. What does this anomalous result mean?

In central Asia evidence of another great late Pleistocene cataclysmic flood episode has recently been found (Rudoy & Baker 1993). For an ice-dammed palaeolake (Kuray) of similar size to that of glacial Lake Missoula, Baker *et al.* (1993) calculate a peak palaeoflow of $18 \times 10^6\,m^3\,s^{-1}$. Additional recent work on field evidence relating to other cataclysmic floods (O'Connor 1993; Teller & Thorliefson 1987) allows a new look at the pattern for these great floods. Baker *et al.* (1993)

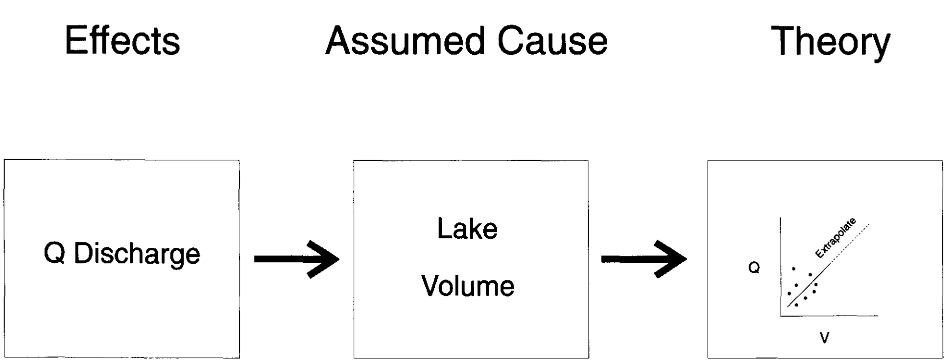

Fig. 3. Inductive inference in which observed effects (glacial lake outburst discharges) are related to their presumed causative factor (lake volume) to yield a generalization (theory).

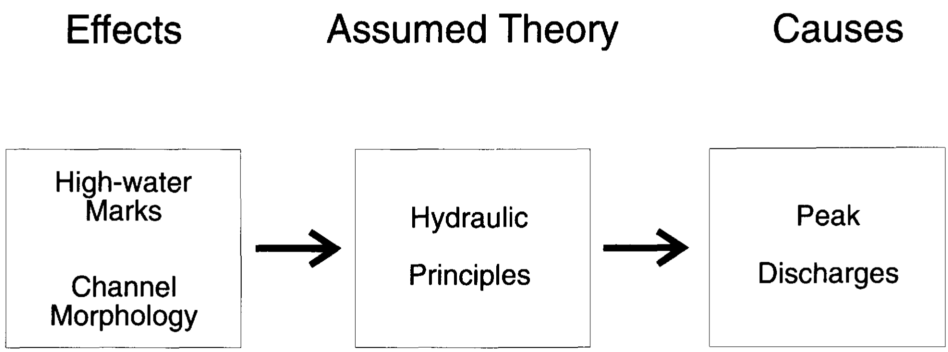

Fig. 4. Retroductive inference in which observed effects (indices of flood processes) are used to infer the causative processes (peak discharges). Theoretical concepts (hydraulic principles) are assumed to apply.

hypothesize that the very big, unusually deep ice-dammed lakes (Missoula and Kuray) may fail in a more cataclysmic fashion, similar to the failure of constructed dams. The glacial spillway floods (Bonneville and Lake Agassiz) occur in a progressive manner, similar to the way small ice-dammed lakes fail in the population studied by Beget (1986). This hypothesis, illustrated in Fig. 5, will require further work. The point is that its discovery was through retroduction. It contributes to enhanced understanding of reality, and it might not have been pursued if one blindly accepted

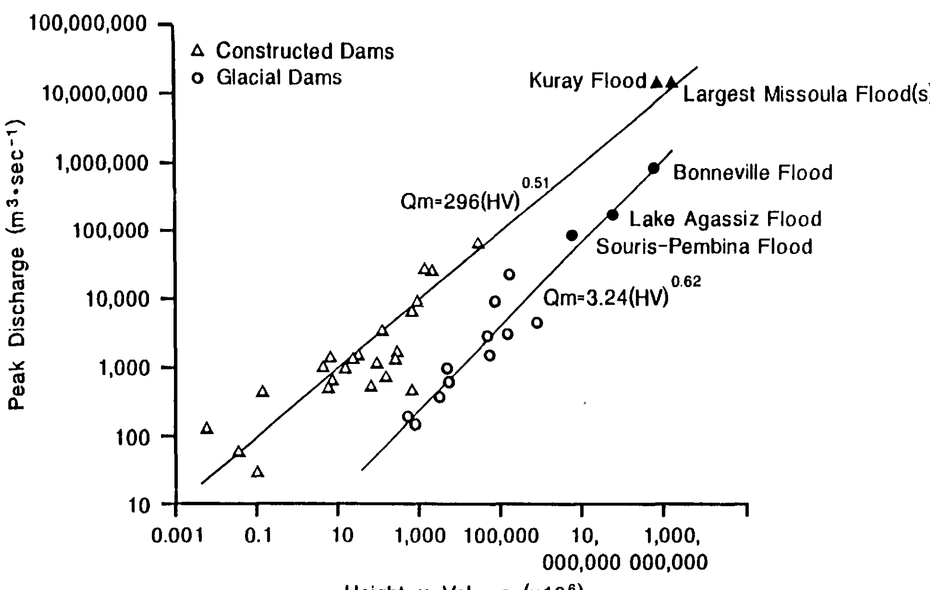

Fig. 5. Relationship of peak discharge (Q_m) to the product of dam height (H in m) times lake volume (V in m^3). Data for failed artificial (constructed) dams form a separate population from those for relatively small glacial lake failures. Some spillway failures of proglacial lakes (black dots) follow the glacial lake curve, while giant ice-dammed lakes (black triangles) follow the constructed dam curve (Baker *et al.* 1993).

the logic of model predictions and verification. Note also that this example employs a numerical method in the retroduction, but it is used as a tool for the inference rather than as a dictate for the way in which data are to be used. Most retroductions are nonquantitative, as illustrated by the insights of Bretz (1923), without which this whole line of inquiry would never have been pursued in the first place.

The consequences of scientific reasoning

Overreliance upon the mathematical predictive approach of the experimental/theoretical sciences has some important consequences for policy in regard to future environmental change. In scientific practice, prediction is a tool used to pose logical consequences of hypotheses that can be subsequently explored through controlled experimentation. Prediction applies to idealized systems, in which the flows of matter, energy, and information can be modeled mathematically. However, policy is driven as much by myth as by method (Girling 1993). Popular myth holds prediction to be accurate preknowledge of future events. The myth could conceivably be true, given immutable, invariant lawlike behaviour to all nature. However, there is a strong argument that the truth of this myth can never be known via the numerical models employed for prediction. Complete confirmation of such models is precluded both logically and by the incompleteness of human access to the relevant phenomena (Oreskes *et al.* 1994).

Human action may be guided by the conceptual generalizations of science, including model prediction, but its basis lies in perception-based experience (see, e.g., Baker 1994*c*). Experience involves concrete particulars. To the degree that these stimulate the mind to grasp reality, these real experiences contribute to human understanding Such experiences can be rather haphazard, and the lack of careful attention accorded them explains many policy failures. However, it is the role of the naturalistic/historical scientist to synthesize experience into a meaningful whole. Herein lies the most important task for the palaeohydrologist.

As a policy is enacted, it must be informed. That information can include attempts to perceive future events, subject to various uncertainties. However, uncertainty will always be present, a fact essential to the existence of science itself (Baker 1992). Action informed by reality, such as the past ranges of water balance palaeohydrologically signified in sediments and landforms, must be subject to continual validation against its consequences. It can, nevertheless, be confidently initiated, given the tendency of future conditions to evolve from past ones. By retaining options for adjustment, policy need not be founded upon a bedrock of scientific fact to achieve success. Rather, it can follow as a natural extension of science itself, transforming its basis of perception into the generality of guidance, always subject to the fallibility of the latter's incomplete formulations.

Discussion

What of the two styles of science that have been so sharply contrasted (Table 1)? There are some who might despair at this characterization of two incompatible approaches to science. However, the critique has focused on the subsumption of one style (reality-oriented) beneath another (theory-oriented). An alternative view might hold for an equilibrium. The eminent biologist Edward O. Wilson recently addressed a similar incompatibility in his discipline (quoted by Barlow *et al.* 1993):

> Much of the history of biology can be expressed metaphorically as a dynamic tension between unit and aggregate, between reduction and holism. An equilibrium in this tension is neither possible nor desirable. As large patterns emerge, ambitious hard-science reductionists set out to dissolve them with nonconforming new data. Conversely, whenever empirical researchers discover enough new nonconforming phenomena to create chaos, synthesizers move in to restore order. In tandem the two kinds of endeavors nudge the discipline forward

Another way to view this incompatibility is to accept that it is fundamental to nature itself. If science does indeed provide 'a mirror to nature' as many philosophers suppose, then the essence which it mirrors is the quantum world of subatomic physics. That world is one of mutually exclusive categories to our understanding but unified reality, as exemplified in the duality particles and waves. The great physicist Niels Bohr resolved this paradox through his idea of complementarity as applicable to our ability to describe the physical system of the subatomic quantum world. Complementarity implies the existence of seemingly incompatible aspects of the description of a physical system, which, nevertheless, are needed to achieve a complete description of that system. As noted by Petersen (1985), Bohr held 'It is wrong to think that the task of physics is to find out how nature is. Physics concerns what we can say about nature'. In similar fashion, it may well be incomplete science to assume that theoretical/predictive Earth-system science, following the example of mathematical physics, will ever provide more than an exceptionally clear exposition of what we can say about future global environmental change. For completely satisfying understanding of that change we will need another insight. Though lacking in the logical rigor that we seek to arbitrarily impose on it, nature's own language, interpreted as signs through the naturalistic/historical approach of sciences like palaeohydrology, provides the needed scientific insight.

To preserve a continually habitable planet through an uncertain future of global environmental change science will need to develop the best possible mathematical models to explain that future. However, just as those models will need to embody the best that scientists can say about the real world, it will also be necessary to explore the real experience of that world for the best that it can say to us. The challenge to palaeohydrology is for its practitioners to devote just as much effort to exploring the real world of Earth experience as they devote to idealizing the abstract world of Earth systems. By preserving this essential tension, science will advance.

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Climate change and flood sensitivity in Spain

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Abstract: The temporal and spatial distribution of more than 2500 records of historical floods in Spain, gathered within a database, were analysed. Our results suggest that the last millennium can be divided into periods with similar flood frequency, magnitude and distribution. This flood variability seems to correspond to changes of the prevailing atmospheric circulation patterns affecting the Iberian Peninsula. The present climatic flood-generating conditions were used in order to identify those changes for each period. In the Atlantic river basins, large floods between AD 1400–1500 and AD 1850–1910 resulted from intense, widespread rainfalls associated with Atlantic west and northwest frontal systems transported by the westerlies. In fact, both periods seem to experience a high hydrological variance, corresponding to a transition between different climatic conditions. Between AD 1150 and 1400 two hydroclimatic periods were distinguished: the first (AD 1150–1290) was characterized by a high flood frequency, especially over the Atlantic basins, some of these floods affecting more than one drainage basins. The flood distribution pattern for the second phase between AD 1290 and 1400, was characterized by a decrease in the number of floods in the Atlantic basins and a relative increase of flooding in the Ebro and Júcar basins. Finally, the AD 1500–1850 interval was characterized by a higher flood frequency, especially in the south and southeastern basins, due to cold pool conditions, as well as by a high irregularity of large floods associated with a high temperature contrast between winter and summer.

Flood magnitude and frequency vary among drainage basins depending upon the network variability in scale and morphometry, but mainly on the weather systems that produce flood events. Undoubtedly, modifications in the atmospheric circulation pattern due to the greenhouse effect will also result in variations in the magnitude, frequency and distribution of floods. Actually, several authors (Knox 1993; Baker 1993) have stressed the high sensitivity of flood occurrence to modest changes in climate which are difficult to be detected using variations in average hydrological conditions. Furthermore, changes in magnitude and frequency of floods during the past can be analysed within the context of time-varying climatic conditions and within a spatial framework of local, regional, and global networks of changing atmospheric circulation patterns (Hirschboeck 1988). The study of the spatial distribution, magnitude and frequency of flood events based on historical and stratigraphical records provide a useful tool in defining the periods with high frequency of flood-producing conditions for each basin and, consequently, to quantify the persistency in the past of anomalous atmospheric circulation patterns.

In areas with high hydrological sensitivity to climate variations, such as the Iberian Peninsula, general circulation models (GCMs) predict a higher severity of floods and droughts. The study of flood temporal and spatial variability within the last millennium provides a real-world scenario for understanding the impact of global warming on extreme hydrological events. In addition, for the major rivers of the Iberian Peninsula, a high correlation between the total annual runoff and the flood magnitude within the instrumental record was found and, therefore, a coincidence between large flood events and anomalous wet years (Benito *et al.* 1994).

In this study, we first describe the main hydroclimatic flood-producing conditions that affect the major drainage basins of Spain. Secondly, we analyse the changes in the flood frequency distribution of the largest historical floods for the last millennium and, finally, we relate those changes to modifications in the atmospheric circulation patterns responsible for these floods.

Historical flood data

The historical data sets of floods in Spain are based on information extracted either directly from a wide range of written documents, such as official and ecclesiastical documents, local chronicles etc., or using previous compilations such as the works by Masachs (1948), Font (1988) and Canales (1989). In these compilations, the checking of every flood information in the original historical document has not been completed. For the current work, the database was filtered using cross references between different sources, and floods affecting only one location have been excluded.

The historical flood database contains more than 2500 records starting from 181 BC in the Duero River, AD 449 in the Tagus River, AD 620 in the Guadiana River,

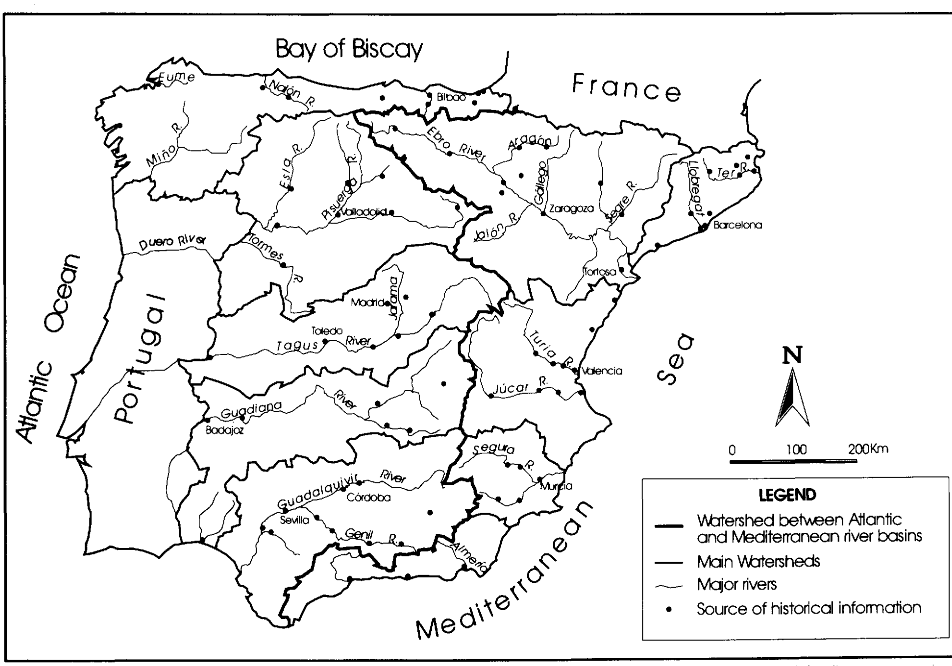

Fig. 1. Location of the main rivers and watersheds in Spain.

AD 100 in the Guadalquivir River, 49 BC in the Ebro River, AD 1088 in the Turia River, AD 934 in the Segura River, AD 1522 in the Northern Rivers, AD 1489 in the Eastern Pyrenean Rivers, and AD 1554 in the Southern Rivers. However, it is only since AD 1400 that flood records were widely documented for the major rivers. In Fig. 1 the location of the main archives and reference points of the historical information are shown. In our data set, each record contains the following fields: name of the river, year, month, day and duration of the flood, coordinates of the river reach affected by the floods and the original historical reference.

The reliability of this information depends on the spatial distribution of the data as well as on the continuity of the historical record. In many cases, the historical flood data sets are incomplete and their errors are not homogeneous, i.e. reports from different sources or epochs have different errors (Pavese *et al.* 1992). In fact, the database is biased since most of the written documents refer to populated areas and/or cities. However, the Spanish population through time have always been concentrated mainly along the middle and lower courses of the major rivers. In addition, for the last 1000 years the largest floods of the major rivers have been recorded in ecclesiastic and official documents of the main cities. One must not forget that the population, around AD 1000, for Valencia, Mérida and Toledo was already of 37 000 inhabitants, Zaragoza 17 000 inhabitants, Córdoba 100 000 inhabitants, Sevilla and Málaga 20 000 inhabitants, Almería 27 000 inhabitants and Granada 26 000 inhabitants (García & González 1994).

Flood magnitude for the historical record is difficult to estimate either quantitatively or qualitatively. At some towns, peak discharges can be calculated using rating curves and flood stages indicated by marks on buildings. Other historical records were referred to these marks as relative flood magnitude. However, most of the floods were recorded because they exceeded the stream channel and caused damages to agricultural lands, bridges and houses. In the second half of the nineteenth century and especially during the twentieth century, the increase of human activities over the floodplain increased the number of reported floods producing damage. A detailed analysis of these records is required to distinguish between the possible human and/or hydroclimatic causes in the increase in the number of reported floods.

Hydrology and flood hydroclimatology in Spain

The Iberian Peninsula is surrounded by the Atlantic Ocean to the west and by the Mediterranean sea to the east (Fig. 1). The influence on weather of these large water masses together with the orographic characteristics of the Peninsula result in a distinct division between the basins draining towards the Atlantic and to the Mediterranean. The Atlantic basins drain 69% of the Peninsula, and the major rivers with up to 1200 km in length are Duero, Tagus, Guadiana and Guadalquivir. Only one main river in Spain, the Ebro, drains towards the Mediterranean. This basin division is related to the structure and the geological evolution of the Central Plateaux and the Alpine Mountain Systems. Besides the geographical and morphostructural controls these drainage basins are also affected by different air masses which due to their characteristics are responsible for different rainfall events in terms of distribution, seasonality and duration.

Circulation types within the Iberian Peninsula

The climate of the Iberian Peninsula is characterized by a clear seasonal and monthly variability, with hot and dry summers, whereas winters are usually mild and relatively wet. This regime is controlled by two main systems: the subtropical anticyclone of Azores during summer and the westerlies in winter, associated with the invasion of cold fronts. Some major seasonal trends can however be identified, each of them associated with the frequency of a given circulation pattern.

From the hydrological point of view and particularly relating to flood analysis in the Iberian Peninsula we must draw our attention to the cyclonic circulation types, which are related to most of the rainfall events occurring in this region. Capel (1981) in his study of the weather types of Spain, using climatological data for the period lasting from 1964 to 1979, has identified the circulation patterns associated with anticyclonic (53.5%) and cyclonic (40.8%) regimes. For the latter he recognized the existence of five main types of circulation: northern flow, northeast, west and northwest, south and southwesterly type, and finally the cold pools.

Northern flow. Associated with an undulating circulation, with the main low pressure centre located at the south of Scandinavia, and a secondary cell over France or over the Occidental Mediterranean, this system is originated by the invasion of a maritime polar air mass over the Peninsula. This cold air mass on its movement south increases its temperature (at the lower levels of the air mass) due to the contact with the sea water surface, and therefore becomes more unstable. This situation, which occurs mainly in early winter and early spring, is responsible for the intense rainfall in the northern part of the Iberian Peninsula, namely in Galicia, Cantabria and Basque Country, and for some minor events at the northern Submeseta and the Ebro Valley.

Northeast flow. Associated also with an undulating circulation, this system is originated by the invasion of a Polar continental air mass (Siberian low pressure cell), and the development of a cold pool affecting the Southwest of Europe. It is often responsible for a general decrease on the temperatures as well as for snow events which affect mainly the eastern part of Spain: Pyrenees, Cantabrian Cordillera, Central System, Mountains of Toledo, Sierra Morena and the Betic Chain, whereas in the northwestern part, Galicia, Cantabria, and the headwaters of the Duero and Ebro valleys, some rainfall events may take place.

West and northwestern flow. This circulation type is associated with zonal flow, where Atlantic disturbances (frontal systems) transported by the westerlies can affect almost all the Iberian Peninsula, crossing it from west to east. This system is responsible for most of the precipitation that falls over Portugal and Spain, as well as for most of the flood events. Nevertheless, the type of trajectory of the zonal flow under cyclonic regime may introduce some changes in the rainfall pattern within the Iberian Peninsula.

In fact, when there is high index zonal circulation over the Atlantic (over 60° N), depressions move far from the Peninsula, and only the northern part of Spain may be affected by the cold-sector air, producing scarce rainfall events over Galicia and Cantabria. A totally different situation concurs when the zonal flow lies between

45° and 50° N. In this case, the Iberian Peninsula is affected by the successive eastward passage of frontal systems, responsible for continuous and persistent precipitation which affects mainly the northern areas, as well as the Atlantic catchments of Duero, Tagus, Guadiana and Guadalquivir. When the zonal flow is lower in latitude (35–45° N), the whole Peninsula is affected. Nevertheless, this circulation type has null or little effect on the southeastern Mediterranean areas (Levante).

South and southwestern flow. A change in the zonal westerlies flow direction by a high pressure cell over Scandinavia and Eastern Europe is the origin of this situation, allowing the invasion of maritime tropical air masses over the southeast part of Occidental Europe, whereas a low pressure cell resulting from the invasion of cold air masses (polar or arctic maritime) is developed in the west of Ireland. This cyclonic cell can, however, be lower in latitude, at 50° N (northwest of Galicia), and remains stable over one week. During this period successive northeasward frontal systems cross the Peninsula, affecting mainly the southern part of Spain as well as the whole Mediterranean area, including the pre-Pyrenean areas of Aragon and Catalonia.

Cold pools. Cold pools cannot be considered as a different circulation type but as a subtype of the southwest flow in altitude (Capel 1981). Nevertheless, their influence on the rainfall distribution and in the flood dynamics of some areas of the Iberian Peninsula is quite important. A cold pool can be defined as a cyclonic cell resulting from the advection of arctic or polar air masses, and can be identified on altitude synoptic charts (500 hPa).

These low pressure cells are frequently associated with three main blocking circulation types: difluent blocking, cut-off-low and omega. In fact, within the Iberian Peninsula 52% of the cold pools are due to cut-off-low (more frequent during early summer and fall), whereas the other two types, difluent and omega, are responsible, respectively for 33% and 15% of the cold pool genesis (Ventura 1987). From the point of view of their seasonal and spatial distribution, Llasat (1991) has determined for the period between 1974 and 1983, that spring and summer are the seasons more affected by cold-pool situations (38% and 25% respectively), whereas fall and winter had registered only 19% and 17% of the cases. Most of these cold pools were found to be centred over the western part of the Peninsula (Galicia and central part of Portugal), as well as in some areas of the Meseta. This general distribution does not, however, present a correlation with the rainfall pattern. In fact, although most of the cells are centred over the western part of the Iberian Peninsula, this is the area where less precipitation is due to cold-pool conditions, whereas the highest amounts of rainfall take place at the southeastern part (Levante). In this area, precipitation over 500 mm/24 hours have been recorded, giving rise to catastrophic flooding. Concerning the seasonal pattern, although the highest number of cold pool cells is registered in spring and summer, the highest rainfall amounts in 24 hours take place at fall and winter. Regarding their inter-annual distribution, cold pools are more frequent in dry years (Ventura 1987), related to changes in the atmospheric circulation: decrease of the zonal circulation and of frontal rains followed by an increase of the blocking situations.

Influence of the circulation types on the rainfall and flood regime

In Fig. 2 two circulation types can be pointed out as the most frequent, as well as the ones with higher influence in the rainfall regime: the southwestern and the west/northwestern flows.

During winter, the west and north-western flows are dominant, closely related to high frequency of the zonal circulation in altitude. This situation most influences the areas affected by the Atlantic air masses, namely the drainage basins of Duero, Tagus, Guadiana, Guadalquivir and even the northern basins of Galicia and Cantabria. These latter ones are however more influenced by the intense rainfall originated by the northern circulation, which can also affect the Ebro and Duero basin headwaters.

In spring, as well as in late winter, with the expansion of the jet and the domain of the undulating circulation pattern, there is a change in the main flow type by increasing the south and southwest flows, which have their highest frequency in late spring. This kind of circulation is responsible for important precipitation volumes at the east and southeast part of Spain, namely on the Mediterranean drainage basins of Júcar, Segura, Ebro, Eastern Pyrenean rivers and on the Southern rivers. Some rivers at the north of Spain also experience a second peak discharge during spring. These spring high waters are due to sudden snow-melt, affecting mainly the rivers fed into the Cantabrian Cordillera as well as the Pyrenean tributaries of the Ebro River.

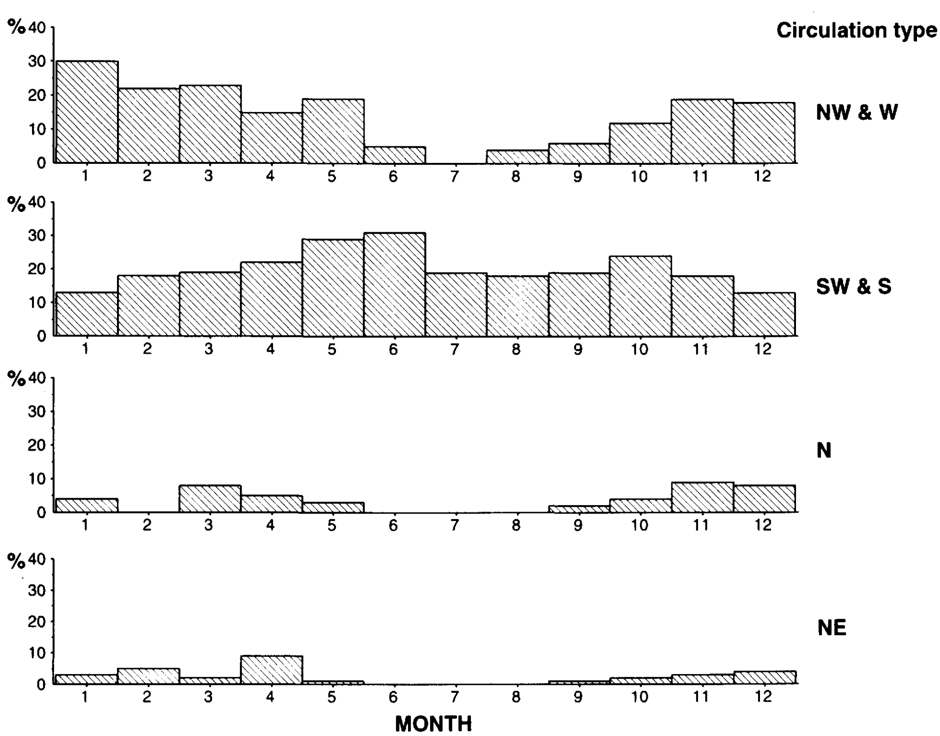

Fig. 2. Frequency of the cyclonic circulation types in the Iberian Peninsula, for the period between 1964 and 1979 (data from Capel 1981).

Further south, in the Guadalquivir basin, the Genil River can also present a second peak in early spring by conveying snow-melt waters from Sierra Nevada.

The summer is characterized by an almost total lack of rain, especially south of the Cantabrian Cordillera. Nevertheless, in the northern part of Spain (Galicia, Cantabria and Basque Country), some flood events may occur, not only to cold pool conditions but associated with an anticyclonic circulation flow. In fact, the advection of the oceanic air-masses (orographic) in a southeastward direction, produces rainfall events which, regardless of their low intensity, are very persistent.

Finally, during autumn there is an increase of the west and northwest circulation, as well as of the southwest type. The latter is more frequent in late autumn, corresponding to an increase of the blocking circulation and the subsequent genesis of cold pools. These cells affect mainly the Mediterranean coast, leading to flood events on the Eastern Pyrenean rivers, Júcar, Segura and also in the Ebro basin and Southern rivers. In the case of the Mediterranean rivers that drain the Iberian Range (Júcar, Segura and Turia), the highest peak discharges are registered during this period. In fact, the mean discharge of these rivers can be multiplied by 11 000 times during the largest flood events (Masach 1950).

Flood variability within the historical record

In the analysis of the temporal variability of flooding during the last millennium, all the historical flood data were taken into account (Figs 3 and 4). However, the analysis concentrated on the variability of the largest flood events or floods reported to affect extensive areas. Concerning the monthly and spatial distribution of the historical floods all data available were also taken into account.

The monthly distribution of the historical floods shows similar patterns to the present monthly flood distribution (Fig. 5). For the major rivers most of the historical floods occurred during winter with a second peak during autumn. Only rivers fed by melting snow show a third peak during spring. The Mediterranean and Northern rivers present a maximum in autumn and a second peak in winter. Despite of the similar distribution pattern between historical and instrumental flood records, major changes in flood frequency and monthly distribution can be found when the largest floods for each individual drainage basins are analysed.

A significant number of floods took place in the Atlantic rivers between AD 1150–1290 (Fig. 3). These floods occurred between December and February, and some of the rainfall events responsible for this situation affected more than one basin. This is the case of the AD 1168 and AD 1258 floods which were extensive to the Duero, Tagus and Guadalquivir basins. Frontal systems crossing the Peninsula from west to east, associated with zonal flow, were responsible for most of this flood events. During this period, several documents mention also the existence of frost in the Tagus and Duero Rivers in AD 1191, 1194, 1201 that are likely to correspond to a more frequent invasion of Polar continental air masses from central and eastern Europe. In contrast to the present flood distribution, no floods were reported in the Spanish Mediterranean rivers during that period, probably due to a lack of historical documentation.

The lack of large flood events was extended to the whole Peninsula for the interval AD 1290–1400, with the exception of the flood events occurred in the Ebro (AD 1325

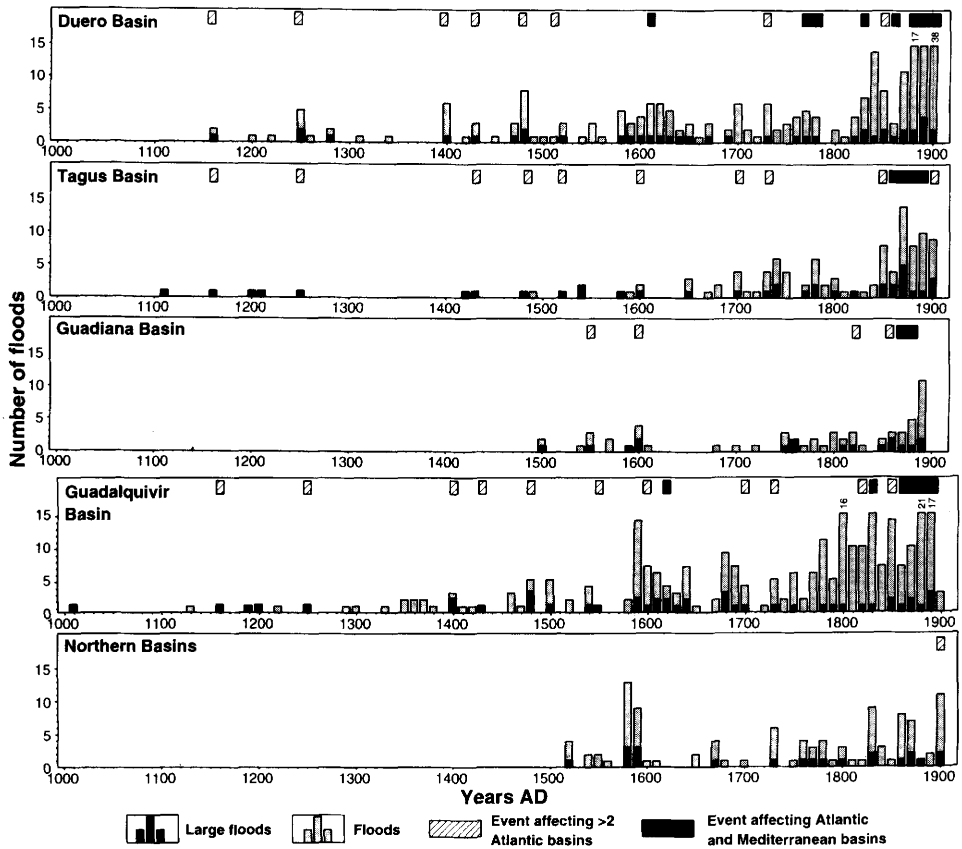

Fig. 3. Historical floods (per decade) in the Atlantic basins.

and 1380) and in the Júcar (AD 1328) rivers. In addition, minor floods were also reported in the Ebro tributaries, taking place in autumn and spring, as well as at some locations along the Guadalquivir (10 floods), Turia (4) and Júcar (2) rivers.

A significant increase in the number of large floods took place between AD 1400 and the first decade of AD 1500. The copious winter precipitation, particularly during AD 1402–1403, 1434, 1485 and 1488, resulted in severe floods which affected most of the Atlantic basins (Duero, Tagus and Guadalquivir, Fig. 6). These winter floods may have been produced by the passage of frontal systems at the surface, associated with a zonal flow low in latitude which brought up continuous and persistent precipitation over the Peninsula. At the same period, the historical records indicate a low number of large floods in the Mediterranean basins. In the Ebro Basin, floods were mainly produced by autumn and spring precipitation probably associated with an undulated circulation pattern with a SW flow pattern whereas in the Turia, Júcar and Segura basins, floods occurred mostly during October and November which may resulted from cold pool conditions.

For the AD 1500–1550 interval, the number of large floods in the Atlantic basins decreased in comparison to the previous century, except in the Guadalquivir Basin (Fig. 6). In contrast, the flood frequency increased in the Mediterranean basins,

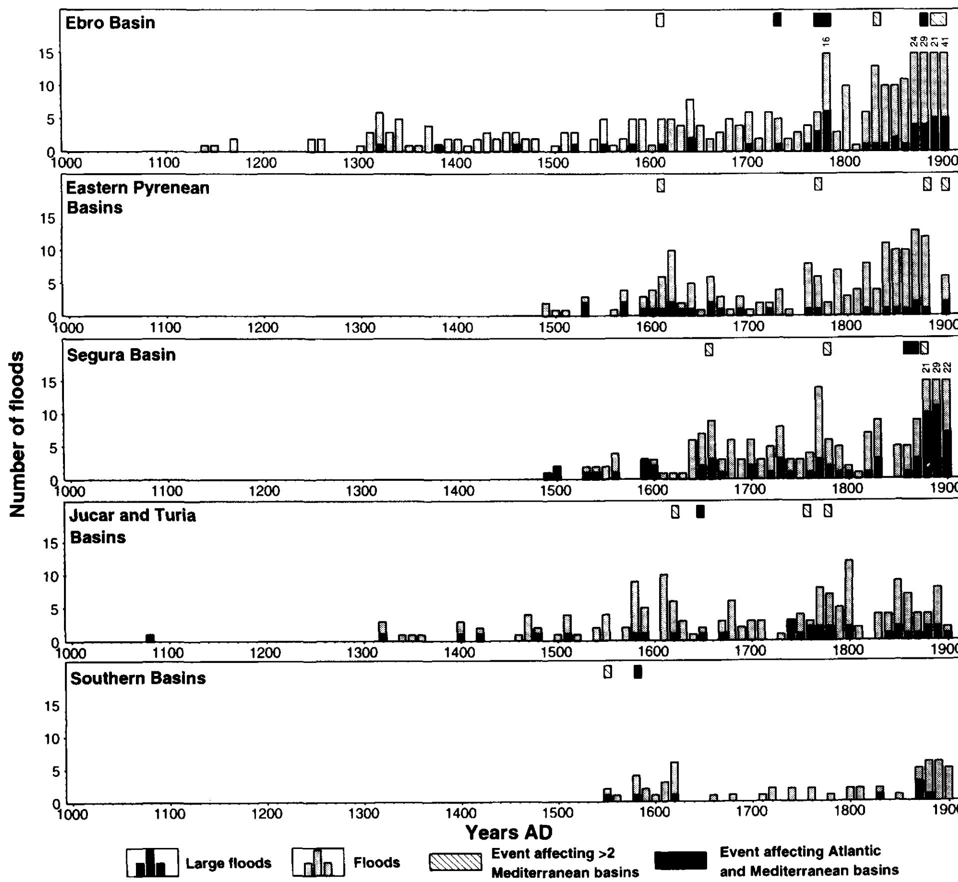

Fig. 4. Historical floods (per decade) in the Mediterranean basins.

especially in the Segura and Júcar basins. This flood frequency and distribution contrasts notably with the previous period and it may represent a shift northwards of the zonal flow and an increase in the undulating circulation types as well as of the blocking situations. A more frequent S and SW cyclonic flow, associated with an undulating circulation, is likely to be the origin of these flood events at the eastern part of Spain during late winter, affecting mainly the Guadalquivir basin. On the other hand, the Mediterranean basins of Júcar and Segura were probably more affected by cold pool cells, linked to blocking situations. Both circulation patterns are related to a higher annual and seasonal irregularity of the precipitation.

The number of large floods between AD 1550 and 1800 increased in comparison to the previous period although only few rainfall events produced floods simultaneously in several basins. In the Atlantic basins, this extensive flooding due to frontal systems transported by the westerlies was reported in AD 1603–1604 within the Duero, Tagus, Guadiana and Guadalquivir rivers, in AD 1626 affecting the Duero and Guadalquivir rivers, in AD 1708 in the Tagus and Guadalquivir, and in AD 1739 in the Duero, Tagus, Guadalquivir and Ebro rivers. In the Duero River, the December 1739 flood was estimated as one of the largest floods in the historical record with peak discharge of 19 000 $m^3 s^{-1}$ (Pardé 1955).

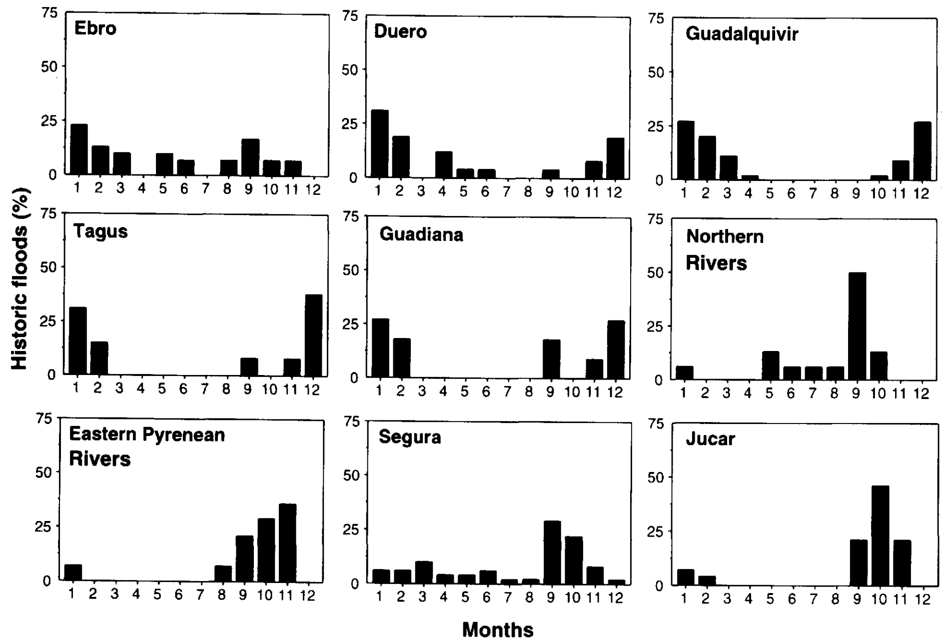

Fig. 5. Monthly distribution of the historical floods in the major rivers of Spain.

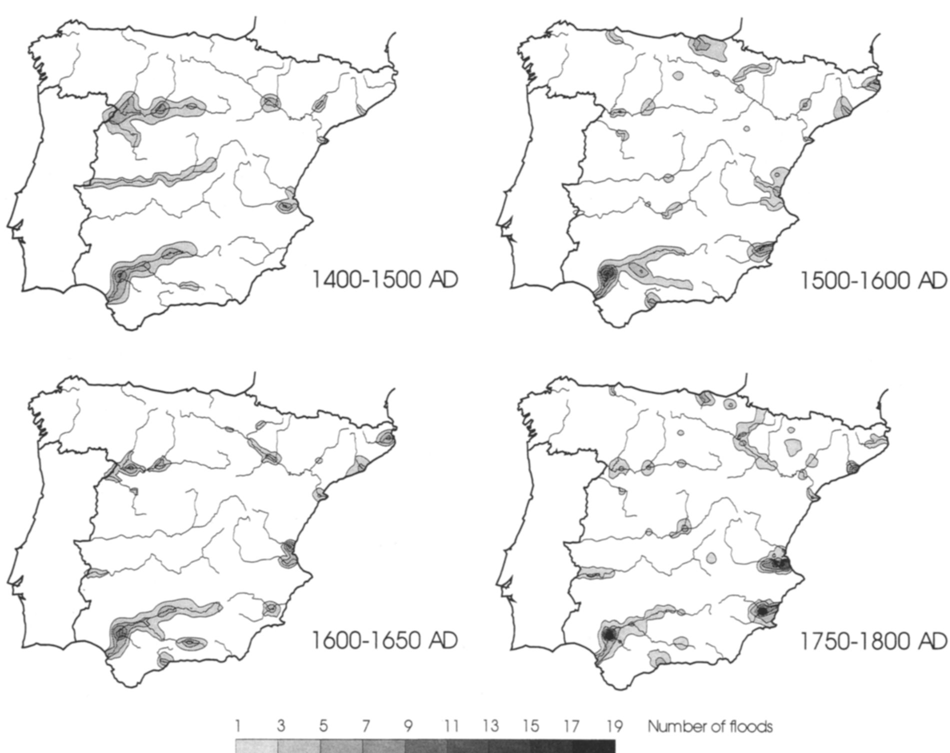

Fig. 6. Distribution of the historical floods in Spain.

In the northern basins the flood frequency increased for the period between AD 1550 and 1610, mainly in autumn and spring with generalized events in AD 1581, 1582, 1586, 1592 and 1593 (Figs 3 and 6). However, this period coincides with the AD 1566–1600 interval reported as the one with more recorded droughts in the northern part of Spain. Therefore, flooding in the northern basins seems to indicate a considerable irregularity in precipitation rather than anomalous wet conditions. The flood events were probably originated by northern flows with invasion of maritime polar air masses during autumn and spring over the Peninsula producing heavy rainfall in the Cantabrian region.

In the Ebro Basin, the frequency of large floods increased between AD 1640 and AD 1650 mainly during winter months at the Ebro headwaters, and between AD 1770–1790 during spring and autumn, affecting the Pyrenean tributaries (Fig. 6). Spring floods in the Pyrenean tributaries were triggered by melt waters that may reflect a more frequent entrance of northeastern flow with Polar Continental air masses. Increased flooding in the second half of the eighteenth century was also recorded in other Mediterranean basins. In addition, high flood frequency was also documented in the Mediterranean basins for the first half of the seventeenth century. These October and November floods were likely originated by cold pool conditions.

A decreased flood frequency with only two large events took place for the first half of the nineteenth century: in AD 1823, affecting the Guadiana and Guadalquivir basins, and in AD 1831 affecting the Duero, Guadalquivir and Ebro rivers. However, in the second half of the nineteenth century the number of floods in the Peninsula increased again (Figs 3 and 4). In the Atlantic basins, severe winter floods took place in AD 1856 covering the Duero, Tagus, Guadiana and Guadalquivir basins, in AD 1860 in the Duero, Tagus and Guadalquivir basins, in AD 1876 in the Tagus and Guadalquivir, in AD 1877 in the Guadiana and Guadalquivir basins, in AD 1881 in the Duero, Tagus, Guadiana, Guadalquivir and Ebro, in AD 1895 in the Duero, Tagus and Guadalquivir, in 1897 in the Duero, Guadalquivir and Ebro, in AD 1900 in the Duero and Ebro and in 1909 in the Duero and Tagus basins. In the Tagus River, the 1876 flood is considered the largest reported flood within historical times which rose the water level more than 30 m above its normal base flow in Alcántara, near the Portuguese border, with estimated discharge about 12 000 $m^3 s^{-1}$. In the Mediterranean basins the number of floods increased as well, with severe floods in the Turia River in 1843 and 1897, in the Segura River in 1879 with a estimated discharge of 4000 $m^3 s^{-1}$ and in the Júcar River in AD 1864 with an estimated discharge of 12 243 $m^3 s^{-1}$ (Solé 1978). The Ebro basin had floods recorded either in winter (some of them simultaneously with the Atlantic basins) or in spring and fall, originated either at the central Pyrenean or Iberian tributaries respectively.

The widespread winter flooding affecting most of the Atlantic basins and even some of the Mediterranean rivers may be considered to represent a significant change in the atmospheric circulation patterns in relation to the previous centuries. The successive eastward passage of Atlantic frontal systems transported by the zonal flow its most likely to be in the origin of these flood events. The persistent precipitation over most of the Peninsula during consecutive years may indicate that the zonal flow was low in latitude. In addition, flooding in the Guadalquivir, Ebro and even in other Mediterranean basins were probably the result of a south and southwestern circulation pattern.

Flood response to climate variability

Most of the historical evidence used by Lamb (1977) to define major climatic periods is based on data reported on Central and Northern Europe. Therefore, one of the major problems in understanding the flood response to climate variability is the inconsistency of these climatic periods over continental scales. This is the case of the Medieval Warm Period considered by Lamb (1977, p. 435) to occur between AD 1150 and 1300. However, in the Mediterranean region Alexandre (1987) indicates that no exceptional winter warmth is evident in the available record until the mid-fourteenth century. In fact, between 1150 and 1250 there are multiple evidences of frost in the Tagus and Duero rivers as well as high waters due to snow-melt in Mediterranean rivers. In addition, global-scale generalizations for relationship between extreme events and shifts in global atmospheric circulation patterns may underestimate the impact of climate change because climate and hydrological responses vary regionally (Knox 1993)

In spite of this, historical flood analysis cannot provide evidence about the direction of the climatic shift, although flood frequency and distribution can provide an indication that climate change has occurred. The historical flood analysis shows different time intervals with homogeneous response of flood frequency and distribution in Spain. These periods may represent major shifts in the atmospheric circulation patterns of Southwestern Europe.

In the Atlantic basins, a similar distribution of large floods was found for the intervals between AD 1400–1500 and AD 1850–1910. These floods resulted from intense, widespread rainfalls associated with Atlantic west and north-west frontal systems transported by the westerlies. This atmospheric circulation pattern was responsible for a spatial distribution of rainfall, which was able to produce severe flooding simultaneously in the Duero (98 375 km^2), Tagus (81 947 km^2), Guadiana (67 500 km^2), Guadalquivir (57 421 km^2) and even the Ebro (85 997 km^2) basins. These two periods with numerous large floods coincide with changes in regional and global temperatures, and probably with the shift of the zonal flow to lower latitudes and vice versa.

The AD 1400–1500 interval seems to coincide with a transition from warm to cool climatic conditions which prevailed from the sixteenth to the eighteenth century in the Iberian Peninsula. The AD 1850–1910 period corresponds to a progressive change towards the warmest conditions of the twentieth century. Similarly, flood magnitude and frequency seems to increase during transitional climatic conditions in the southwestern US (Ely *et al.* 1993) suggesting a higher hydrological variance.

The historical records, from the sixteenth to the eighteenth century, suggest the existence of a colder phase in the Iberian Peninsula with a climax in the seventeenth century. During these centuries several rivers were frozen such as the Ebro River in Tortosa, 15 km from the coast, in AD 1503, 1506, 1572, 1573, 1590, 1607, 1623–1624, 1648–1649, 1678–1679, 1693–1694 (with 3 m in thickness) and 1696–1697, and the Tagus, at Toledo, in AD 1529, 1530, 1536, 1678–1679 and 1696–1697. In addition, severe winters with snow and frost, even in the southeastern part of Spain, were reported for 1503–1511, 1529–1539, 1541–1548 and especially after 1560 until 1850. In opposition to the cold winters, summers are generally described as very dry and hot, with several locust plagues in 1508, 1519, 1546–1551, 1570–1576, 1586,1591, 1618, 1621, 1629,1638–1641, 1648–1650, 1656, 1666, 1669, 1671–1672, 1675, 1685, 1687–1688 and 1693 (Font 1988). This climatic phase seems to be influenced by two main air masses: invasions of cold polar continental air masses, associated with the north-eastern flow during winter and early spring, and warm tropical and subtropical continental air masses with a southern flow, associated with a shift of the Saharan depression towards the Peninsula, during summer.

The hydrological response of the main rivers to this prevailing circulation pattern resulted in a higher flood irregularity with a decrease of the frequency of large floods and particularly in the number of extensive large events in the Atlantic basins. These large floods occurred only around AD 1600 and 1700 (Figs 3 and 4). However, floods seem to be more frequent in the southern and southeastern river basins (Figs 4 and 6), mainly during autumn, although most of them were described as moderate-magnitude floods, with the exception of the flooding in the Segura River in the 1600s. These floods were likely to be originated by a higher frequency of cold-pool cells. In the present time, these cells are more frequent in dry years, and related to a decrease of the zonal circulation followed by an increase of the blocking situations.

Conclusions

Historical evidence illustrates the high sensitivity of flood magnitude and frequency to climate variability within the last millennium. The two periods of extensive large-magnitude floods between AD 1400–1500 and AD 1850–1910 correspond to transitional climatic conditions. During these periods, zonal flow probably located at lower latitudes prevailed over meridional flow. A higher irregularity of large floods was found between AD 1500–1850, associated with a very high temperature contrast between winter and summer. However, for this period, a high frequency of floods in the south and southeastern basins due to cold pool conditions was reported. This flood distribution is associated with a high frequency of the undulating circulation which is also responsible for the high temperature contrast. These changes of the flood pattern with climate suggests that the predicted global warming will lead to major modifications in the flood distribution, magnitude and frequency within the Southwestern Europe. In addition, flood sensitivity to climate gives rise to questions over the applicability of the standard flood frequency analysis estimating high-magnitude low-frequency floods without taking into account the changing hydroclimatic conditions.

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Geochronologies and environmental records of Quaternary fluvial sequences in the Guadalope basin, northeast Spain, based on luminescence dating

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Abstract: Geomorphological investigations of Quaternary fluvial deposits in the Guadalope basin, NE Spain, have revealed an extended record of river and drainage basin response to long term environmental change. Investigations have focused on a 13 km long reach centred on the town of Castelserás and a 4 km long reach at Mas de Las Matas. The study reach at Mas de Las Matas is located at the confluence of the Rio Guadalope and its principal tributary the Rio Bergantes. The Castelserás reach is situated 16 km downstream from the Bergantes confluence within a relatively narrow bedrock-confined valley.

Luminescence dating provides a geochronology indicating that deposits in both study reaches span the Mid-Pleistocene through to the historic period. The Mid-Pleistocene was characterized by large-scale (locally up to 40 m) aggradation and incision cycles associated with tributary stream and slope input of sediment. More gradual alluviation in oxygen isotope stages 4 and 5 was followed by renewed dissection and aggradation immediately preceding the Last Glacial Maximum. At least four river terraces were formed during this period, with heights above present river level of between 10 and 15 m. River instability continued after the glacial maximum and into the Holocene, when the most recent major alluvial unit was formed 3000 to 5000 years ago.

Progressive entrenchment of the Guadalope valley floor during the Pleistocene would appear to reflect long-term regional uplift with channel aggradation and dissection episodes linked to the climate changes shown by vegetation and oxygen isotope records. Rates of downcutting increased during oxygen isotope stages 2 and 3. This change probably occurred in response to a reduction of coarse sediment delivery to the main channel from tributary catchments, an increase in trunk stream competence and/or catchment uplift.

The proximity of Spain to the North Atlantic Ocean implies the region is likely to be sensitive to perturbations in the ocean-atmosphere system, such as change in North Atlantic thermohaline circulation (Paillard & Labeyrie 1994), Heinrich events (Bond *et al.* 1992) or change in North Atlantic deep water formation (Keigwin *et al.* 1994). Furthermore, in this area, the sensitivity of rivers to Quaternary environmental change, together with conditions promoting the formation and preservation of river terraces, result from high basin relief (Fig. 1), long-term regional uplift and extensive

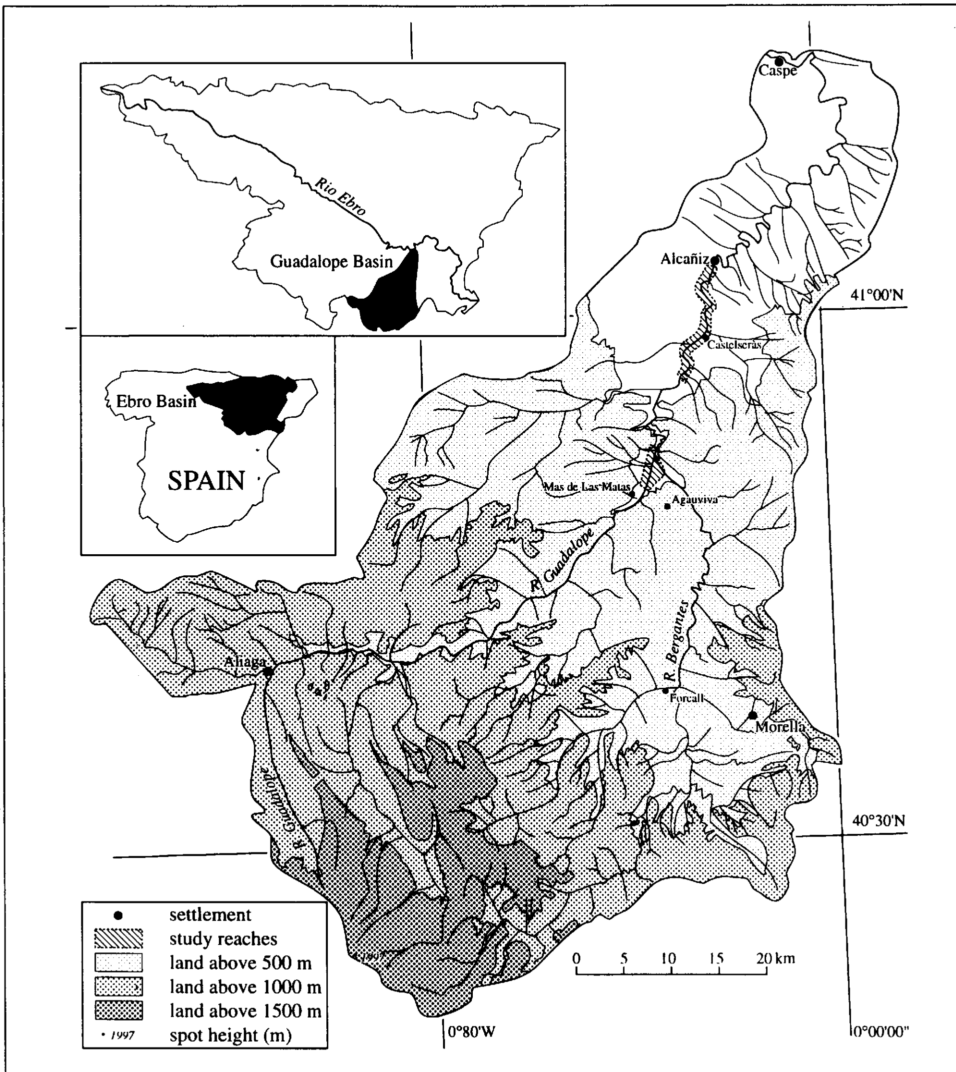

Fig. 1. Relief and drainage network of the Guadalope Basin. Locations of study reaches are indicated, as are places referred to in the text.

outcrops of sedimentary rock susceptible to mechanical degradation and erosion. North Atlantic climate oscillations have been registered in pollen records in the region, most notably at Lake Banyoles (Pérez-Obiol & Julià 1994). Detailed, high-resolution climate records from Greenland ice (e.g. GRIP ice core, Dansgaard *et al.* 1993) and lacustrine/peat bog sequences from Banyoles (Pérez-Obiol & Julià 1994) and Padul (Pons & Reille 1988) are now available, but there is a need to investigate how rivers have responded to the climate changes registered in these records during the Pleistocene. The main limitation in determining fluvial response to such environmental change has been the lack of a suitable dating technique which can be applied directly to fluvial material of Holocene and Pleistocene age. Geochronologies of Quaternary fluvial sequences, however, potentially constitute a long (albeit discontinuous) continental record of Quaternary environmental history.

Luminescence dating has been applied to fossil soils in Late Pleistocene terraces of the Júcar, Serpis and Turia rivers by Prószynska-Bordas *et al.* (1992) using thermoluminescence (TL). Similarly, Rendell *et al.* (1994) have used TL and optically stimulated luminescence (OSL) to date aeolian deposits overlying Pleistocene alluvium in the Júcar and Guadiana rivers. Infra-red stimulated luminescence (IRSL) has been applied to date Holocene alluvium in the Rio Regallo in the Ebro basin and a good agreement was found between luminescence ages of river sediment and AMS ^{14}C dating of incorporated charcoal (Macklin *et al.* 1994).

Dating techniques, such as luminescence, which are based on direct dating of a sediment unit and are able to estimate the age of the depositional event itself, can provide significantly improved chronological resolution. Determination of the drainage basin and river response to the impact of climate and tectonic forcing requires the precision obtained using direct sediment dating. Using this approach, river aggradation and incision episodes may be related to the climatostratigraphic framework provided by pollen, marine, ice core and lacustrine records.

This paper seeks to address the problem of determining cause and effect relationships in a river catchment and relate river response to the regional and global records of climate change during the middle and late Quaternary. This is enabled, primarily, by developing a luminescence-based geochronology of alluvial deposits in the Guadalope basin.

Study area and methodology

Study area

The Rio Guadalope is a major south-bank tributary of the Rio Ebro, draining an area of 3892 km^2. Its headwaters rise in the Sierra de Gúdar which reaches an altitude of 2019 m. The highest point in the catchment is 1997 m above sea level and 45% of the catchment lies above 1000 m (Fig. 1). Much of the catchment is dominated by the Iberian massif 'basement and cover' structure, where Palaeozoic rocks in the Montes de Maestrazgo are fractured into blocks and in places thrust over Mesozoic and later material. This geological setting has produced strongly deformed fault- and fold-guided relief above *c.* 500 m. Beyond the highly deformed upland (i.e. $<$500 m, Fig. 1), the Guadalope flows across relatively undisturbed, gently northeast dipping Miocene conglomerates, sandstones and silts. Macklin & Passmore (1995) cite evidence for regional uplift of this area of the Ebro basin during the Quaternary.

In the deformed upland of the Sierra de Gúdar, the river is entrenched in a gorge. The exception to this is the very upper reaches of the trunk valley which have been infilled by colluvium. Within the upland reaches, preservation of alluvium is confined to areas where the valley floor widens, as for example at the confluence between the Guadalope and its primary east-bank tributary, the Rio Bergantes (Fig. 1).

Two reaches of the Guadalope basin have been investigated. First, the alluvial basin at Mas de Las Matas has developed at the confluence zone between the Guadalope and Bergantes and is situated towards the northern edge of the upland area (Fig. 1). This reach has an extensive area of Quaternary alluvium (*c.* 4 km wide), which has been mapped, sampled and dated. The modern river at Mas de Las Matas has a gravel-bed, low sinuosity, single thread channel and is currently incising into its floodplain which consists largely of vertically accreted sands and silts.

The second study reach, at Castelserás, lies in the Iberian massif piedmont. The river here flows across more uniform lithologies and is entrenched and presently confined in a narrow bedrock slot. The valley, however, is locally up to 2 km wide and alluvium is well preserved. A 13 km long reach has been mapped between Castelserás and Alcañiz (Fig. 1). The modern river in this second study reach is characterized by a confined meandering pattern, with a narrow (<100 m) floodplain comprising vertically accreted sands and silts, circumscribed by bedrock bluffs. North of Alcañiz the Guadalope cuts through a sandstone–conglomerate escarpment, forming a series of very sinuous, deeply entrenched bedrock-confined meanders where alluvium is poorly preserved. Alluvium in both reaches is exposed in a number of gravel pits, road cuttings and river bank sections.

Climate and vegetation

Mean annual precipitation in the Guadalope basin varies from 300 to 400 mm in the Alcañiz–Caspe region to 700 mm in the Sierra de Gúdar. Much of the lower catchment is classified as semi-arid and the remainder as dry or very dry (Confederacion Hidrografica del Ebro: Memoria 1946–1975). The resulting hydrological regime of the catchment is classed as 'Pluvial incierto' (unreliable pluvial). January mean temperatures vary from 5–10 (Caspe) to 0–5°C (Morella), whilst the July mean ranges from >25 (Caspe) to 15–20°C (Morella). The net water balance varies from −500 mm (Caspe) to 0 mm (Morella). Evergreen coniferous forest covers parts of the upper catchment, but much of the vegetation cover throughout the Guadalope basin (where uncultivated) is maquis or garrigue.

Methodology

Fluvial landforms in both study reaches were mapped by field walking and plotted on 1:25 000 scale aerial photographs and enlarged 1:50 000 topographic maps. Breaks of slope >1 m were recorded. Terrace heights were measured using an aneroid barometer.

Sedimentary exposures were photographed and logged. Where exposed, fine-grained material (sands and silts), either capping or within a terrace unit, was sampled for sedimentological analysis and IRSL sediment dating. Samples for IRSL dating were collected from beneath a black tarpaulin using light-tight black plastic tubes. Where the sediment was particularly indurated, blocks of sediment were cut and then wrapped in light-tight bags.

Luminescence dating

A number of luminescence dating techniques has been developed to date sediments, notably thermoluminescence (TL), optically stimulated luminescence (OSL) and infra-red stimulated luminescence (IRSL). The most appropriate means to date alluvial sediments are optically stimulated, i.e. OSL and IRSL. These techniques measure the luminescence signal most sensitive to light and therefore optical bleaching. In the potentially light restricted environment of a sediment-laden, turbulent river, sensitivity of the luminescence signal to zeroing is important. TL, proven in dating well-bleached aeolian landforms and loess deposits (Wintle 1993), may be less reliable in dating alluvial deposits of unknown bleaching history (Fuller *et al.* 1994). The comparative sensitivities of TL and IRSL signals to bleaching in a water column have been demonstrated by Ditlefsen (1992). In this paper, IRSL measurements are made using potassium feldspars separated from the coarse grain fraction (>100 μm) of the sediment.

Age determination using luminescence dating is based on the following equation:

$$\text{age} = \frac{\text{ED (Gy)}}{\text{dose rate (Gy/ka)}}$$

ED determination

ED (equivalent dose) is determined using a partial bleach methodology (Wintle & Huntley 1980). The ED is a measure of the dose absorbed by the sediment and this is built up by decay of the naturally occurring radioactive elements ^{238}U, ^{232}Th (and their daughter products) and ^{40}K which liberates electrons that become trapped in crystal lattice defects within the sediment grain. The number of trapped electrons builds up with time. Stimulation with light (here IR to produce IRSL) or heat (to produce TL) dislodges trapped electrons which recombine at luminescence centres (also crystal defects) producing a photon of light per recombination. The number of photons emitted is therefore a measure of the number of electrons trapped and thus a measure of the dose built up in the sediment.

The use of a partial bleach methodology avoids any potential overbleaching (bleaching of luminescence signal beyond that at deposition) which may result if the total bleach or additive dose methodology was applied (cf. fine-grained experiments of Fuller *et al.* 1994). The light spectrum used in the partial bleaching of this Spanish alluvium was restricted. Berger & Luternauer (1987) indicate the solar spectrum to be severely attenuated below 500 nm and above 690 nm in turbid water akin to that in a river. To avoid overbleaching in the laboratory by using wavelengths not necessarily experienced during transport or at deposition, Berger (1988, 1990) suggests that only wavelengths >550 nm should be used for optical bleaching. A 500–690 nm window is used in this study to construct partial bleach curves (Fig. 2). Bleaching wavelengths were restricted using a Schott BG39 and Corning 3-67 filter in combination. The power of bleaching beneath a solar simulator (SOL2) was further reduced by adding a Kodak-Wratten 0.9ND neutral density filter, achieving a bleaching power of 0.35 mW cm^{-2} in a SOL2.

IRSL measurements were made using a Daybreak 2000 manual IRSL system, detecting signal transmitted through a Schott BG39 filter only (50% light transmission between 340 and 600 nm); a neutral density filter (0.15 ND) was used in signal detection for samples M8, M13, M14, M15 and C1 in which irradiated IRSL signals exceeded 300 000 counts/second (Daybreak photomultiplier tube detection limit). All sample aliquots underwent natural normalization (1.0 s IR stimulation) to remove the effects of inter-disc variation in natural IRSL signal. A 1.0 s IR stimulation using the Daybreak 2000 system causes a less than 1% reduction in signal. Irradiation doses applied were dependent upon sample sensitivity to doses given in preliminary tests;

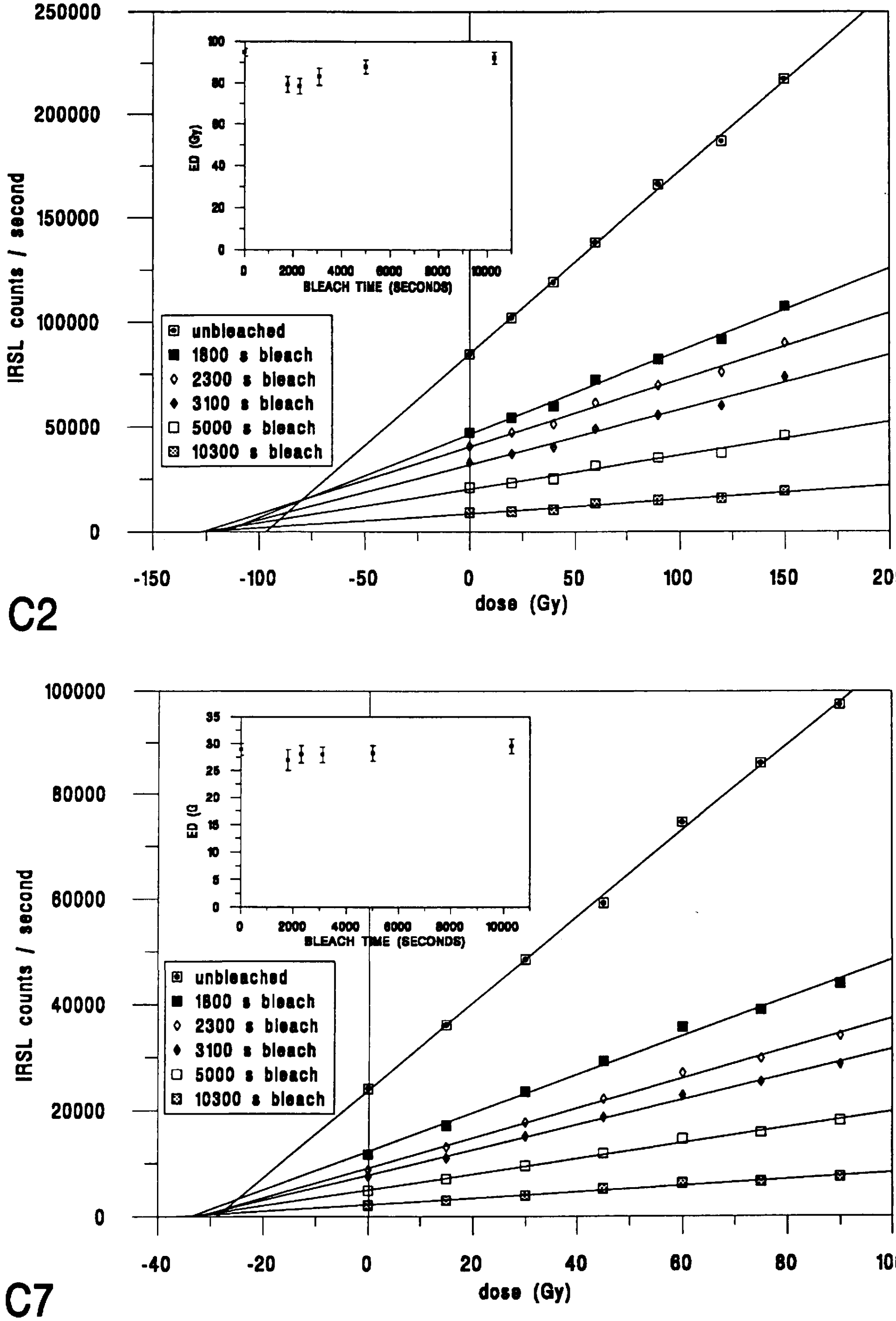

Fig. 2. ED determinations: examples of partial bleach methodology are given (C2 and C7). ED variation with bleach time (where appropriate) is shown in the inset graph. Note: these graphs are illustrative, displaying simple curves fitted to mean values. Actual ED calculation used Grün's software (pers. comm. 1991) in which the growth curve is a best fit line accounting for scatter between values.

maximum beta doses, delivered using a Daybreak irradiator at a $^{90}Sr/^{90}Y$ source strength of 3.8 Gy/minute for coarse grains (> 100 μm), varied from 30 to 800 Gy. Two representative sample types (C2 and C7) give an indication of the dose regimes applied (Fig. 2). A preheat of 140°C of 62 hour duration was used, removing thermally unstable signal induced following irradiation. Each sample was then stored for 24 hours before measurement (1.0 s IR stimulation) – partial bleach cycles using nine partial bleach times (180, 500, 900, 1300, 1800, 2300, 3100, 5300 and 10 300 seconds duration) reducing IRSL at *c.* 10% intervals until *c.* 90% of original (subsequent to preheat) IRSL was removed. The five longest bleaches (≥50% IRSL reduction) were considered in ED calculation (Fig. 2); the EDs determined using shorter bleaches do not give consistent results (Fuller *et al.* 1994). In most cases, the EDs for each sample fall within errors of each other and a mean value is calculated (cf. Table 1). However, for sample C2, the additive dose ED (intersection with x-axis) is appreciably higher than that of the first three partial bleach EDs shown (Fig. 2). This may be used as a maximum ED, but suggests ED overestimation which the partial bleach methodology seeks to avoid; consequently, in sample C2 and other samples displaying this ED distribution, the first three EDs are used to determine the mean. An additive dose methodological approach was adopted for older samples with exponential growth curves, given the high degree of error in defining the intersect of two exponential curves. A distinction is made between those samples displaying linear growth and those displaying exponential growth (Table 1).

Table 1. *Additive dose and partial bleach ED values (Gy) for samples collected from the Rio Guadalope*

Bleaching time (seconds)	0 (AD)	1800	2300	3100	5300	10 300
Mas de Las Matas study reach						
M2[l]	1.97 ± 0.28*	1.29 ± 0.37*	1.36 ± 0.31*	1.37 ± 0.36*	1.51 ± 0.35*	
M4[l]	8.26 ± 0.30	6.10 ± 0.45*	6.15 ± 0.39*	6.51 ± 0.42*	6.97 ± 0.41	7.35 ± 0.41
M5[l]	11.31 ± 1.11	8.10 ± 1.47*	8.34 ± 1.49*			10.06 ± 1.25
M6[l]	75.43 ± 1.26	65.12 ± 2.25*	68.45 ± 21.16*	70.77 ± 5.19*	71.18 ± 1.66	73.31 ± 1.51
M7[l]	54.47 ± 1.60*	51.89 ± 2.37*	51.00 ± 2.10*	51.38 ± 2.17*	51.54 ± 2.08*	53.26 ± 2.04*
M8[e]	147.25 ± 18.68*					
M9[l]	21.78 ± 1.74*	20.31 ± 3.03*	21.43 ± 2.93*		20.34 ± 2.25*	
M11[l]	69.70 ± 1.99*					
M12[l]	21.99 ± 0.50*	21.04 ± 1.57*	19.33 ± 1.52*	20.15 ± 1.83*	20.82 ± 1.20*	21.03 ± 1.11*
M13[e]	277.45 ± 4.14*					
M14[e]	462.84 ± 107.52*	338.80 ± 91.04	349.44 ± 90.49	374.90 ± 96.50	410.07 ± 100.59	444.32 ± 107.29
M15[e]	293.50 ± 16.39*	229.23 ± 29.81	249.19 ± 44.98	256.52 ± 45.98	275.60 ± 34.48	286.17 ± 29.30
Castelserás study reach						
C1[e]	303.62 ± 8.63*	267.71 ± 16.37	266.24 ± 15.18	260.50 ± 15.16	289.85 ± 13.84	287.36 ± 10.47
C2[l]	94.72 ± 1.78	79.15 ± 3.85*	78.36 ± 3.75*	83.00 ± 4.12*	87.75 ± 3.39	92.00 ± 2.77
C4[l]	54.54 ± 2.32*	49.23 ± 2.82*	52.74 ± 3.79*	52.19 ± 3.37*	53.26 ± 3.03*	53.56 ± 2.47*
C5[l]	49.29 ± 3.31*	43.79 ± 3.90*			49.82 ± 5.14*	
C6[l]	28.15 ± 0.92	24.90 ± 1.10*	25.74 ± 1.17*	26.01 ± 1.11*	27.22 ± 1.23	25.95 ± 1.89
C7[l]	28.93 ± 1.07*	26.95 ± 1.91*	28.02 ± 1.61*	27.97 ± 1.47*	28.19 ± 1.43*	29.53 ± 1.32*
C8[l]	34.52 ± 0.94*	33.49 ± 1.46*	33.33 ± 1.42	33.74 ± 1.53*	33.37 ± 1.56*	33.94 ± 1.01*
C9[l]	67.88 ± 1.60*					

[l] Denotes linear growth curve; [e] denotes exponential growth curve.
* EDs used (where multiple, value in Table 2 represents mean of those denoted).
AD indicates ED obtained by additive dose method.

Dosimetry

The total dose rate is a function of the sample's radioactivity and it is determined by measuring the constituent external α, β and γ dose rates and internal β dose rate. U and Th (and their daughters) emit α, and β particles and γ rays accompanying decay; ^{40}K emits β particles and γ rays accompanying decay. Measurement of the α activity of the sample in a thick source alpha counter (TSAC) provides a measure of the U and Th contents within the sediment, from which the external α dose rate can be calculated. Beta and γ dose rates are measured directly in the laboratory and field, using a thick source beta counter (TSBC) (Sanderson 1988) and γ spectrometer respectively. Internal dose rates (radioactivity from within the grain) are measured by determining K content. Cosmic ray dose rates are estimated using calculations from Prescott & Hutton (1988). Water content (important in the attenuation of radioactivity in the sediment) is estimated based on water content at the time of sample collection (close to zero) and at saturation (measured in the laboratory). Grain size is also considered in the dose rate calculations, being influential in radiation attenuation. The alpha efficiency value is taken as that used in coarse grain feldspar dating (0.2, cf. Mejdahl 1987). A summary of dose rate data is given in Table 2.

IRSL age correction

For samples giving luminescence ages over 100 ka, it has been suggested that the age needs to be corrected to take account of long term loss of luminescence signal (Mejdahl 1988; Duller 1994). In this study, correction is applied using the formula:

$$T = \tau(1 - e^{-1/\tau})$$

where T is luminescence age derived from the ED and dose rate, τ is the mean life (i.e. lifetime of decay), and t is the geological age. For the current study τ is assumed to be 1 million years, as implied for feldspars from Australia (Huntley *et al.* 1993). No correction for the effect of shallow traps (as reported by Mejdahl & Christiansen 1994) was applied. For the four oldest samples, the corrected ages are given in the final column of Table 3.

Alluvial morphologies

Mas de Las Matas

A total of 20 terrace surfaces has been identified in the Mas de Las Matas study reach. Six of these relate to the Rio Bergantes, and 14 have been identified in the Rio Guadalope. Heights above river channel vary from 43 m (MT1) to 3 m (MT14) with MT1 standing *c.* 15 m above the next highest surface, MT2. Correlations between the Guadalope (MT) and Bergantes (B) terraces may be made on the basis of height above modern river level, allowing for variation in gradients between the two systems. It appears that terrace B6 correlates with MT13, B5 with MT10, B4 with MT9, B3 with MT8 or MT7, B2 with MT6 and B1 with MT5.

Terraces MT13 (5 m), MT9 (7.5 m) and MT6 (15 m) are the most extensive surfaces and can be traced throughout much of the reach with both the high and low

Table 2. *Summary of dose rate data (after attenuation)*

Sample	Sample depth (m)	Grain size (μm)	Water content (%)	Internal β dose rate (μGy a^{-1})	External α dose rate (μGy a^{-1})	External β dose rate (μGy a)	External γ dose rate (μGy a^{-1})	Cosmic ray dose rate (μGy a^{-1})	Total attenuated dose rate (μGy a^{-1})
Mas de Las Matas									
M2	0.75	125–212	10 ± 5	525 ± 163	369 ± 202	1405 ± 138	815 ± 83	191 ± 19	3305 ± 313
M4	1.1	180–212	10 ± 5	588 ± 48	90 ± 47	552 ± 47	470 ± 43	182 ± 18	1882 ± 113
M5	1.6	180–212	10 ± 5	653 ± 53	145 ± 77	850 ± 81	369 ± 32	171 ± 17	2188 ± 146
M6	0.8	125–212	10 ± 5	568 ± 143	328 ± 180	923 ± 87	517 ± 48	190 ± 19	2526 ± 262
M7	0.4	125–150	17 ± 9	534 ± 50	510 ± 271	1030 ± 128	622 ± 81	199 ± 20	2895 ± 380
M8	1.3	180–212	10 ± 5	680 ± 55	274 ± 123	1001 ± 87	584 ± 560	178 ± 18	2717 ± 217
M9	1.2	125–212	10 ± 5	507 ± 127	332 ± 183	1128 ± 99	702 ± 70	180 ± 18	2849 ± 272
M11	0.8	125–212	10 ± 5	563 ± 142	309 ± 172	1175 ± 97	607 ± 59	190 ± 19	2844 ± 264
M12	1.1	180–212	14 ± 7	685 ± 56	138 ± 73	748 ± 81	466 ± 50	182 ± 18	2219 ± 167
M13	3.1	180–212	11 ± 5	653 ± 53	106 ± 56	379 ± 38	224 ± 16	143 ± 14	1505 ± 98
M14	0.9	180–212	12 ± 6	702 ± 57	126 ± 67	747 ± 68	307 ± 27	187 ± 19	2069 ± 137
M15	2.6	150–212	10 ± 5	705 ± 117	195 ± 106	797 ± 62	339 ± 28	152 ± 15	2188 ± 181
Castelserás									
C1	2.1	180–212	5 ± 5	593 ± 48	163 ± 86	658 ± 64	322 ± 27	161 ± 16	1897 ± 137
C2	0.2	180–212	14 ± 7	653 ± 53	89 ± 47	346 ± 50	208 ± 17	205 ± 20	1500 ± 103
C4	1.6	180–212	7 ± 7	740 ± 60	119 ± 64	594 ± 106	367 ± 36	171 ± 17	1991 ± 163
C5	0.5	125–150	19 ± 9	437 ± 50	113 ± 62	1064 ± 151	642 ± 90	196 ± 20	2452 ± 262
C6	0.6	180–212	13 ± 7	631 ± 42	84 ± 45	517 ± 53	280 ± 25	194 ± 19	1706 ± 111
C7	1.6	150–212	14 ± 7	559 ± 93	155 ± 84	826 ± 92	408 ± 42	171 ± 17	2120 ± 187
C8	1.7	125–150	15 ± 8	456 ± 41	266 ± 141	1062 ± 125	426 ± 48	169 ± 17	2379 ± 246
C9	1.3	180–212	15 ± 8	658 ± 53	72 ± 38	404 ± 61	276 ± 27	178 ± 18	1589 ± 115

Table 3. *Summary of estimated IRSL ages*

Sample	Terrace unit	Grain size (μm)	ED (Gy)	Dose rate (μGy a^{-1})	IRSL age (a)	Corrected IRSL age (ka)
Mas de Las Matas						
M2	MT14	125–212	1.34 ± 0.35	3305 ± 313	405 ± 112	–
M4	B6/MT13	180–212	6.25 ± 0.42	1882 ± 113	3 320 ± 229	–
M5	B6/MT13	180–212	8.22 ± 1.48	2188 ± 146	3 756 ± 721	–
M6	B5/MT10	125–212	68.11 ± 3.2	2526 ± 262	26 964 ± 3 071	–
M7	B4/MT9	125–150	52.26 ± 2.06	2895 ± 380	18 052 ± 2 749	–
M8	MT5	180–212	147.25 ± 18.68	2717 ± 217	54 195 ± 8 121	–
M9*	MT7	125–212	20.97 ± 2.49	2849 ± 272	7 360 ± 1 121	–
M11	MT7	125–212	69.70 ± 1.99	2844 ± 264	24 508 ± 2 380	–
M12	MT11	180–212	20.73 ± 1.29	2219 ± 167	9 342 ± 940	–
M13	MT2	180–212	277.45 ± 44.14	1505 ± 98	184 352 ± 31 698	204^{+40}_{-38}
M14	MT1	180–212	462.84 ± 107.52	2069 ± 137	223 702 ± 54 034	253^{+72}_{-67}
M15	MT2	150–212	293.50 ± 16.39	2188 ± 181	134 141 ± 13 389	144^{+16}_{-15}
Castelserás						
C1	CT6/fan	180–212	303.62 ± 8.63	1897 ± 137	160 054 ± 12 423	174^{+15}_{-14}
C2	CT7	180–212	80.17 ± 3.91	1000 ± 103	53 446 ± 4 501	–
C4	CT9	180–212	52.59 ± 2.97	1991 ± 163	26 414 ± 2 456	–
C5	CT11	125–150	47.63 ± 12.35	2452 ± 262	19 425 ± 5 447	–
C6*	CT8	180–212	25.55 ± 1.13	1706 ± 111	14 976 ± 1 178	–
C7*	CT8	150–212	28.27 ± 1.47	2120 ± 187	13 335 ± 1 365	–
C8*	CT6	125–150	33.73 ± 1.32	2379 ± 246	14 178 ± 1 567	–
C9	CT7	180–212	67.88 ± 1.60	1589 ± 115	42 718 ± 3 253	–

* Tributary fan material overlying terrace surface.

terrace surfaces being fragmentary. Small fans have developed at the mouth of tributary valleys, upbuilding existing terrace surfaces and the highest terrace surfaces (MT1 and MT2) may well consist partly of fan material, situated, as they are, close to a large west-bank tributary. A large NW–SE-trending palaeochannel is preserved in the surface of MT5; its position and size suggests it is a proto-Bergantes indicating that the Guadalope–Bergantes confluence has moved *c.* 3 km downstream since MT5 times.

Tributary valleys in the Mas de Las Matas reach generally grade to the modern channel. This contrasts with the Guadalope at the Castelserás study reach, where trunk stream incision has outstripped that of the tributaries, leaving many hanging (Macklin & Passmore 1995).

Castelserás

Eleven terrace surfaces have been previously identified in the Castelserás reach by Macklin & Passmore (1995), of which six (at 4 9, 12, 17.5, 19 and 23 m above present river level) can be traced throughout the reach. Five higher terraces (29, 34.5, 45, 56 and 81 m above river level) occur only at the northern end of the basin. The most extensive units are CT6 (23 m), CT7 (19 m) and CT8 (17.5 m). Prominent relict alluvial fans are developed at the mouths of a number of tributary streams and grade to terrace units CT6 or CT7. Tributary alluvial fan deposition slowed shortly after CT8 formation and as the result cutting of barrancas followed. Incision of the Guadalope has been more rapid than its tributaries, many tributaries have been left hanging above the trunk river level. The lower terraces CT9 (12 m) and CT10 (9 m) are less horizontally extensive than the earlier fills and lie (partly or entirely) within a bedrock trench 150–500 m wide.

Alluvial sedimentology

Alluvial deposits at Castelserás (described by Macklin and Passmore, 1995) and Mas de Las Matas, on the basis of their architecture and sedimentology fall into one of four major lithofacies types.

(1) Terrace units characterized by numerous multi-storey lenticular or tabular intersecting sheets of clast supported, matrix rich gravels with minor sand and silt belts and channel fills. Such sediment packages are typical of aggrading, low sinuosity braided river environments (Ori 1982; Miall 1985) and is the most common architectural type in terrace units at both Mas de Las Matas and Castelserás.

(2) Terrace units characterized by horizontally extensive inclined heterolithic stratification (IHS) (Thomas *et al.* 1987) and a thick associated fine member. Such features are well developed in CT8 and are consistent with meander lobe development (Campbell & Hendry 1987). In CT9, MT11 and MT14, a lower gravel member is not exposed, but thick upper fine members suggest vertical accretion overbank in a floodplain environment.

(3) Terrace units where the lower part of the sequence is dominated by IHS, more typical of a high sinuosity river system; while in the upper part cut and fill multi-channel system features predominate.

(4) Terrace units characterized by horizontally continuous, relatively thin (<1 m), flat-bedded relatively angular gravel and minor sand sheets. This type of sequence is often found in a tributary–trunk stream location and represents alluvial fan sedimentation.

A notable difference occurs between the sedimentology of higher terrace units and the lower, more recent alluvial fills (MT11 to MT14 at Mas de Las Matas and CT9 to CT11 at Castelserás). The former are dominated by gravels, with very little fine material, whilst the latter are characterised by sands, with relatively little gravel. This suggests a change in the calibre of sediment delivered to the river, which may relate to the stripping of catchment slopes, exhausting coarse sediment supply, enhancing stream competence and trunk stream-tributary coupling.

Alluvial chronologies

Figure 3a shows a schematic cross section of CT6 and later terraces at Castelserás and Fig. 3b shows a cross section of the complete sequence of alluvial units in the Guadalope at Mas de Las Matas; IRSL ages refer to those in Table 3. Table 4 summarizes the alluvial stratigraphy of both reaches, correlations between

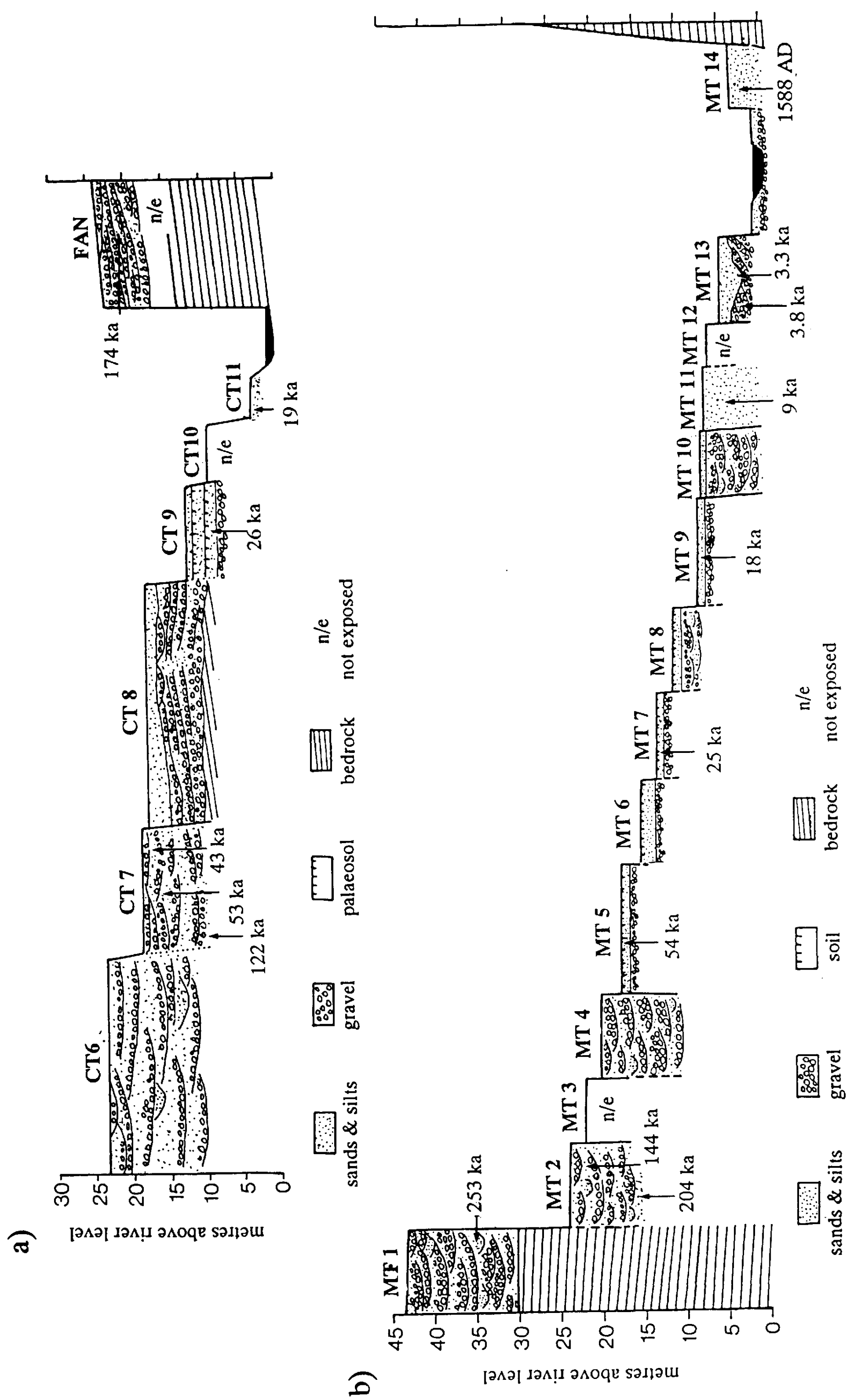

Fig. 3. Schematic cross-sections of the Guadalope valley at **(a)** Castelserás and **(b)** Mas de Las Matas, showing sedimentary architectures of Late Pleistocene river terrace units and tributary stream alluvial fan.

Table 4. *Alluvial records of the Rio Guadalope at Castelserás and Mas de Las Matas compared*

Castelserás					*Mas de Las Matas*				
Alluvial terrace unit	Height above modern river (m)	Maximum observed thickness (m)	IRSL age (ka)	Fluvial style	Alluvial terrace unit	Height above modern river (m)	Maximum observed thickness (m)	IRSL age (ka)	Fluvial style
CT1	81	1.5	–	Gravel-bed river					
CT2	56	Unknown	–	Unknown					
CT3	45	<10	–	Gravelly high-/low-sinuosity river	MT1	43	>10	253^{+72}_{-67}	Gravelly low-sinuosity river
CT4	34.5	Unknown	–	Unknown					
CT5	29	Unknown	–	Unknown					
CT6	23	>15	174^{+15}_{-14}	Gravelly low-sinuosity river	MT2	24	>7	204^{+40}_{-38} to 144^{+16}_{-15}	Gravelly low-sinuosity river
					MT3	22	Unknown	–	
CT7	19	10	122^{+19}_{-12} to 43 ± 3	Gravelly low-sinuosity river	MT4	20	10		Gravelly low-sinuosity river
					MT5	17.5	>2	54 ± 8	Unknown
CT8	17.5	11	–	Gravelly high-sinuosity river					
					MT6	15	4	–	Gravelly low-sinuosity river
CT9	12	>4	26 ± 2	Unknown	MT7	13	>2	25 ± 2	Gravelly low-sinuosity river
CT10	9	Unknown	–	Unknown	MT8	11	3	–	Gravelly low-sinuosity river
CT11	4	Unknown	19 ± 5	Unknown	MT9	8	3	18 ± 3	Gravelly low-sinuosity river
Fans	17–23	Unknown	15 ± 1 to 13 ± 1	Alluvial fan	MT10	7.5	8	–	Gravelly low-sinuosity river
					MT11	7	>2	9 ± 1	Sandy high sinuosity river
					MT12	6.5	Unknown	–	Unknown
					MT13	5	>3	3.8 ± 0.7 to 3.3 ± 0.2	Gravelly low-/high sinuosity river
					MT14	3	1	0.4 ± 0.1	Sandy high sinuosity river

Castelserás and Mas de Las Matas terraces are made on the basis of height above river level of the terrace surfaces and IRSL ages (where available) of subadjacent units.

The earliest alluvial units are CT1–5 in the Castelserás reach near Alcañiz. The surface of these terraces lie between 29 and 81 m above present river level. No datable material was available from these units. Although it is not possible to produce a precise estimate of age for these highest terraces, comparison with the dated MT1 (43 m) at Mas de Las Matas (probably correlative with CT3 (45 m) at Castelserás) indicates these were probably deposited before 253^{+72}_{-67} ka BP and are of Mid- and Early Pleistocene age (Fig. 4). Freeman (1975) and Alonso & Garzón (1994) have mapped units in the Rio Jarama (Tagus Basin) at equivalent heights to those at Castelserás (85 m, 60–70 m and 40 m). Terraces lying at 40 m above river level have also been identified in the Júcar, Serpis and Turia rivers by Payá & Walker (1986). The IRSL age estimate of MT1 at Mas de Las Matas suggests aggradation towards the end of (oxygen isotope) stage 8 although the error limits could shift this anywhere between mid stage 6 and stage 9. If we accept 253 ka, this aggradation may have occurred in the transition between an interglacial period and a globally cold, glacial phase in climate, according to oxygen isotope records (Martinson *et al.* 1987) and ice core records (e.g. Jouzel *et al.* 1993).

Incision of MT1 (Mas de Las Matas) occurred at some point within stage 7. Aggradation of MT2 may have been initiated at the end of stage 7/early stage 6 (204^{+40}_{-38} ka) and continued until at least 144^{+16}_{-15} ka, (IRSL age at the top of MT2 (24 m)) throughout a glacial period (stage 6). An IRSL age of 174^{+15}_{-14} ka within a fan unit which grades to CT6 (23 m) also indicates extensive aggradation during a glacial phase (stage 6) in the Castelserás reach, with tributary valleys delivering large quantities of sediment to the trunk river. CT4 and CT5 at Castelserás have no height correlatives preserved in Mas de Las Matas, with equivalent units probably having been reworked and eroded explaining the large height difference between MT1 (43 m) and MT2 (24 m).

Aggradation of MT2 ceased some time after 144 ka (Table 4, Fig. 4). An additional terrace is present at Mas de Las Matas, MT3 (22 m), is bracketed between 144 and 122 ka, deposited in the transitional climate between stage 6 and substage 5e. The terrace unit may have been removed from Castelserás through subsequent reworking, or may represent a greater complexity of alluvial response to climate change in a confluence zone.

Assuming CT7 and MT4 to be correlatives (19 m and 20 m respectively above river level), then aggradation in both reaches began at 122^{+19}_{-12} ka towards the end of the thermal maximum at substage 5e (122 ka). Palaeoclimate reconstructions from marine pollen records in the Adriatic Sea (Turon 1984) and mires and lakes in France (Pons *et al.* 1992) show a marked decline in arboreal pollen at the 5d–e boundary. Early substage 5d is characterized by low mean temperature and high total precipitation. At the Cova Negra Palaeolithic site, Valencia, a period of cooler and much wetter climate is also evident, dated to 117 ka (corrected to 124 ka) (Prószynska-Bordas *et al.* 1992). Two further IRSL age estimates obtained from CT7 give 53 ± 5 ka and 43 ± 3 ka for samples collected at the middle and at the top of CT7 respectively. However, there is evidence for a greater instability in the alluvial system at Mas de Las Matas. The top of MT5 has an IRSL age of 54 ± 8 ka, indicating that while CT7 was aggrading downstream, MT4 at Mas de Las Matas had incised and

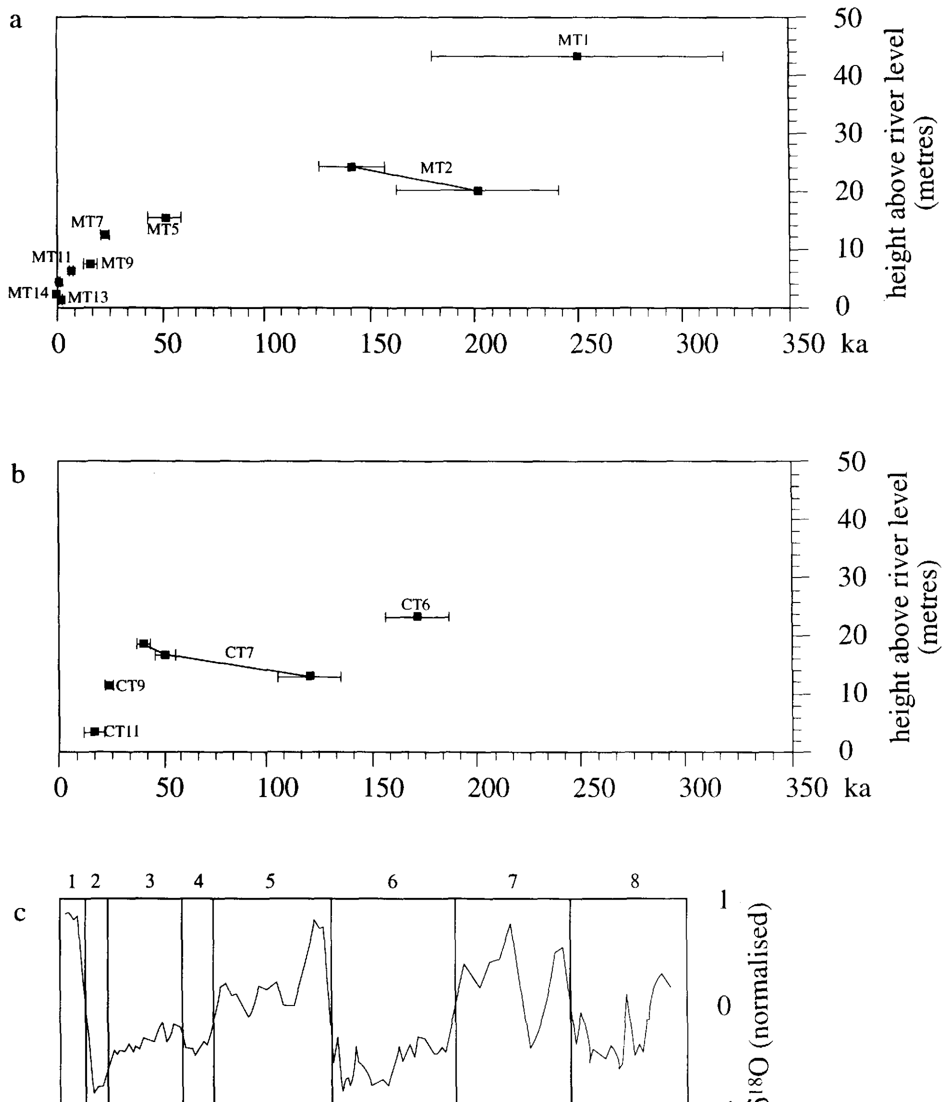

Fig. 4. A geochronology of alluviation in the Guadalope basin at (**a**) Mas de Las Matas and (**b**) Castelserás, compared with (**c**) the marine oxygen isotope stratigraphy (after Martinson *et al.* 1987).

filled (MT5). MT5 aggradation may relate to stage 4–3 transition, with cutting occurring soon after 54 ka (sample taken from capping material), in early stage 3. MT4 was probably incised sometime between 122 and 54 ka.

Trenching of CT7 at Castelserás, however, can be more tightly constrained, occurring after 43 ka. Subsequent infilling of the valley floor occurred (CT8) to 17.5 m, which has a similar height to MT6 (15 m) at Mas de Las Matas. Both CT8 and MT6 are at present undated, but are bracketed between units for which there is age control. MT6 was deposited and incised between 53–25 ka and CT8 between 43–26 ka, both probably in stage 3. CT8 represents a change in fluvial style from that of a braided river to an actively migrating coarse grained wandering river overlying a bedrock strath (Macklin & Passmore 1995). The predominance of lateral erosion suggests the Rio Guadalope is close to either static or dynamic equilibrium (cf. Bull 1991). In the Castelserás reach, all major tributary fans lie above or are truncated by CT8, suggesting very little coarse sediment was being delivered to the trunk stream at this time by these tributaries. This would suggest reduced catchment yields of coarse sediment and this may account for the difference in fluvial style.

This change appears to have occurred after 43 ka. Both units (CT8 and MT6) deposited after this period have thick upper fine members. This suggests a change in fine-sediment delivery rates which may relate to a decline in precipitation during this period which is recorded elsewhere in western Europe in the Les Echets and Grand Pile pollen records (Guiot *et al.* 1989). An increase in aridification suggested by these proxy records would have reduced stream competence, leading to an alluvial system increasingly dominated by fine sediment. Stage 3 was characterized by frequent, small amplitude climate oscillations (Martinson *et al.* 1987) and in NW Europe, it may be divided into four phases of warm/wet conditions at 59–62 ka, 49–56 ka and 35–42 ka and 28–31 ka, with a cold and arid phase centred on 45 ka (Baker *et al.* 1993). It is uncertain whether these can apply to southern Europe, given the strong climatic gradient suggested by Rousseau and Puissegúr (1990), however, Pérez-Obiol & Julià (1994) identified a clear interstadial event 30 to 27 ka BP from Lake Banyoles in northeastern Spain.

The upper fine member of CT9 (12 m) at Castelserás has been dated as 28 ± 4 ka by Durham Luminescence Research Laboratory (reported in Macklin & Passmore 1995) and in this study to 26 ± 2 ka (IRSL age estimates), equating to the end of the interstadial event identified in Lake Banyoles (Pérez-Obiol & Julià 1994). An IRSL age of 25 ± 2 ka has been obtained from a fine member capping MT7 (13 m) gravels, from which artefacts have also been recovered. This corresponds (within error), with the age of CT9.

Trenching of MT7 occurred after 24 ka, during stage 2 the next stage of alluviation represented by MT8 is undated, but can be bracketed to sometime between 24 ka and 18 ± 3 ka (MT9). It most probably correlates with CT10, which is bracketed between 26 ka and 19 ka (CT11). MT9, dated at 18 ka, therefore correlates with the lowest terrace mapped in the Guadalope at Castelserás, CT11, which has an IRSL age of 19 ± 5 ka. Stage 2 appears to represent an intensive period of fluvial activity in the Guadalope with at least three phases of alluviation and incision, whereas a single cold stage aggradation has been suggested by previous studies in the region (Payá & Walker 1986). The complex nature of climate change during the Last Glacial Maximum is perhaps reflected in the Guadalope record.

At Castelserás a series of fan deposits overlying CT6 and CT8 have IRSL age estimates of 15–13 ka indicating significant tributary delivery of sediment during the Older Dryas stadial. At Mas de Las Matas, alluviation occurred between 18 ka and 9 ka (MT10) (Table 4), but at present no alluvial units of this age have been recognised or dated in the Castelserás reach.

MT11 has been dated to the early Holocene with an IRSL age of 9 ± 1 ka. Incision occurred after 9 ka and was followed by a major period of alluviation at Mas de Las Matas (MT13) dated to 3.8 ± 0.7 ka (at the base) and 3.3 ± 0.2 ka (towards the top), with up to 3 m of aggradation in 500 years. In addition there are two minor alluvial fills dated to between 9 and 3.8 ka (MT12) and 405 years (MT14).

Discussion

Luminescence dating of alluvial units in the Guadalope basin suggests major aggradation phases in the Mid-Pleistocene occurred during periods of climate transition between interglacials and glacials. This is suggested in terrace units CT7 and MT2, where aggradation began during the climatic deterioration preceding substages 5e and 7a respectively. (Fig. 4). A similar pattern is also evident in MT1, although aggradation is dated to a phase of improving climate between stage 8 and stage 7. The deposition of MT7 and CT9 at 25 and 26 ka, respectively, coincides with a period of climate transition shown by the GRIP oxygen isotope record (Fig. 5). The Lake Banyoles pollen record (Pérez-Obiol and Julià 1994) shows this was also a wetter phase in northern Spain. Aggradation in the Guadalope basin at *c.* 19 ka (MT9 and CT11) relates to a short cooling trend at 20 ka, associated with a Heinrich event (H2) in the North Atlantic (Bond *et al.* 1993). Deposition of MT11 appears to date to climate amelioration in the early Holocene. In the absence of detailed regional land-use or climate records for the later Holocene, at present it is not possible to attribute alluviation at 3.8–3.3 ka and 0.4 ka to climatic or anthropogenic causes. However, introduction of agriculture, destabilizing catchment soils, together with suitable climate conditions providing abundant runoff, will readily supply sediment to the fluvial system prompting aggradation. MT12 may correspond with the Cerezuela unit in the Rio Regallo which is dated (by OSL and radiocarbon dating) to 5840–4780 years BP (Macklin *et al.* 1994). Between 3 ka and 2.3 ka, 8–9 m of aggradation has been found in the Rio Jarama (Alonso & Garzón 1994). This event is of a similar magnitude, aggrading up to 9 m in 700 years, compared with 3 m in 500 years in the Guadalope. The sedimentology of these comparable events is also similar, consisting of basal gravels overlain by upwardly fining sands and silts, suggesting a similar cause, probably a meandering system. No evidence for equivalent incision is found in the Guadalope, but a correlative incisive period was probably responsible for cutting into MT11 or MT12 in the early Holocene. MT14 is a historic fill. The terrace is limited in extent and the sediment dated represents a flood unit, giving an age of 405 ± 112 years (AD 1476–1700). Deposition may be associated with the Little Ice Age climate deterioration. Incision of 2–3 m occurred in the Rio Jarama sometime between 390 ± 80 radiocarbon years (AD 1407–1663 cal. (2σ)) and the present (Alonso & Garzón 1994), equivalent (within errors) to incision in the Guadalope following deposition of MT14.

Trunk stream alluviation in the Guadalope basin does not appear confined to arid, cold stages, as has been suggested by other workers in Spain investigating less well dated alluvial sequences (e.g. Payá & Walker 1986), rather it seems to have been initiated by climate deterioration from interglacial or interstadial conditions. A mechanism for this response may be a southward shift in the jet stream, a consequence of which would be an enhanced precipitation delivery to the Mediterranean during the winter (Prentice *et al.* 1992) and higher rates of catchment erosion and runoff. The proximity of the Guadalope basin to the North Atlantic may be critical in explaining the pattern observed. The GRIP ice core suggests Atlantic temperature to have been highly variable in the Late Pleistocene, linked, primarily, to Heinrich events which

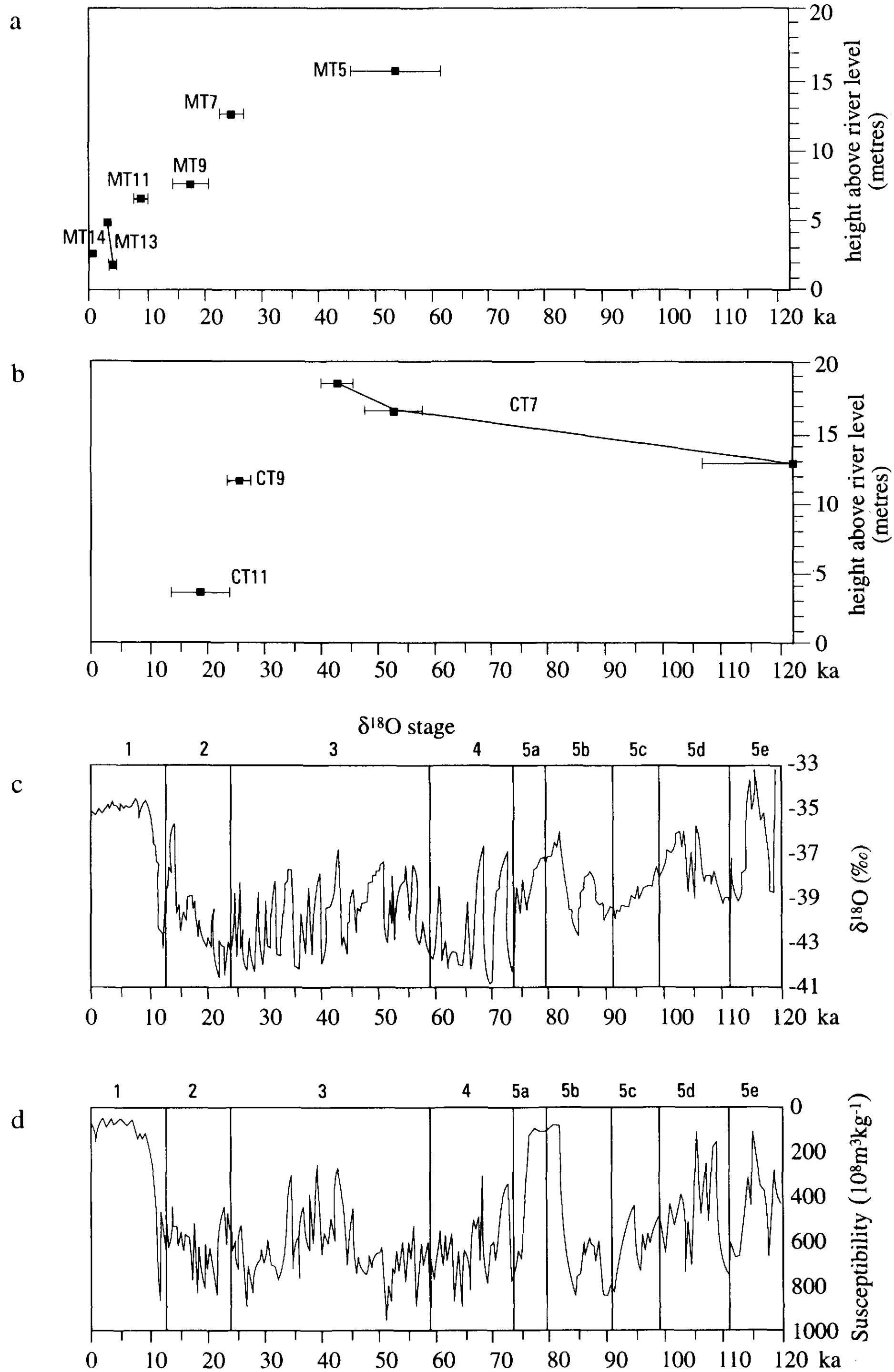

Fig. 5. A geochronology of alluviation in the Guadalope basin at (**a**) Mas de Las Matas and (**b**) Castelserás, compared with (**c**) the oxygen isotope stratigraphy in the high resolution GRIP ice core and (**d**) the Lac du Bouchet magnetic susceptibility record (after Thouveny *et al.* 1994).

cause a sudden cooling, although subsequent warming is also rapid. The role of thermohaline circulation is thought to be critical in forcing these temperature shifts, whereby heat advection from the tropics is increased in response to a reduced meltwater flux following ice sheet collapse (Paillard & Labeyrie 1994). Therefore, even during general cooling, increased evaporation from the North Atlantic may be sufficient to deliver additional precipitation to the northern parts of the Iberian peninsula. This would be likely to promote aggradation, by causing major erosion of catchment slopes unprotected by vegetation which may be unable to respond to such short-term ocean-driven climate pulses.

In summary, it is proposed that major valley alluviation in the Guadalope basin tends to be associated with transitional phases of climate, during the last interglacial–glacial cycle and possibly in earlier periods of the Pleistocene. This contrasts with interpretations of Pleistocene alluvial sequences elsewhere in Spain and Europe, notably the Thames and Rhine in which gravel aggradation episodes are apparently confined to cold stages (cf. Brunnacker *et al.* 1982; Bridgland 1988; Gibbard 1989). However, a lack of detailed chronometric control and the fragmentary nature of the alluvial deposits in these systems may conceal responses similar to those identified in the Guadalope catchment.

Conclusions

Geomorphological investigations of Quaternary fluvial deposits in the Guadalope valley, northeast Spain, have revealed a record of alluvial response to long-term environmental change during the Mid- and Late Pleistocene, Holocene and historic periods. Eleven alluvial terraces, ranging from 4 to 81 m above river level, have been identified along a 13 km long reach of the Rio Guadalope south of Alcañiz. Twenty surfaces have been identified in a 4 km long confluence zone with the Rio Bergantes, ranging from 3 to 43 m above river level. Luminescence dating suggests that deposition of these alluvial units dates back to the Mid-Pleistocene.

The earliest fluvial unit is dated to the Mid-Pleistocene (probably stage 8 glacial). A complex sequence of large scale cutting and filling occurs up to stage 5. A major phase of aggradation from substage 5e to early stage 3 can be identified at Castelserás, whilst at Mas de Las Matas during this period two cut and fill episodes may be identified. Following 43 ka, four episodes of cutting and filling are documented in both reaches during the latter part of stage 3 and into stage 2. Four Holocene alluvial units are evident: the oldest deposited at 9 ka, a second unit dated before 4 ka and a third to 3.8–3.3 ka; a phase of minor alluviation is dated to the sixteenth century.

Region-wide aggradation/incision events can be tentatively established by comparing the Guadalope response with nearby catchments (notably the Jarama and Júcar) during both the Late Pleistocene and Holocene. However, the lack of dating control in the latter catchments does not enable detailed correlation between alluvial units to be made with the Guadalope at present.

Luminescence dating has built up a detailed geochronology of the alluvial response to climate change in the Rio Guadalope spanning the last 250 000 IRSL years. It has permitted, for the first time, a relatively detailed view of river response to environmental change in previously poorly documented periods in the Iberian Peninsula, notably during the last interglacial–glacial cycle, as well as providing a window into the Mid-Pleistocene.

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Magnitude and frequency of Holocene palaeofloods in the southwestern United States: A review and discussion of implications

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Abstract: Data about the magnitude and time of occurrence of palaeofloods from the lower Colorado River basin enable us to test two long-standing hypotheses which have affected many studies and applications in the field of flood hydrology. The two hypotheses are (a) the existence of an upper boundary to flood magnitudes and whether there is a possibility of determining it from the existing data, and (b) the random occurrence versus clustering of the large floods through time.

Earlier observations on regional flood envelope curves indicated the existence of an upper limit for flood magnitudes, but these studies limited their conclusions because of the short length of the systematic gauged data. This limitation is overcome here because palaeoflood data cover a much longer period of observation. Palaeoflood studies provide information about the largest individual floods experienced in many rivers in a specific region occurring over the last millennia. In the southwestern US, this information demonstrates that, even when the length of observational data increases to centuries and millennia, there is no change in the stabilized, regional envelope curves constructed from gauged and historical flood records. This pattern supports the hypothesis of an upper limit to flood magnitudes and points to a method for testing this hypothesis in other regions. There are surprising similarities between the envelope curve of the palaeoflood data and the envelope curve for the gauged and historical data in the lower Colorado River basin. These similarities indicate that in regions of the world where flood data is sparse envelop curves based on palaeoflood studies can provide basic data for engineering design purposes and other hydrological applications.

The random occurrence of large floods in time is tested by constructing chronologies for the largest palaeofloods in several basins in the lower Colorado River basin. These chronologies indicate a clustering of the large floods in specific time periods. The similarity between the various time periods characterized by high- and low-flooding and other palaeoclimatic indicators from the southwestern Untied States seems best explained by a climatic control on flood frequency over the last 5000 years.

Several assumptions have underpinned scientific thought in the research field of statistical flood hydrology for many years. Among those which are commonly used in hydrology are (a) the existence of an upper boundary to flood magnitudes, and

(b) the random (v. clustered) occurrence of large floods through time. The difficulties in resolving which and/or whether these concepts are valid arose from the limited lengths of the flood records. These records are usually short, in most cases shorter than 100 years. Here we overcome this limitation by adding regional palaeoflood data that span the last several centuries to millennia. This addition enables us to detect patterns in both the temporal distribution and the variations in the magnitudes of the largest floods. Agreement or disagreement of the regional flood records with the above assumptions will result in the acceptance or rejection of the assumptions. A limited number of regions in the world have sufficient palaeoflood information to apply this approach. By far the largest available palaeoflood data set from a single region is from the lower Colorado River basin in Arizona and southern Utah, which will be the focus of this review.

Palaeoflood hydrology – methodology and available data

Fine-grained flood deposits and other palaeostage indicators provide a long record of large floods (Baker 1987*a*). During large floods in stable bedrock canyons, fine-grained sand and silt fall rapidly out of suspension in areas of markedly reduced flow velocity, such as back-flooded tributaries and eddies at channel irregularities (Kochel & Baker 1982; Kochel *et al.* 1982; Baker 1987*a*). At some sites, stratigraphic records of multiple floods span several centuries or millennia, and the individual flood deposits can be distinguished through sedimentological criteria. Several types of information can be extracted from such field settings: (a) the minimum stage of the largest flood, (b) a chronology and a minimum stage estimates of other large floods, and (c) estimated magnitudes of the floods represented in the stratigraphic record. This last type of data can be extracted only if the setting and channel morphology are appropriate for hydraulic modeling, which is used to produce instantaneous peak discharge estimates from the minimum stage estimates derived from the heights of the deposits (Baker 1987*a*; O'Connor & Webb 1988). In this study we are interested in two aspects of the palaeoflood records. The discharge of the single largest flood in each basin will be compared with the largest floods in the modern instrumental record and used to address the question of the existence of an upper boundary to flood magnitudes in the region. Second, the chronologies of the largest floods in each basin are used to produce a regional data set of palaeofloods and examine the temporal distribution of large floods throughout the region. Information on both aspects from as many rivers as possible in the lower Colorado River basin is required to address these questions effectively. Figure 1 shows sites that are suitable for producing chronologies, sites suitable for indicating magnitudes of the largest floods, and sites that are useful for both applications.

Evidence of the largest floods is selectively preserved, deposits from smaller floods lie closer to the active river channel and are more likely to be removed by subsequent erosion (Ely & Baker 1990). The ages of the floods were largely determined by radiocarbon dating of associated organic or archaeological material (e.g., Enzel *et al.* 1994). These dates establish the chronology of the individual palaeoflood deposits preserved in a site and aid in determining the total length of the record chronicled by the deposits.

Palaeoflood discharge estimates are calculated by comparing the heights of flood palaeostage indicators with water-surface profiles calculated by the step-backwater

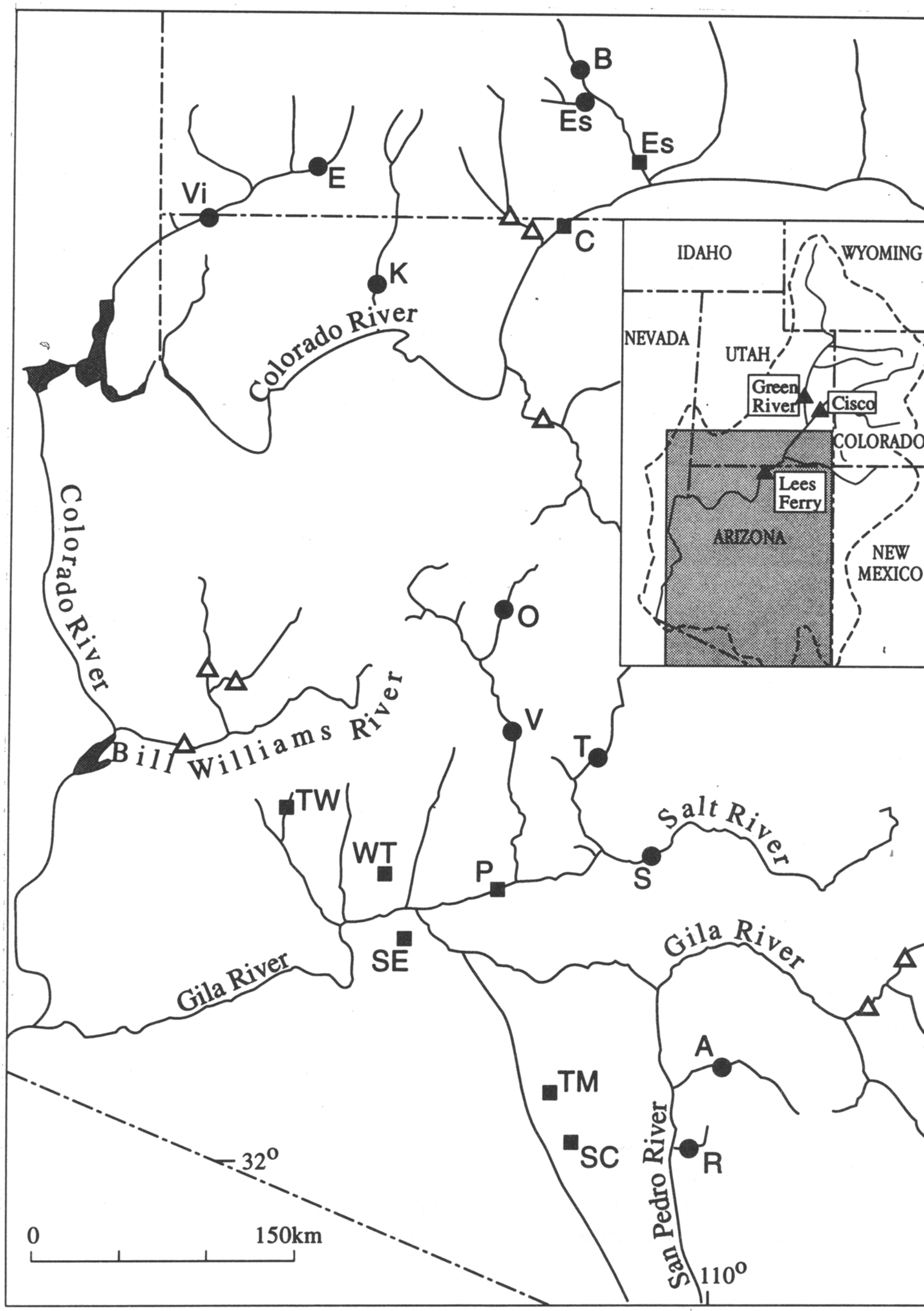

Fig. 1. Map showing the location of palaeoflood sites in Arizona and southern Utah, which yielded data (**a**) only for palaeoflood chronologies (open triangles); (**b**) only for maximum of peak discharges of palaeofloods (squares); and (**c**) for both chronologies and maximum peak discharges of palaeofloods (circles). The letters are in reference to Table 1. The inset map shows the entire Colorado River basin.

Table 1. *A summary of the single largest palaeoflood magnitude from several drainage basins in the lower Colorado River (Enzel* et al. *1993)*

River/site*	Code in Fig. 1	Drainage area (km^2)	Maximum peak discharge of palaeoflood (m^3s^{-1})	Length of record (years)	Data source
Colorado River, AZ	C	279 350	13 600–14 200	*c.* 4000	O'Connor *et al.* (1994)
Verde River, AZ	V	14 240	5 000	2000	Ely & Baker (1985); O'Connor *et al.* (1986*a*)
Salt River, AZ	S	11 150	4 100–4 600	2000	Partridge & Baker (1897); O'Connor *et al.* (1986*a*)
Salt River, AZ†	P	33 650	8 500–9 900 11 300–12 700	1000+	Fuller (1986)
Tonto Creek, AZ	T	1 630	800–1 000	*c.* 500	O'Connor *et al.* (1986*a*); Ely *et al.* (1988)
Aravaipa Creek, AZ	A	3 160	970	*c.* 900	Roberts (1987)
Redfield Creek AZ	R	285	350–400	*c.* 1000	Wohl (1989)
Oak Creek, AZ	O	1 213	1 350	≫100	Melis (1990); Melis (pers. comm.)
Virgin River					
East Fork, UT	E	840	800–850	1000+	Enzel, Ely & Webb (unpublished data)
Virgin River, AZ	Vi	10 306	1 700–1 900	1000+	Enzel *et al.* (1994)
Kanab Creek, UT	K	5 370	400–600‡	*c.* 500	Smith (1990); Smith (pers. comm)
Escalante River§, UT	Es	820	700–750		Webb (1985); Webb
	Es	1 900	1 250–1 550		and Baker (1987);
	Es	3 290	1 850–2 100	1000+	Webb *et al.* (1988)
	Es	4 430	860–940		
Boulder Creek, UT	B	450	350–450	500+	O'Connor *et al.* (1986*b*)
Tortolita Mts, AZ	TM				House (1991); House (unpublished data)
Cochie	C1	9.8	60–80	–	
Wild Burro	WB	11.1	120–150	–	House *et al.* (1991)
	WB	18.1	200–300		
Ruelas	RU	6.0	80–100	–	
Prospect	Pr	9.6	40–50	600+	
Cañada Agua	Ca	4.7	30–50		
White Tank Mts, AZ	WT	14.6	57–142	–	House (unpublished data); CH2MHill & French (1992)
Tiger Wash, AZ	TW	220	283–382	–	House (unpublished data), CH2MHill & French (1992)
Sierra Estrella, AZ	SE	2.8	21–29	–	House (unpublished data), CH2MHill & French (1992)

Table 1. *(continued)*

River/site*	Code in Fig. 1	Drainage area (km^2)	Maximum peak discharge of palaeoflood (m^3s^{-1})	Largest flood (years BP)	Data source
Santan Catalina Mts.	SC				Martinez-Goytre *et al.* (1994)
Cañada Del oro		80	110	–	
Sutherland		55	95	–	
Pima		10	55	–	
Tanque Verde		110	240	–	
Youtcy		24	50	–	
Buehman		103	270	–	
Edgar		74	130	–	
Alder		46	50	–	

* See Fig. 1 for locations. The palaeofloods are plotted in Figs 1 2, and 3 according to site code except for the Tortolita Mts. palaeoflood estimates which are represented in Fig. 1 as TM.
† Salt River downstream of the confluence of the Verde River; expanding flow can cause large overestimation.
‡ The actual largest flood was larger than reported by Smith (1990). He identified in the field evidence for a larger flood located in a channel reach which is difficult to model (Smith, pers. comm. 1991).
§ Different sites on the river.

method (O'Connor & Webb 1988). The elevation of a given deposit provides a minimum estimate for the peak stage of the associated flood, though the underestimation is not large. In many cases, the heights of the deposits closely approximate the actual stage of the flood peak (Ely & Baker 1985; O'Connor *et al.* 1986*a*; Baker 1987*a*; Partridge & Baker 1987; Kochel & Ritter 1987; Baker & Kochel 1988). Kochel (1980) estimated that deposit height was 10% less than actual water surface elevation and Webb (pers. comm. 1992) and Greenbaum (pers. comm. 1992) have documented silt lines and other high water marks 50 to 90 m higher than associated deposits. In several of the palaeoflood studies cited in Table 1 the results are based on silt lines, scour lines, or debris that indicate maximum flood stage. Thus, the quoted discharges are directly associated with the largest flood that has occurred at the site over the period of the palaeoflood record. While we stress that the palaeoflood discharge studies based solely on the height of flood deposits are minimum estimates of the peak discharge, considerable experience demonstrate that discharge underestimation is probably 20% or less (e.g., Kochel *et al.* 1982). Baker (1987*a*) reviews the field observations that justify this conclusion.

The palaeoflood methodology used for data reported herein corresponds strictly to the 'slackwater deposit and palaeostage indicator' (SWD-PSI) technique (Baker 1987*a*) applied either to stable-boundary reaches or to reaches with well-known geometry. Data obtained through other palaeoflood reconstruction techniques, including regime-based procedures and palaeocompetence studies, are not included in Table 1. In that table the discharges are listed according to the ranges reported by the original researchers and they are the largest palaeoflood discharges in each of the studied basins in Utah and Arizona (Fig. 1). All of these palaeoflood sites were studied by past and present researchers at the Arizona Laboratory for Paleohydrological and Hydroclimatological Analysis in the Department of Geosciences at the University of Arizona. Therefore, we are sure that the field and laboratory procedures used to extract the data were very similar in all cases. As no evidence was mentioned by the original authors as to causes of floods other than precipitation, we assume that all the listed palaeofloods were formed through rainfall-runoff processes. Thirteen of these palaeofloods are the largest in at least the last 500 years, and eight are the largest during the last 1000–4000 years (Table 1).

Upper boundary to flood magnitudes

The hypothesis

The perspective that there should be a limit to flood magnitude and that this upper bound is related to the area of the specific drainage basin can be traced to classical studies in hydrology, including the seminal work by Horton. Horton (1936, pp. 437–438) was probably the first to hypothesize about the existence of an upper limit to flood magnitudes related to basin size:

> Flood magnitudes always continue to increase as the recurrence interval increases, but they increase toward a definite limit and not toward infinity. This is believed to be the more rational from of expression. No terrestrial stream can produce an infinite flood. A small stream cannot produce a major Mississippi River flood, for much the same reason that an ordinary barnyard fowl cannot lay an egg a yard in diameter: it would transcend nature's capabilities under the circumstances.

Since then, many studies have shown that floods are related to drainage area and that this variable is important in predicting flood magnitudes (e.g., Dooge 1986). In the southwestern United States, Benson (1964) concluded that drainage area is by far the most important basin characteristic in estimating flood magnitudes. For Arizona, his conclusion was supported by Roeske (1978, p. 33), who showed that drainage area is the only statistically significant variable. However, most researchers have used the drainage basin area as a parameter only in estimating magnitude of rare flood, not the magnitude of the maximum expected flood. On the other hand, envelope curves encompassing all the maximum flood peaks (discharge plotted vs. area) experienced either regionally (e.g., Creager 1939; Crippen & Bue 1977; Georgiadi 1979; Crippen 1982) or globally (Costa 1987) were used to advocate the existence of a maximum flood per drainage area. The existence of an upper limit to flood magnitude is not an assumption restricted to the advocates of the utility of envelope curves, the concept also underlies the more common, deterministic approach taken in using rainfall-runoff models of the 'worst case scenarios' for design purposes.

An interesting pattern emerges from the construction of regional envelope curves of maximum discharges. Several researchers have noted that incremental increases in the temporal and spatial base of the observational record impart progressively smaller changes in the form and position of envelope curves of peak discharge vs. drainage area (e.g., Creager 1939; Matthai 1969; Crippen 1982; Wolman & Costa 1984; Costa 1987). While this phenomenon can be explained by probabilistic reasoning (Yevjevich & Harmancioglu 1987), it has also been hypothesized that the apparent recent stabilization in the envelope curves encompassing the maximum floods in the Untied States is indicative of the existence of an upper limit to flood magnitudes (Wolman & Costa 1984; Costa 1987) and is not simply a stochastic phenomenon. Wolman & Costa (1984) also hypothesized that hydroclimatological processes and basin characteristics sustain the main control on the upper limit of flooding. The hypothesis of an upper limit to flood magnitudes has been difficult to substantiate because of obvious limitations on the rate of accumulation of observational data. In this study we present a means of overcoming the above limitations. Rather than waiting a very long time for sufficient future data, we accumulated data on floods which have already occurred in the region. Thus results from 25 palaeoflood hydrological studies in the lower Colorado River basin covering the last several centuries to millennia are added to the database previously composed of only modern and historical data covering 100 years at best. This augmentation of the flood record extends the effective length of observation at individual sites by hundreds to thousands of years and thereby allows for an independent evaluation of the hypothesis that an enveloping curve with a sufficiently broad spatial and temporal data base stabilizes at a position approximating a natural upper bound to flood magnitudes in a given region. There are two possible results from this exercise: either the additional palaeoflood data will alter (i.e., raise) the regional envelope curve based on modern and historical data, or the maximum palaeoflood discharges will fall within the existing curve. If the curve remains unchanged by the addition of palaeoflood data, the results then are consistent with the hypothesis of an upper limit to flood magnitudes.

Data and results from the southwestern United States

The US Bureau of Reclamation (1990) constructed an envelope curve for the largest gauged and historical floods in the Colorado River basin (Fig. 2 curve C) and we use this as the modern envelope curve. The 26 curve-controlling floods are marked in Fig. 2. This curve was originally constructed only for drainage basins with areas greater than about 258 km^2 (100 sq miles), and we extended it to the smaller drainage basins for the purpose of this study (Fig. 2). This extension is somewhat problematic, because it leaves out three estimations of flood peaks above the curve; Bronco Creek near Wikieup, Arizona; Eldorado Canyon at Nelson Landing, Nevada; and Dragoon Wash at St. David, Arizona (Fig. 2 numbers 4 5, and 6 respectively). To include these three floods at or below the envelope curve will demand a major discontinuity in the shape of the curve. In 1987, Costa suggested that any flood estimate that falls above the envelope curve should be carefully reexamined. These three specific estimations were challenged earlier by several authors (see Enzel *et al.* 1993, who also summarized critiques by Carmody, 1980 and Malvick, 1980). More recently, House & Pearthree (1995) demonstrated that the published Bronco Creek flood magnitude is a large overestimation. Therefore, we accepted the extension of the US Bureau of Reclamation's curve as is.

Curve A in Fig. 2 encompasses all of the largest flood magnitudes in the United States as reported by Costa (1987). Curve B is defined by the palaeoflood data from the lower Colorado River basin (Table 1). A comparison between the United States curve and the gauged and historical data from the Colorado River basin indicates

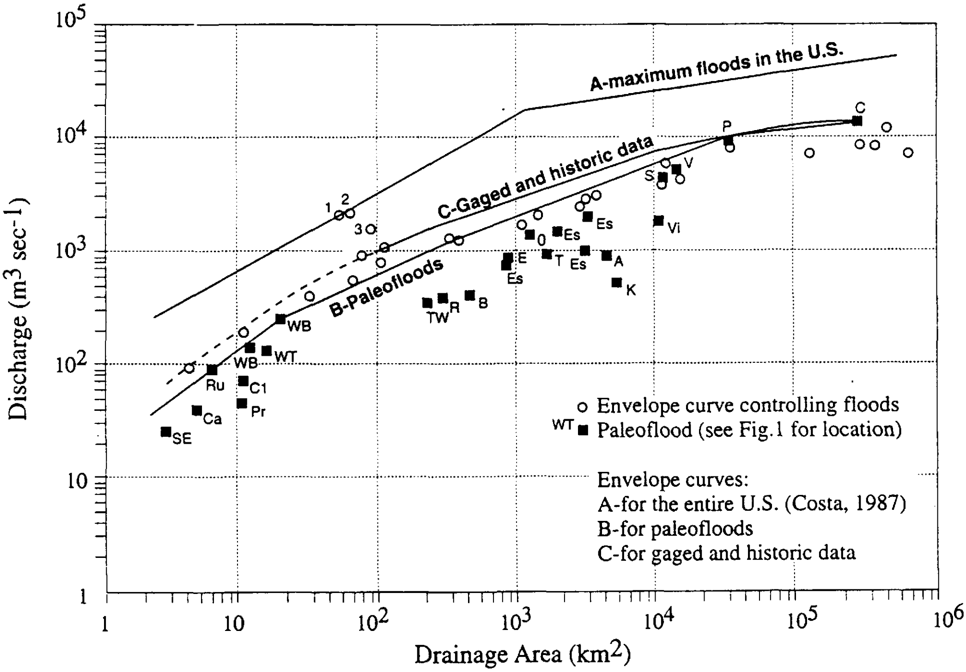

Fig. 2. Envelope curves of the largest floods in the United States (curve A; from Costa 1987), the documented palaeofloods from Table 1 (curve B), and the largest gauged and historical floods (curve C; solid part from US Bureau of Reclamation (1990) and dashed part is our extension of the curve to drainage basins with smaller areas). The numbers 1 2, and 3 next to the three envelope-curve-controlling floods indicate the three floods discussed in text.

that drainage basins in the Colorado River basin produce systematically smaller floods than drainage basins in other regions of the United States. Although we are concerned about the accuracy of the modern discharge estimates, we directed our efforts at identifying the general position and trend of the upper bound and its relation to the palaeoflood data. Therefore, curve C (Fig. 2) encompasses all of the floods except those which are obviously controversial discharge estimates. This curve can be used as a tool to identify those floods which warrant a reexamination, similar to the suggestion by Costa (1987), and the practice by Carmody (1980) and Malvick (1980).

It is demonstrated in Fig. 2 that in the Colorado River basin, a substantial increase in the temporal scale of flood records does not change the position of the envelope curve based on the gauged and historical data. The palaeoflood discharges fall on or below the curve which envelopes the largest gauged and historical floods (Fig. 2). Although these palaeoflood discharges represent much longer periods of record and are usually larger than modern floods in the individual rivers where they were studied, they nevertheless are remarkably similar to the magnitudes of the largest modern or historical floods in the region. The relation between palaeoflood discharges and the regional envelope curve for the Colorado River basin is consistent with the concept that there is a physical or hydrometeorological limit on the magnitude of the maximum flood that can be expected in a given drainage (Costa 1987).

Because no other palaeoflood data base similar to the one for the lower Colorado River basin exists, the only way to further test the emerging pattern is on a subset of the data on a somewhat smaller and more hydroclimatically homogenous area. We chose the southern Arizona subregion because of data availability. The maximum floods that occurred in drainage basins within arid and semiarid southern Arizona are shown in Fig. 3. To produce a hydroclimatologically homogeneous data set we excluded from this figure gauging stations on the main stem of the Gila River, basins that drain to the Gila River from the north and have headwaters in high elevations, and one station that is clearly affected by urbanization in Tucson, Arizona. The resulting curve for the southern Arizona subregion is slightly different from curve C for the entire Colorado River basin. Smaller basins (0.2–1 km^2) were included in the curve for southern Arizona because the palaeoflood data for the smaller sized basins are exclusively from that region. Palaeoflood magnitudes were estimated for five small drainage basins in the Tortolita Mountains north of Tucson, Arizona, for one basin the White Tank Mountains, for Tiger Wash near the Harquahala Mountains, and for one basin in Sierra Estrella west of Phoenix, Arizona (Fig. 1, Table 1; House 1991; House *et al.* 1991; P. K. House, unpublished data, CH2MHill & French 1992). Although only one radiocarbon date is available for these palaeofloods, field evidence and relative age dating indicate that they are the largest floods to have occurred in these ungauged basins during at least the last several hundred years (Baker *et al.* 1990; House 1991).

All of these palaeoflood discharges plot on or below the envelope curve constructed from the gauged data from southern Arizona (Fig. 3). Martinez-Goytre *et al.* (1994) recently provided an additional eight maximum palaeoflood discharge estimations from southern Arizona. None is higher than the regional envelope curve shown in Fig. 3. In addition, none of the large floods which occurred in Arizona in January and February 1993 (several were the largest on record),

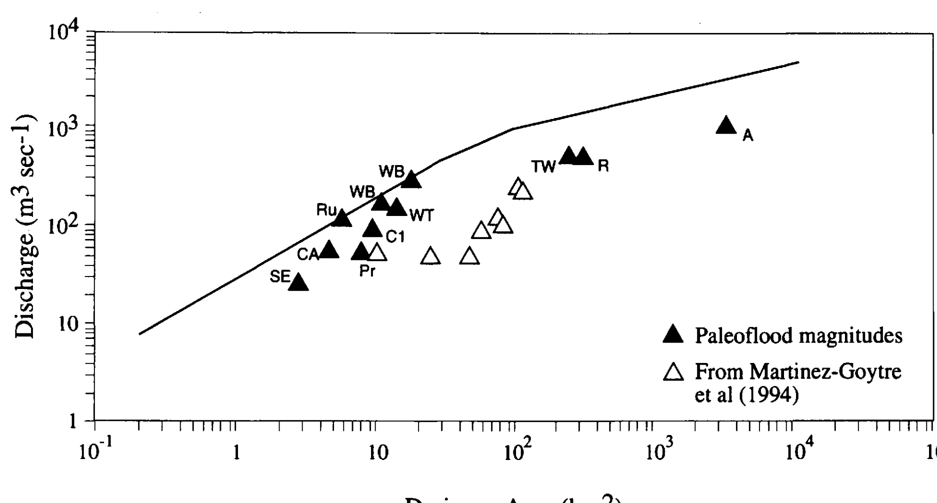

Fig. 3. Envelope curve for southern Arizona floods and palaeofloods from that region (letters are in reference to Table 1). Open triangles are the largest paleofloods in basins draining the Santa Catalina mountains (SC in Table 1) in Arizona estimated by Martinez-Goytre *et al.* (1994).

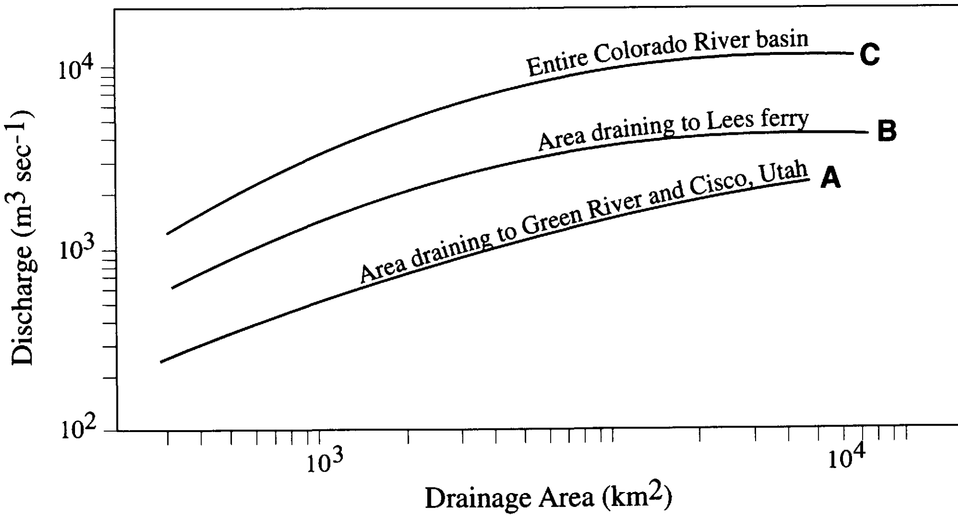

Fig. 4. Envelope curves for: (A) the upper Colorado River drainage basin (all gauging stations in the basin from the headwaters to Green River, Utah and Cisco, Utah on the Green and Colorado Rivers, respectively; see inset in Fig. 1); (B) the upper and middle Colorado River basins (all gauging stations from the headwaters to Lees Ferry, Arizona; Fig. 1); (C) the entire Colorado River basin (similar to curve C in Fig. 2).

exceeded the regional envelope curve (House 1993). The relationship between the palaeoflood discharges and the regional envelope curves is consistent with the assumption of an upper limit to flood magnitudes in both the entire Colorado River basin and the southern Arizona subregion. Analysis of other subregions within the Colorado River basin indicates that they have substantially different envelope curves for maximum instantaneous flood peaks (Fig. 4). Different types of storms produce the envelope-shaping peak floods in each subregion, indicating that hydroclimatology plays a major role in defining the curves. Although it is assumed by the authors that hydroclimatology is the cause for the upper limit of floods (Wolman & Costa 1984), additional inquiries into these phenomena are needed.

The existence of a natural upper limit would raise questions about the basic assumption, intrinsic to frequently used models in the probabilistic approach of flood-frequency analyses, that the upper bound on flood magnitudes cannot be determined and that the largest floods are beyond the range of practical concerns. The patterns presented in Figs 2 and 3 indicate that the upper bound is not beyond that range.

Clustering of the largest floods

To resolve the question of random versus clustering influences on the timing of maximum flood peaks we used evidence of 251 palaeofloods derived from palaeoflood chronologies on 19 rivers in the lower Colorado River basin (Fig. 1). The flood chronologies from these sites are listed, summarized, and discussed in detail by Ely (1992) and studies by Ely & Baker (1985), O'Connor *et al.* (1986*a*, *b*, 1994), Partridge & Baker (1987), Roberts (1987), Webb *et al.* (1988), Wohl (1989), Melis (1990), Enzel *et al.* (1994) and Ely (in press). These chronologies are based on cultural artifacts and on more than 150 radiocarbon dates (listed in Ely 1992). The dated floods were placed into time intervals of 200 years each throughout the last 5600 radiocarbon years. These intervals are suitable to the resolution of the method and data. Unfortunately, there is no way to divide the record into smaller intervals. Combining the palaeoflood information into one data set allows a greater confidence in the emerging temporal patterns in the occurrence of large floods than would information based on single studies of individual rivers.

The combined record (Fig. 5) of palaeofloods from the lower Colorado River area indicates a distinct pattern of large floods clustering in specific time episodes during the last 5000 years (Ely *et al.* 1993). Periods of numerous large floods span 5–3.6 ka, 2.2–0.8 ka and 0.6–0 ka. The episodes of few to no large floods are equally important from a climatic standpoint. The longest and most pronounced period of very few large floods is 3.6–2.2 ka. Another significant hiatus in large floods occurred from 0.8–0.6 ka, which shows a significant drop in the number of large floods immediately after an episode (1–0.8 ka) of particularly frequent high-magnitude floods. Periods of numerous large floods correspond to high lake levels (Mehringer & Warren 1976; Smith 1979; Enzel *et al.* 1989, 1992; Waters 1989; Stine 1990; Enzel & Wells in press) and groundwater discharge (Benson & Klieforth 1989) in the southwestern United States, global neoglacial advances (Rothlisberger 1986; Wigley 1988), and frequent strong El Niño events (Anderson 1993) (Fig. 5). Enzel *et al.* (1989) and Enzel & Wells

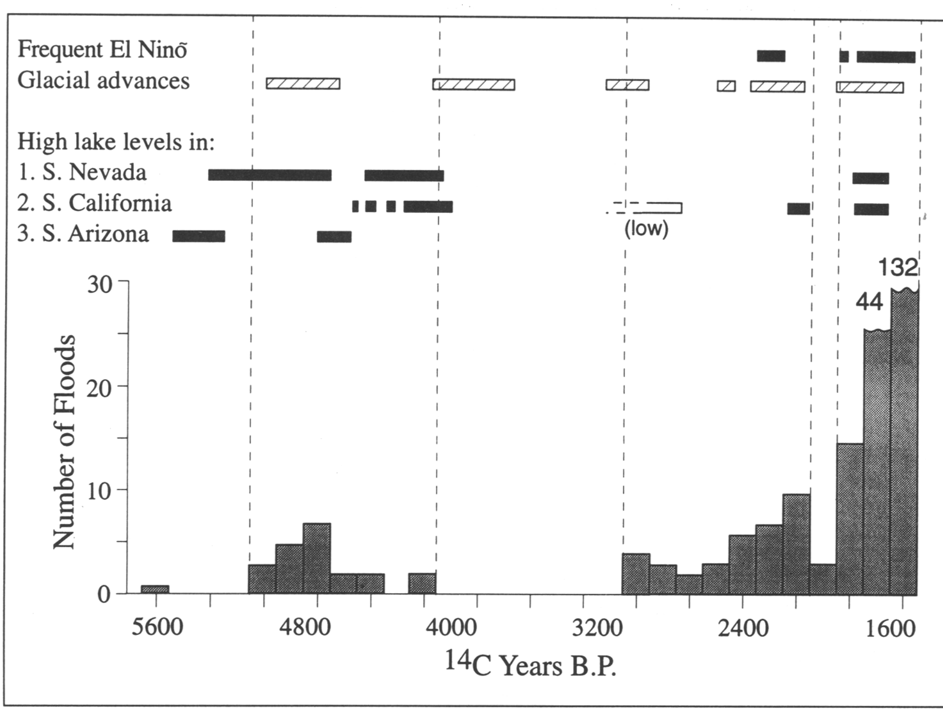

Fig. 5. Composite chronology of the palaeofloods in the lower Colorado River basin in Arizona and southern Utah. The numbers of palaeofloods are arranged according to 200-year intervals. All ages are uncalibrated ^{14}C dates. See text for references of the other palaeohydrological and palaeoclimatological records. Modified from Ely *et al.* (1993).

(in press) have demonstrated that the formation and maintaining of short duration Holocene lakes in the terminus of the Mojave River in southern California could have been caused only by a large increase in the frequency of floods with magnitudes similar to the largest observed in the modern record (Enzel 1992).

The increase in flood frequency during the last 400–600 years is apparent from Fig. 5. This can be either due to a bias in the methodology of palaeoflood hydrology or a true phenomenon. To try to detect a preservation bias we omitted chronologies with a record shorter than 400 years and normalized the number of floods by the number of rivers which produced data for each interval. The results (Ely 1992) did not alter the basic observation of increased flooding in the last several centuries. Therefore, the resulting pattern is most likely not merely an artifact of differential preservation. Other palaeohydrological evidence also indicate an increase in annual runoff in the Southwest (D'Arrigo & Jacoby 1991) and channel incision (e.g., Cooke & Reeves 1976; Webb *et al.* 1991) in the late nineteenth–early twentieth centuries, which is consistent with increased flooding in the region during that time.

We assume that the causes of the largest floods during the late Holocene are similar to those of the modern floods. If the modern, regional oceanic and atmospheric conditions necessary to produce the largest floods are anomalous and unique, then similar conditions must have occurred to produce the large flood peaks. Modern storms which produce the largest floods in the sites summarized here are mainly winter North Pacific frontal storms and late-summer and fall storms associated with Pacific tropical cyclones over northwestern Mexico in conjunction with mid-latitude low-pressure troughs (Ely *et al.* 1994). Local summer convective storms are significant flood-producing storms only in basins much smaller than those used to construct the palaeoflood chronology (Ely 1992).

We have determined that during both winter and tropical storms associated with large floods in the study area, the atmospheric circulation shifts the storm track southward toward the southwestern United States. The winter storm track is shifted far to the south of its normal position (Enzel *et al.* 1989; Ely *et al.* 1994). An unusually low pressure anomaly off the coast of California, and a high pressure anomaly over the Aleutians or the Gulf of Alaska are characteristics of the winter floods (Ely *et al.* 1994). Similar atmospheric patterns caused large winter floods in southern California (Enzel *et al.* 1989; Enzel & Wells in press). Tropical-cyclone floods exhibit a similar low-pressure anomaly and an extended blocking high-pressure anomaly in the central North Pacific in addition to the entrainment of an eastern Pacific tropical storm (Ely *et al.* 1993). On the longer, late Holocene time scale, periods of increased flooding are associated with distinct, persistent changes in regional climate, large-scale atmospheric circulation patterns (Enzel *et al.* 1992; Ely *et al.* 1993), and an increased frequency of strong El Niño events as reflected both in the Nile record (Anderson 1993; Quinn 1993) (Fig. 5) and in the El Niño-like warm sea surface temperatures off the coast of California (Pisias 1978). The variations in flood frequency over the last 1000 years agree exceptionally well with these records (Ely *et al.* 1994).

Combining the upper boundary approach with the suggested climatic forcing on the occurrence of the largest floods reveals the potential effect of climatic variations on the magnitudes for the rarest flood events in the region. The influence of climatic variability on the occurrences of extreme floods has been recognized at several time scales, and mechanisms to explain this association have been suggested (Knox 1983;

Webb 1985; Baker 1987*b*; Enzel *et al.* 1989; Ely and Baker 1990; Ely 1992; Ely *et al.* 1994; Enzel 1992; Webb & Betancourt 1992). Climate has also varied over different temporal scales during the late Holocene (e.g., Bradley 1985), the period for which palaeoflood data are available. However, not one Colorado River tributary where a palaeoflood study has been performed has produced a flood with a magnitude greater than the flood expected from the envelope curve of the modern record. In addition, a hydrological model of the Holocene lakes in the Mojave River terminus (Enzel *et al.* 1989) showed that although the floods that produced these lakes were similar to the modern largest floods, they could not have been much larger than the largest observed in the modern and historical records (Enzel 1992). The increased frequencies of large floods during distinct time periods over the last several thousand years were not necessarily associated with greatly increased peak flood magnitudes.

Summary

Combining palaeoflood data from as many rivers and streams as possible allows important patterns to emerge. These patterns are unseen from data sets limited in their spatial and temporal coverage. The palaeoflood data have provided evidence for the existence of an upper limit to peak flood magnitudes in a region. Our results suggest that this upper limit can be estimated from the combination of modern, historical, and palaeoflood data (Enzel *et al.* 1993). This concept has been discussed in previous studies, but lack of detailed, high-quality, long-term data prevented the discussions to extend beyond suggestions. We stress that not all the rivers in the region will experience floods equivalent to the maximum discharge magnitude delineated by the envelope curve. The largest flood within a specific basin in the region can be smaller than the curve, but will not exceed the curve.

The palaeoflood data also provided a regional chronology of large floods which exhibited a distinct clustering of floods through time. Comparison with other palaeoclimatic and palaeohydrologic data indicates a strong climatic control on the temporal distribution of floods in the southwestern United States (Ely *et al.* 1993).

These patterns hold major implications for the study of flood hydrology, in particular to the increased understanding of the relationship between climate variability and floods, and application of flood hydrology toward a better understanding of the maximum expected flood on a given river. This research demonstrates the applicability of palaeohydrological information parameters in testing concepts and assumptions that have long affected scientific thought in surface-water hydrology.

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The response of geomorphic systems to climatic and hydrological change during the Late Glacial and early Holocene in the humid and sub-humid tropics

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Abstract: Dated alluvial stratigraphies indicative of late Quaternary environmental change in the humid tropics have increased, but the database remains inadequate and the intensity and duration of wet–dry oscillations and responses of hillslopes and river systems remain poorly understood. Dry conditions at the Last Glacial Maximum were marked by semi-arid landforms with reduced stream activity. Large palaeofloods, valley-floor erosion, channel cutting and flood deposition occurred at the Pleistocene–Holocene transition after 13 000 BP. Distinctive floodplains and stratigraphies characterized by multiple shifts from lateral to vertical accretion were built initially over a period of nearly 2 ka after 9 500 BP during the early Holocene pluvial in Africa, Kalimantan and Amazonia during and after re-establishment of the lowland rainforests. Several wet–dry climatic oscillations followed in the mid-Holocene period and are marked by alluvial cut and fill sequences and by slightly thinner but coarse textured floodplain overbank sediments.

Late Pleistocene and Holocene signals of climate change are becoming more widely available for the tropics, mainly from studies of pollen and lake sediments. However, there is a gap in our knowledge between the inferences about vegetation and climate and our understanding of river behaviour and landscape dynamics as responses to the changes. That understanding needs to come, in part at least, from the analysis, dating and interpretation of alluvial deposits. Unfortunately, there is a dearth of published alluvial evidence from within the humid tropics (Thomas 1994).

There are several problems associated with environmental and palaeohydrological interpretations of alluvial sediments. It is not always easy to distinguish the impacts of regional changes in controlling factors such as catchment relief, tectonics, and base level, from the effects of site sensitivity on the response to single high magnitude events or to secondary threshold effects (complex response) within the geomorphic system (Schumm 1977). Progressive stripping of relict late Tertiary weathering mantles during episodes of rapid landscape change must also have a cumulative impact and also limit the potential for repeated events at the same site.

The impact of human occupation on both pollen spectra and on lake sedimentation (Flenley 1988) must also clearly lead to other uncertainties of interpretation, and research into the impacts of possible anthropogenic vegetation change, as early as 26 000 BP in Irian Jaya (Haberle *et al.* 1991) and occurring more widely by 7–5 ka BP, on landscape dynamics and river behaviour is at a primitive stage.

A second general problem facing palaeohydrological interpretation concerns the nature of past climate, particularly rainfall regimes, where close analogies with current climates may be inappropriate. The notion of geomorphologically effective rain events based on the concept of changing magnitude and frequency over time and space is arguably the most appropriate way forward in this field (Wolman & Miller 1960; Ahnert 1987; Ehbergen & Imeson 1989). Unfortunately, existing palaeoenvironmental data and models of palaeoclimate do not usually afford such a detailed understanding. It is now accepted that climate changes can be abrupt, occurring over 10^2 years rather than 10^3 years, and that particular climatic patterns may persist for a few hundred years only (Street-Perrott & Perrott 1990; Gasse & Van Campo 1994). Consequently low resolution studies of sediments and peats can be misleading. On the other hand, high-resolution studies will converge with concepts of flood frequency and inter-annual climate variability and event stratigraphy.

Equally worrying are the doubts expressed concerning the supposed relationships between climate, vegetation and sediment yield on the one hand, and predictions about stream power, erosion and deposition on the other. Much of our thinking in this regard for regions outside the recently glaciated areas derives from work done in the temperate continental and the arid zone climates (Langbein & Schumm 1958; Schumm 1965, 1968; Knox 1972, 1984; Bull 1991). Thus it is common to indicate a peak of sediment production during periods of rapidly changing climate, as when rainfall increases ahead of vegetation recovery (Knox 1972, 1983; Thomas & Thorp 1980, 1992; Roberts & Baker 1993). In a comparable manner rivers are expected to aggrade their valleys when sediment yield is high but discharge is erratic, and to erode their beds when forested conditions promote broader flood peaks and lower sediment concentrations. However, not all of this reasoning is matched to field evidence and some simple models of these kinds may be very misleading. Several issues can be listed:

(1) The model of long periods of 'stable' climate punctuated by intervals of rapid change is not adequate to account for landscape complexity, because it largely ignores the magnitude–frequency properties of climate; additionally, 'stable' climates may not have persisted throughout the periods indicated on most models.

(2) The rainfall regime is critical to any arguments concerning stream activity. Runoff trends vary with storm size and antecedent moisture and can be almost continuous under equatorial conditions. In monsoonal climates, on the other hand, there is a period of major annual floods, creating a distinctive fluvial regime (Gupta & Dutt 1989), different from that of the drier seasonal savanna climates found in parts of Africa and S America. In his analysis of *rainfall erosivity* in tropical climates in West Africa Roose (1977, 1981) found immense variations across the region in the erosivity of rainfall such that slight spatial shifts in vegetation communities can induce major changes in sediment yield. His model is unlikely to apply in all tropical climates, but where the rainfall generating mechanism is similar throughout an area it may have great relevance to debates about landscape response to rainfall and vegetation changes.

(3) The production of sediment from hillslopes may arise from mass movements favoured by saturated soil conditions, as well as by surface wash which is promoted by open vegetation. But in neither case are connections to stream channels always direct, and sediment stores in the landscape play a major role in controlling delivery into rivers, a point made forcibly by Church & Slaymaker (1989) with respect to glaciated terrain. Thus, while highland streams may respond to the magnitude and frequency of

landslides coming directly into the channel, plains rivers create floodplains over 10^4 years, and these form the major sources of sediment entering the channel.

(4) Late Quaternary events were superimposed on landscapes produced during many previous cycles or oscillations of climate. Those landscapes contain glacis, fans, terraces and a multitude of hillslope forms, some parts of which act as barriers to the entry of sediment into present-day rivers and strongly influence morphostratigraphy of Holocene floodplains.

It is clear, therefore, that appraisal of this group of problems requires a close scrutiny of the processes involved and of their periodicities and controlling factors (Eybergen & Imeson 1989). However, published knowledge of current catchment geomorphological behaviour in response to meteorological events is limited (see Thomas 1994 and Douglas & Spencer 1985 for collations of data). For these and other reasons, the interpretation of many alluvial forms and deposits must depend on site properties, their sensitivity to change and their regional settings. The analysis of material from vertical cores taken from enclosed depressions does not always satisfy this requirement, while the interpretation of lake levels is notoriously difficult, where basin morphology, catchment size, altitudinal variation and hydrogeology can all be extremely variable (Gasse & Van Campo 1994). However, the use of a large number of observations has allowed world-wide correlations to be offered for Late Quaternary lake levels (Street & Grove 1979; Street-Perrott *et al.* 1985), and this comparative method is important for other studies, including those of hillslope morphology and stream sedimentation.

In this review the alluvial responses to environmental change in the humid tropics will be confined to the period spanning the recovery from the LGM through to

Table 1. *Chronology of late Quaternary environmental change in the humid tropics*

Radiocarbon Years BP	Probable environmental conditions (recorded examples in brackets)
3 100–2 400	Possibly drier, accompanied by deforestation and human occupation, and continuing to the present day. West Africa (Ghana, post-2400 years BP Brazil)
3 400–3 100	Increased humidity in forested tropics with rising discharges, several lesser oscillations of humidity followed
4 200–3 400	Mid Holocene dry phase, probably quite severe (Africa, Brazil)
5 500–4 200	Declining humidity in some areas of humid tropics (dry excursions in Amazonia, 5500; 4800)
7 000–5 500	Increased humidity and modest rise in lake levels
7 800–7 000	Reduced lake levels and river discharges in W & E Africa, Brazil
10 500–8 000	Second humid period with high lake levels and discharges; re-establishment of forest
11 000–10 500	Dry, cool interval in many areas; low lake levels (Younger Dryas)
12 500–11 000	Rapid warming with unstable climates and prolonged heavy rains in tropical Africa, very high lake levels (world-wide)
13 000/12 000–post-22 000	Becoming cold in uplands and dry in most lowlands; by 18 000 tree-line depressed 1000 m, rainfalls possibly reduced by 50% (most sites)

Main Sources: Absy *et al.* (1989, 1991); Adamson *et al.* (1980); Butzer 1980; Kershaw (1978, 1992); Schubert (1988); Street & Grove (1979); Talbot *et al.* (1984); Thomas & Thorp (1980, 1985); Thorp *et al.* (1990); Street-Perrott & Perrott (1990); Gasse & Van Campo (1994).

c. 3 ka BP, when the environmental signal in alluvial stratigraphies becomes strongly mixed with that from human activities. The chronological framework for this period is tentatively generalised and summarized in Table 1.

Naturally, such a summary table hides many of the disagreements and differences between regions in the number, onset and duration of major climate changes. Distinguishing the effects of human environmental modification from those of climatic change is a further problem. Preliminary indications suggest that early deforestation in the lowland humid tropics *c.* 10–4 ka BP was probably small scale, short-term and spatially very patchy and its impacts may have been short-lived and localized, until widespread agriculture associated with larger populations spread across many tropical areas in a diachronous fashion, perhaps from about 5 ka BP onwards, but mainly during the last 2–3 ka (Flenley 1988; Hope & Tulip 1994; Jolley *et al.* 1994).

Conditions during the Late Glacial Maximum

The duration of ice-age aridity varied from region to region appearing post 22 ka BP in most areas and, according to some writers, ameliorating by 16.5 ka BP in SE India, (Van Campo 1986), by 15–14 ka BP in the west Zaire basin (Kadomura 1995) whilst elsewhere, dry climates seem to have persisted until well after 14 ka BP.

Evidence of geomorphological conditions during the Late Glacial Maximum throughout the present humid and seasonally humid tropics includes:

- deep sand-filled desiccation cracks in the weathered phyllite beneath shallow tributary valleys floors now buried beneath early Holocene colluvial and alluvial gravels in southern Ghana (Junner 1943; Hall *et al.* 1985);
- an apparent absence of alluvial sedimentary units in both headwaters and trunk streams between 25 ka BP and 13 ka BP in Amazonia (Van der Hammen *et al.* 1992*a* & *b*), West Africa (Thomas & Thorp 1980, 1985; Hall *et al.* 1985) and Kalimantan (Thorp *et al.* 1990, Thorp & Thomas, 1992);
- dry lake beds and swamps;
- regolith-stripped slopes found widely under present day forest and other woodland areas;
- stonelines within the forest and savanna zones;
- palaeopans, lunettes and other dune sands in Venezuela (Tricart 1982, 1985);
- large fans of course material as in the Pantanal (Mato Grosso, Brazil) (Klammer 1982), the Orinoco Llanos (Clapperton 1993*a*);
- widespread colluvium (Thomas 1994).

Collectively these indicate relative aridity and a suppression of the characteristic fluvial processes which justify suggestions of rainfall reductions of 30–66% compared to present values for large areas of the lowland tropics (Rossignol-Strick & Duzer 1979; Verstappen 1980; Peters & Tetzlaff 1990; Crowley & North 1991; Heaney 1991: Thomas 1994; Van der Hammen & Asby 1994). These conclusions pose questions regarding the vegetation cover at the LGM which are impossible to answer on the scale of the global tropics. In mountain areas, reduced temperatures depressed treelines by *c.* 1 km, but the impact of these changes on the lowlands remains uncertain. The appearance of montaine flora in lowland pollen spectra from west Africa (Maley 1991; Giresse *et al.* 1994) and Panama (Bush *et al.* 1992) has been used as an argument against the refugia theory of Haffer (1969, 1987), but in terms of landscape ecology and dynamics this may be misleading. Forested conditions clearly persisted throughout the last glacial cycle in certain favoured areas that may have included western equatorial Africa (Preuss 1990; Maley 1991) and western Amazonia (Van der Hammen & Absy, 1994). But conditions were probably dry enough after *c.* 20 ka BP for 5–7 Ma for many areas of former forest to become mosaics of open woodland and grassland (Giresse *et al.* 1994; Van der Hammen & Absy 1994). Arguments about biodiversity in relation to so-called refugia are clearly contentious, but the persistence of lowland TRF during the last glacial cooling (−4° to 6°C) must have occurred in climatically and edaphicaly favoured (refuge or heartland) areas, albeit in modified form. In any case major reductions in rainfalls (30 to 60%) led to the reduction of water surpluses and dwindling of runoff and streamflow (Thomas & Thorp 1980; Van der Hammen & Absy 1994) beneath the prevailing vegetation cover. Towards the margins of the humid tropics, it seems likely that strengthened Trade winds, less frequent storm activity and reduced rainfall could have combined to produce landscapes of very open woodland.

The late Pleistocene–Holocene transition

The onset of the late Pleistocene–Holocene transitional period is of great interest geomorphologically in the humid tropics and there is evidence for a period of unstable and often very wet conditions beginning around 12.5 ka BP and lasting *c.* 1.5 ka. It is remarkable how frequently this date is referenced around the globe.

In Africa, wetter conditions appear to have become established near the Equator by 13 ka BP but did not become effective in the southern margins of the Sahara until 10 or even 9 ka BP (Fabre & Petit-Maire 1988). This illustrates the time-transgressive nature of climate changes when viewed across regions such as west Africa (Fairbridge 1976). It may also mean that the number of 'effective' fluctuations in rainfall will have varied significantly between areas, being fewer in the more humid tropics, where the forests probably survived drier episodes except in unfavourable edaphic sites.

Large palaeofloods on the Nile and the Niger and also in smaller catchments in the seasonally humid tropics of west Africa commenced after 12.7 ka BP. The overflow of Lake Victoria and the deposition of the Sheik Hassan silts 30 m above the floodplain of the Nile at Wadi Haifa (the Wild Nile) have been given dates of 12.5–11.5 ka BP (Paulissen 1989). The first major depositional events of this period in the middle Birim River of Ghana and in the Bafi-Sewa headwaters in Sierra Leone date to the same period (12.7–12.4 ka BP) and are associated with extensive erosion of valley floors (Thomas & Thorp 1980; Hall *et al.* 1985; Thorp & Thomas 1992).

On the other hand, the major pulse of off-shore sedimentation, at least from major African rivers, appears to have taken place after 12 ka BP, with the sedimentation peaks in the Niger delta dated to *c.* 11.5 ka BP (Pastouret *et al.* 1978) and in the Congo estuary to 11.23 ka BP (Giresse & Lanfranchi 1984), whilst the lower sapropel muds of the eastern Mediterranean are bracketed between 11.76 ka BP and 10.44 ka BP (Rossignol-Strick *et al.* 1982). The sedimentation rates increased by four times in the Congo estuary and by 18 times in the Niger delta at these times. It is tempting to suggest that the lag between fluvial erosion in the interior and the peak in offshore delta sedimentation reflects the sediment transfer time between the two, involving sediment storage on lower slopes and in floodplains.

After 13 ka BP, hillslopes were probably subjected to more frequent and powerful sheetflood events and tributary channel extension led to the evacuation of colluvium from hillslope hollows. A fluvial landscape reasserted itself, by reactivating drainage lines. Prolonged rain, falling on regolith with high-rainfall acceptance, could also have led to landsliding, and to continued widespread fan formation especially in modern savanna areas. These fans were later abandoned or trenched as the sediment supply from upstream sources diminished and regular stream flows, albeit seasonal, became established.

During the very large flood events which appear to have occurred during this phase, the critical power of streams appears to have been enough to remove coarse, often boulder sized, channel sediments, leading to erosion of 'bedrock' channels in Sierra Leone and Ghana. In eastern Zambia modern 'dambos' draining hillfoot zones have indistinct or no stream channels today, but exhibit gravel terraces 2–3 m above the valley floors, and also contain 3 m of fine gravel and sandy sediment infilling buried channels.

Towards the end of this transition period, and following the Younger Dryas, with increasing vegetation cover and soil development, it may be surmised that flows became more seasonally regular and that flood peaks on many lowland tropical rivers probably became broader and less high. This would have allowed rivers to build alluvial plains and in due course to become single thread meandering streams. There are, however, few dated records of floodplain overbank alluvia from this period.

The early Holocene Pluvial

Following the dry-cool Younger Dryas episode, the subsequent development of pluvial forested conditions, marked by high lake levels and by fluvial sedimentation, is clearly attested in the environmental records for which there are abundant early Holocene (mostly post-10 ka BP) dates. What is not clear, however, is how long these conditions persisted, neither is it agreed what rainfall mechanisms were predominant and therefore what the magnitude and frequency of rainfall events may have been. However, Lézine & Vergnard-Grazzini (1993) have suggested a rainfall 300 mm wetter than today for the west African tropics at this time, a figure also used by Kershaw & Nix (1989) for northeast Queensland (although they recognized that the figure could be as high as 800 mm).

This interglacial pluvial is sometimes represented as persisting well into the middle Holocene, but after *c.* 8 ka BP there is evidence of drier conditions lasting perhaps 500 years in Ethiopia, western Sahara and Ghana (Gasse & Van Campo 1994; Talbot & Johannessen 1992), with a return to greater humidity of climate after *c.* 7 ka BP. A severe dry phase followed in Africa after 5 ka BP and Van der Hammen *et al.* (1992*a*, *b*) indicates three dry phases between 5 and 3 ka BP in Colombian Amazonia. Thus whilst there is considerable spatial synchroneity around the 11–10 ka BP and the 4–3 ka BP cooler and drier periods there is considerable spatial variation in their onset and termination and as more detail becomes available, the division of the Holocene climates into longer periods of climatic stability separated by shorter periods of rapid change may become less tenable.

Records from W Africa, Australia and Brazil suggest the re-establishment of lowland forest between 9 and 8 ka BP (Talbot & Delibrias 1980; Markgraf 1989; Servant *et al.* 1989; Kadomura & Maley 1994), in contrast to the absence of records for forest cover during the Pleistocene–Holocene transition. It is tempting, therefore, to suggest that in many areas the forest taxa were no longer present locally, and that migration of the rainforest from refuge areas such as western Amazonia and western equatorial Africa took up much of the period and was arrested for 500–1000 years during the Younger Dryas. The environmental controls over this recovery may have included the unstable climate of the late Pleistocene and earliest Holocene, and also the absence or continued loss of soil cover in many areas. It has been hypothesised that the period in itself was one of rapid hillslope erosion and landscape change (Thomas & Thorp 1980, 1992; Hall *et al.* 1985; Roberts & Baker 1993), and this must have opposed the re-establishment of equilibria (biostasie) in landscapes that had experienced 6–7 ka of dry conditions.

The development of maximum pluvial conditions after the Younger Dryas event is recognised not only in the second Mediterranean sapropel, which accumulated in the period from *c.* 9–8 ka BP, but is clearly recognized in the several published dated alluvial stratigraphies. In West Africa, in the Birim River of Ghana, early Holocene floodplain sedimentation occurred after 9 ka BP and continued for more than 1000 years during which time several 2–3 m gravel units were formed together with a final 5–7 m thick fine member overbank unit. This records a major aggradation of the valley floor, burying 5 m thick coarse gravels which had already accumulated during the Late Pleistocene transition period between 13 and 12 ka BP within a deep bedrock channel incised into the pre LGM, late Pleistocene bedrock surface (Hall *et al.* 1985). The overbank fine members of the early Holocene floodplain are dominated by clays, silts and fine sands in contrast to more sandy textured later Holocene floodplains. In Sierra Leone, the Bafi-Sewa headwaters exhibit a similar time concentration of channel sediments between 9.5 and 7.8 ka BP, and their overbank fine members also contain organic rich clay facies within generally finer textured sediments than those of the later Holocene inset floodplains.

In equatorial western Kalimantan, floodplain sedimentation also commenced around this time with dates from basal gravels beneath the floodplains of the Mandor and Raya rivers returning dates of 9.97 ka BP and 10.25 ka BP respectively (Thorp *et al.* 1990; Thorp & Thomas 1992). In their recent study of Holocene palaeochannels of the Yom river in the central plain of Thailand, Bishop & Godley (1994) identify several Holocene channel and floodplain formations. The early Holocene floodplain appears to have been constructed by higher discharges than the later units. Their calculations indicate a reduction factor of ×3–4 for bankfull discharge (Qbf) between the early and the middle Holocene discharges. The greater humidity of the early Holocene is attested also by Löffler *et al.* (1984) and by Nutalaya *et al.* (1989) as quoted by Bishop & Godley (1994).

The Holocene sands and clays of the Blue (B) and White (W) Niles also show comparable flood peaks in the early/mid-Holocene, according to Williams (1980); Adamson *et al.* (1980): 8.4–8.1 ka BP (W); 7.5 ka BP (B); *c* 7 ka BP (B & W). But the picture is less clear in the late Holocene. High flows on the Congo/Zaire River dating to the period 10–8 ka BP are referred to by Preuss (1990).

In the Caquetá valley (Colombia), late Pleistocene to early Holocene floodplain formation commenced after *c.* 12.6 ka BP and persisted until around 11 ka BP, after which the sequence was partly eroded during the Younger Dryas/El Abra stadial (Van der Hammen *et al.* 1992*a*, *b*). After 10 ka BP they record renewed sedimentation and 'possible extensive inundation' which has again been evident in the last 3000 years due, they think, to Andean deforestation.

Results from some of these rivers are summarized in Table 2. During the period from *c.* 10.25 to 8 ka BP forested condition became widespread (Lézine &Vergnaud-Grazzini, 1993). Sediment supply was initially high with a wide range of textural sizes but may have declined to fine textured suspension loads with forest closure. Although deep-seated landslides are common in the humid tropics today, there is also abundant evidence of palaeolandslides in hilly terrains within savanna areas. Discharges appear to have been greater than today and single thread rivers built thick overbank depositional sequences. Stabilization of stream banks by fine silts and by vegetation would have inhibited lateral migration and irregularly sinuous, box-shaped channels, as in the Birim, became common on many rivers. The compromise between vertical accretion and overbank flooding meant that the latter became rarer, and occasional (10^2 year) high floods probably led to avulsion and splay formation on floodplains and to the evacuation of sediments from the small valleys, including the *dambos* and *bolis*.

During and after the mid-Holocene, possibly from as early as 7.8 ka BP and certainly after 5 ka BP drier conditions periodically recurred possibly lasting *c.* 500 years, but the forest remained largely intact although after 5 ka BP it became increasingly subject to agricultural burning and clearance. It is not clear what the impact of these drier periods was on the rivers of the time, except that they must have experienced diminished flows within channels designed for much larger discharges; possibly nothing much happened. However, bank instability and cavitation may have occurred, and during subsequent high floods, some degree of lateral stream migration, together with the generation of large amounts of fine sediment from bank erosion would have characterised the system. After *c.* 5 ka BP alluvial stratigraphies record several periods of enhanced floodplain erosion and rebuilding, supra-basal gravel reactivation and minor cut and fill assemblages.

Of interest are the erosional changes which terminated the major sedimentation periods. Those between the Transition and the early Holocene pluvial are perhaps

Table 2. *Dates of late Quaternary sedimentation in Blue (B) and White (W) Nile, Caquetá (Colombia), Koidu Basin (Sierra Leone) and Birim (Ghana) in C^{14} years* BP

Nile*	Caquetá†	Birim‡	Koidu§
		>2090–< 1780	
			c. 3000–*c.* 2300
			>4300–*c.* 3200
7 000 (B & W)			
7 500 (B)			
8 400–8100 (W)	10 000–8800–7500 > 9500–7800		
12 500–11 000 (B & W)	12 600–11 000 > 12 700– < 12 400–12 500– < 10 500		

* After Williams (1980); Adamson *et al.* (1980).
† After Van de Hammen *et al* (1992*a*, *b*).
‡ Ater Hall *et al.* (1985).
§ After Thomas & Thorp (1980).

more easy to explain than the erosion between the early Holocene floodplains and those built during the last 5000 years. *Ad hoc* hypotheses might be advanced invoking complex response or externally driven changes in water and sediment discharges after *c.* 7 ka BP, or the effects of high magnitude flow events during the several periods of reduced overall precipitation. However, as Bull (1991) has commented, rivers experiencing long term degradation due to tectonic factors, erode their beds whenever conditions for rapid sedimentation are absent. This wider view probably affects many of the river reaches quoted in this study.

Concluding discussion

At the Pleistocene–Holocene transition climates are thought to have become unstable, but the changes to rainfall amount and distribution are little understood. Widespread evidence for very high floods on rivers both large and small suggests a major increase in storm size and frequency in low latitudes impacting on landsurfaces partially adjusted during the preceding 5000–9000 dry years to more arid conditions, while the pan-tropical rise in lake levels indicates a major change in water balance, reflecting enhanced rainfall totals rather than reduced evaporation.

However, while this transition can be explained largely by insolation forcing, subsequent Holocene fluctations in tropical rainfall cannot, and variations in sea surface temperatures, or ocean currents, also fail to offer an explanation. Gasse & Van Campo (1994) have recently proposed that feedback mechanisms associated with the land surface conditions themselves may have been largely responsible. With initial post glacial warming and recovery of the monsoon rainfall mechanisms, came the expansion of forest vegetation, wetlands and lakes, decreasing albedo and increasing methane production into the atmosphere, reinforcing the insolation mechanism. The evaporation of the moisture in soils and lakes would subsequently require large amounts of solar energy and lead to a cooling of the surface, reversing the rainfall trends, and according to these authors, leading to a drying out of the land and a repetition of the cycle of rainfall fluctuation. They also point out that the declining magnitude of the fluctations follows the curve of N hemisphere insolation during the Holocene.

High runoff and sediment yields would have characterized many hillslopes at the time of the first major increase in rainfall and frequent debris flows combined with fan building and trenching may have taken place. Many 'head' deposits and former valley fills in hilly terrain may have been excavated, but there is evidence from the dambo valleys of the Nyika Plateau in Malawi (Meadows 1985) and from the Inyanga Highlands of Zimbabwe (Tomlinson 1974) to suggest a period of rapid sedimentation post 12 ka BP. It also seems likely that deeper seated landslides would have been generated as climates became increasingly humid. Rivers were subject to very high flood peaks which combined to create major fluxes of water and sediment in large catchments. There may be an important association here between rapid hillslope erosion, especially from earlier sediment stores, deposition of coarse sediments, the flushing of clays through the catchment system and the marked accumulation of kaolinitic clay in deltas and estuaries.

The Younger Dryas was probably important in the tropics as a period of drier climates lasting more than 500 years and this must have halted the spread of the forests outward from their refugia or heartland areas. But the lake level data do not suggest that the severity of climate was everywhere as great as during the LGM. Hillslope erosion would have continued and colluviation may have been widespread at this time, but fluvial activity was diminished.

The Holocene pluvial phase post 10 ka BP led to much higher lake levels than before and to renewed fluvial activity on a scale comparable with the late Pleistocene in some areas, but apparently did not lead to major sedimentation events in lower reaches and off-shore deltas. This has been interpreted as an indication of the stabilisation of the landscape as rainforest and other woodland vegetation was finally re-established between 9.5 and 8.5 ka BP, according to location and site conditions. But it must also reflect the build-up of sediment stores in early Holocene floodplains, where both lateral and vertical accretion deposits remain stored in large volumes.

This also seems to have been a period of widespread mass movement and colluviation. Palaeolandslides are common features throughout much of the humid tropics, but it is not known when these were first initiated and landslides are seldom dated. Nevertheless, an impression is gained of landscapes that have experienced sometime in the Holocene a phase of quite unparalleled morphogenetic activity which was triggered by a concentration of high magnitude rainfall events.

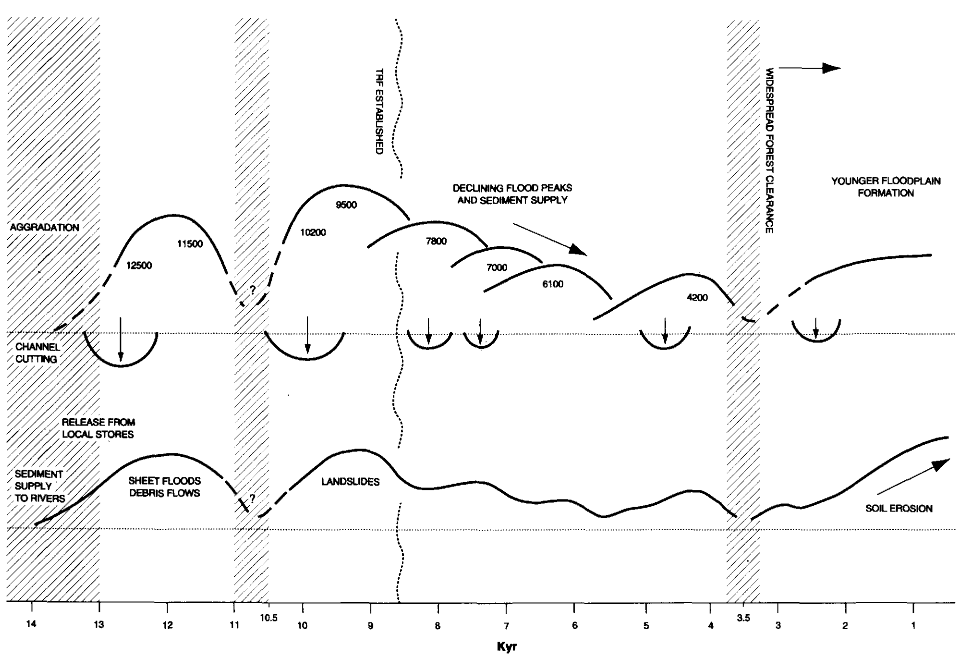

Fig. 1. Late Quaternary stream activity in the humid tropics of West Africa Curve peaks denote timing of major sedimentary units; troughs/arrows indicate likely periods of channel cutting. Diagonal shading indicates periods of dry climate. Reprinted from Quaternary Science Reviews, 14, Thomas & Thorp, pp. 193–207, 1995, with kind permission from Elsevier Science Ltd, The Boulevard, Langford Lane, Kidlington OX5 1GB, UK.

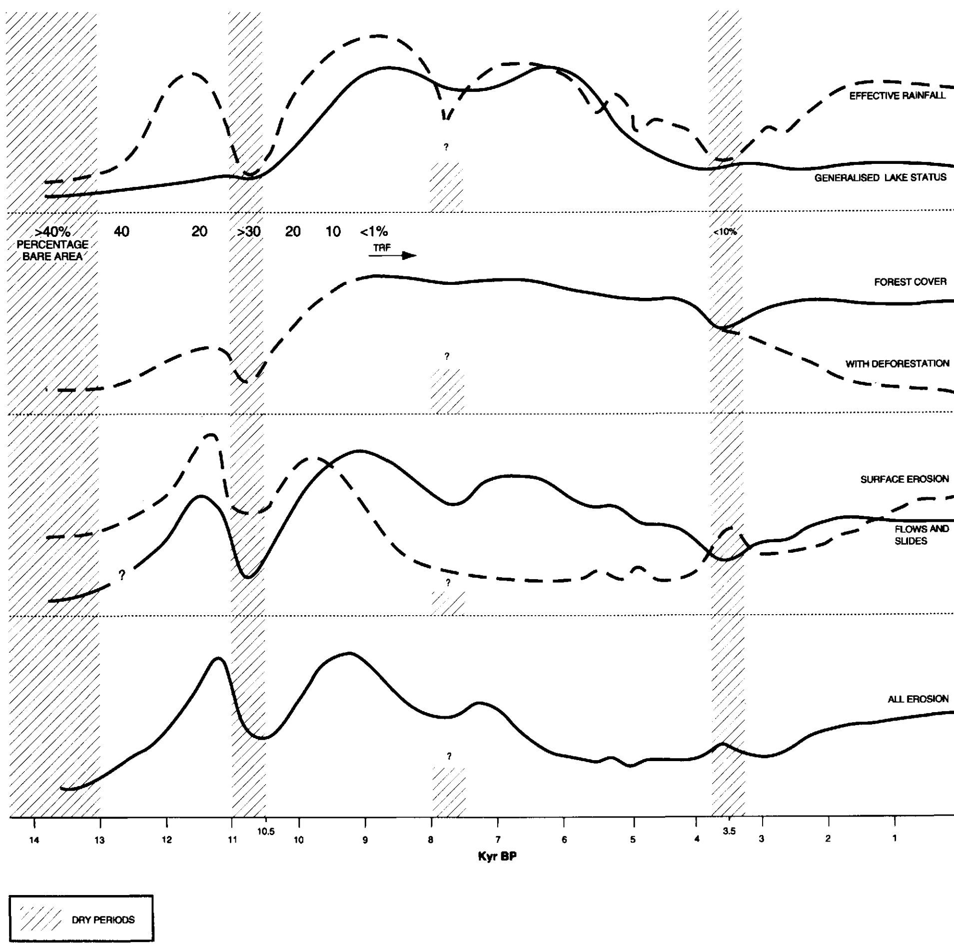

Fig. 2. Climate, environment and land surface processes in the humid tropics during the last 14 ka. Reprinted from Quaternary Science Reviews, 14, Thomas & Thorp, pp. 193–207, 1995, with kind permission from Elsevier Science Ltd, The Boulevard, Langford Lane, Kidlington OX5 1GB, UK.

It can be hypothesized, therefore, that the early Holocene pluvial period, as it progressed, may have led first to further debris flows and fan accumulation, then subsequently to fan and terrace dissection and channel cutting as the amount and calibre of load supplied to rivers diminished after *c.* 9.5 ka BP. But the forest cover did not prevent the development of deep seated landslides and the episodic delivery to local fluvial systems of copious sediments in flood events.

Since the early Holocene pluvial many small channels have become infilled with sediment, creating the widespread alluviated and partly channelless valleys or 'dambos', many of which conceal buried channels. These channel fills have different ages, perhaps recording singular events of floodplain destruction and reconstruction, but the reduction of stream power in the mid-Holocene is possibly one reason for their wide occurrence.

The accompanying figures (Figs 1 & 2) attempt to summarize the findings of this review. But they are presented in the knowledge that they involve some speculation and that the data base for determining the sedimentary and environmental histories of tropical floodplains remains inadequate. A concerted programme of research directed towards the analysis of representative catchments in different tropical environments is required in order to provide well-constrained chronologies of alluvial and colluvial deposits.

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Palaeo and historical flood hydrology, Indian Peninsula

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Abstract: The Indian Peninsula has many rivers that flow through bedrock gorges, providing excellent sites for the emplacement and preservation of late Holocene slackwater sediments and palaeostage indicators. Preliminary surveys of palaeoflood hydrology at these sites and the available historical data (since 1700) indicate considerable spatial and temporal variability of the largest flood events. There appear to be complex associations with the monsoon and with the El Niño–Southern Oscillation phenomenon that will require further study as more palaeoflood records are developed at the study sites. Data available thus far indicate a modern (post-1950) epoch of very large floods that may exceed the magnitudes of events over the last millennium.

In the last two decades the significance of palaeoflood and historical records in flood frequency analysis has received greatly increased attention by engineers and hydrologists. Both palaeoflood and historical records provide additional information that complements short-term systematic records (Baker *et al.* 1988; Jarrett 1991; Georgiadi 1993; Frances & Salas 1994). In India, where the hydrological records are extremely short, the value of historical and palaeoflood data could be considerable. Geomorphological studies of Indian Peninsula rivers in the last few years have shown that palaeoflood data have potential as indicators of long-term variability in monsoon and tropical storm activity related to regional and global phenomena.

Flood hydrology of the Indian Peninsula

The Peninsula region of India is characterized by long, low-gradient rivers draining east from the Western Ghats. In the north lie the two largest west-flowing rivers: the very flood-prone Narmada and Tapi (Fig. 1). The peninsular rivers are monsoonal in nature, and the hydrological characteristics of these rivers differ markedly from those of humid river systems. These rivers are characterized by an extreme flow variability, including very severe floods during the monsoon season (Rajaguru *et al.*

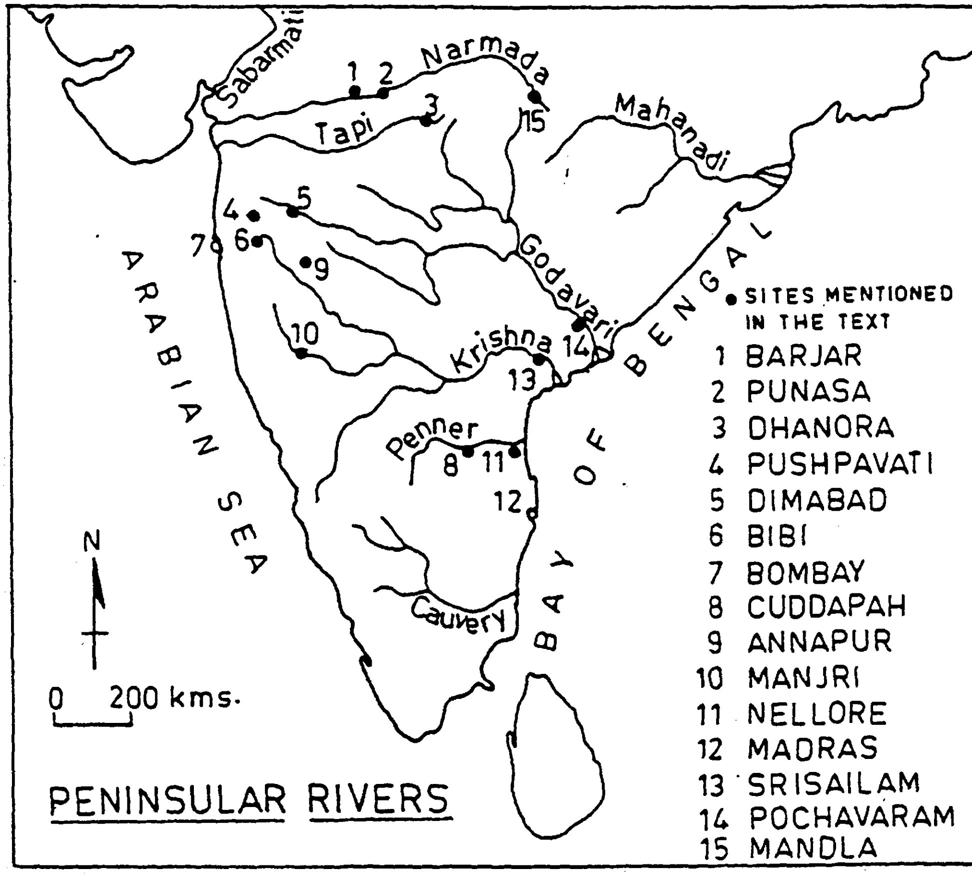

Fig. 1. Study sites in the Deccan Peninsula region.

1995). Monsoon-related storms are the major cause of high-magnitude floods in the modern records. Available gauge data indicate that high-magnitude floods are relatively common, with the most recent major events occurring in 1953, 1959, 1961, 1968, 1970, 1972 and 1984. The magnitude of megafloods on the Narmada (59 000 $m^3 s^{-1}$) and the Godavari (78 000 $m^3 s^{-1}$) rank among the global highest for

Table 1. *Peak discharges of Peninsular rivers*

River	Catchment area (km^2)	Total length (km)	Peak discharge on record ($m^3 s^{-1}$)
Mahanadi	132 100	858	44 827
Godavari	313 389	1465	78 800
Krishna	259 000	1400	33 810
Tapi	65 145	724	41 700
Narmada	98 796	1057	59 400
Cauvery	87 900	1340	12 913
Penner	55 213	597	14 716

their respective drainage areas (Rodier & Roche 1984), and remarkable peak flows occur on other rivers as well (Table 1).

Historical records

Historical flood information prior to the instrumental records includes (1) written historical accounts and the Imperial Gazetteers, (2) high-water marks on buildings, temples and bathing ghats, and (3) verbal communication from the local people.

Table 2. *Late historic floods, AD 1700–1900*

Year	River(s)	Remarks
1714*	Sabarmati	Damage to the city of Ahmedabad
1727*[f]	Tapi	Boats sailed over Surat city wall
1739	Sabarmati	Damage to the city of Ahmedabad
1755	Sabarmati	Ahmedabad damaged
1776*[p]	Tapi	Flood water reached Surat city wall
1782	Tapi	Flood due to cyclone
1810	Tapi	Flood on Tapi
1817*	Krishna, Bhima	Fields inundated
1822*[p]	Tapi, Narmada	65 villages destroyed along Tapi
1823*[f]	Narmada, Bhima	Crops destroyed
1825†	Krishna, Bhima	Flooding due to heavy rains
1835†	Tapi	Floods due to passing cyclone
1837†	Tapi, Narmada	Floods on Tapi and Narmada
1843*[f]	Tapi	Floods on Tapi
1849*[f]	Tapi	Floods on Tapi
1851*[p]	Krishna, Bhima	Crops destroyed in several villages
1852†	Penner	Widespread damage
1855†	Mahanadi	Many people washed away
1866*	Mahanadi	Large area inundated
1868*†	Sabarmati	Disastrous floods
1872*[p]	Tapi, Girna	Destruction to property, bank overflow
1873†	Tapi	Floods on Tapi
1874*†	Penner, Krishna	Banks overflowed, widespread damage
1875	Tapi, Sabarmati	Two iron bridges and part of town washed away
1876†	Penner, Tapi	Water level in Tapi higher than 1849
1878*†	Tapi, Narmada	Floods due to excessive rains
1882†	Krishna, Penner	Penner overflowing by 28.5 feet
1883†	Tapi, Mahanadi	Town of Surat under 20 feet of water
1884*†	Tapi	Floods on Tapi
1891*†	Tapi, Narmada, Godavari	Extensive inundation in Godavari
1892	Mahanadi	Floods in Puri District
1893	Penner	Widespread damage
1894	Tapi, Narmada	Floods on Narmada, Tapi and Girna
1896*†	Penner, Krishna	Floods on Krishna and Penner
1900*†	Godavari	Extensive inundations

* El Niño events.
† ENSO events after Quinn & Neal (1987) and Quinn (1990).
[p] El Niño/ENSO previous year.
[f] El Niño/ENSO following year.

According to the written historical accounts, several devastating floods occurred on the Narmada, Tapi, Mahanadi, Krishna and Godavari Rivers and their tributaries during the last 300 years. Although the relative magnitudes or the stages were not recorded, field evidence indicates that the flooding prior to 1900 was, by and large, lower in magnitude than modern extreme floods.

A summary of the historical data (Table 2) indicates a large spatio-temporal variability in flooding between 1700 and 1900. In addition to the historical descriptions and reports of the floods themselves, historical accounts also make extensive reference to severe damages to tanks, bridges and towns due to disastrous cyclones and storms. For instance, cyclone-related destruction occurred in 1785 (Madras), 1787 (Godavari), 1803 (Cuddapa), 1820 (Nellore), 1832 (Godavari), 1851 (Cuddapa and Bellary), 1857 (Nellore), 1874 (Arcot) and 1876 (regional).

Palaeoflood records

Palaeoflood information is generally obtained from the geological evidence left by floods as slackwater flood deposits (SWD), and as palaeostage indicators (PSI) such as silt lines and scour lines (Baker *et al.* 1988). Rivers of the Indian Peninsula (Fig. 1) provide excellent geomorphic situations for the deposition and preservation of palaeoflood slackwater deposits (Table 3). Where these rivers are confined in bedrock canyons, vertical accretion of fine-grained sediment from suspension becomes the dominant mode of flood-plain formation. The most common site for the deposition of mainstream-derived SWD is at the mouths of small tributaries in canyon and gorge sections. Such tributaries have very small drainage basins, and are relatively inefficient for eroding and removing the slackwater sequences. Other common geomorphic settings are along the inner margins of gorge meanders. The majority of SWD reveal multiple flood sedimentation units, delineated by abrupt changes in texture, colour and composition (Kale *et al.* 1994). The individual flood units are often capped by silt–clay, organic detritus or colluvial material. At some places, archaeological materials (hearths, pottery, bangles, etc.), leaf litter, and flood debris occur at the tops of individual flood sedimentary units. The textural and structural characteristics of the sediments suggest that these units represent very rapid deposition from suspension.

Table 3. *Palaeoflood sites on Peninsular rivers*

River	Gorge	Lithology	Type of PSI–SWD evidence
Narmada	Marble	Dolomite	BB, SL, SWD
Narmada	Punasa	Quartzite	SWD, BB, SL
Narmada	Dhadgoan	Basalt	SWD, SL
Tapi	Bhainsadehi	Basalt	SL, BB, SWD
Aner	Satrasen	Basalt	SWD
Puspavati	Mandvi	Basalt	SWD, BB, SL
Choral	Barjar	Quartzite	SWD, BB
Godavari	Papi	Gneisses	SL, SWD
Krishna	Jaldurgh	Granite	SL, BB
Krishna	Srisailam	Quartzite	SWD, BB, SL

SWD, slackwater deposits; BB, boulder berms; SL, scour/tree line.

Many SWD gorge sites have now been identified on the Indian peninsular rivers (Table 3). Sites on the Narmada River near Punasa (Fig. 1) were first discovered by V.R. Baker in February 1988. The best known SWD records occur at Narmada-Chain Nala junction (Kale *et al.* 1994; Ely *et al.* in press). Backflooding from the Narmada seems to have favoured deposition of flood deposits at this site. The series of flood deposits at the Chain Nala were the main source of palaeoflood information for the central Narmada River. The flood deposits consist of 4 to >22 flood units (Fig. 2). Radiocarbon dates from the flood units ranged between ultramodern and <1700 years BP (Ely *et al.* 1993, in press).

Calculations of the discharges associated with the Punasa Gorge SWD–PSI evidence were performed using the step-backwater computer modeling procedures.

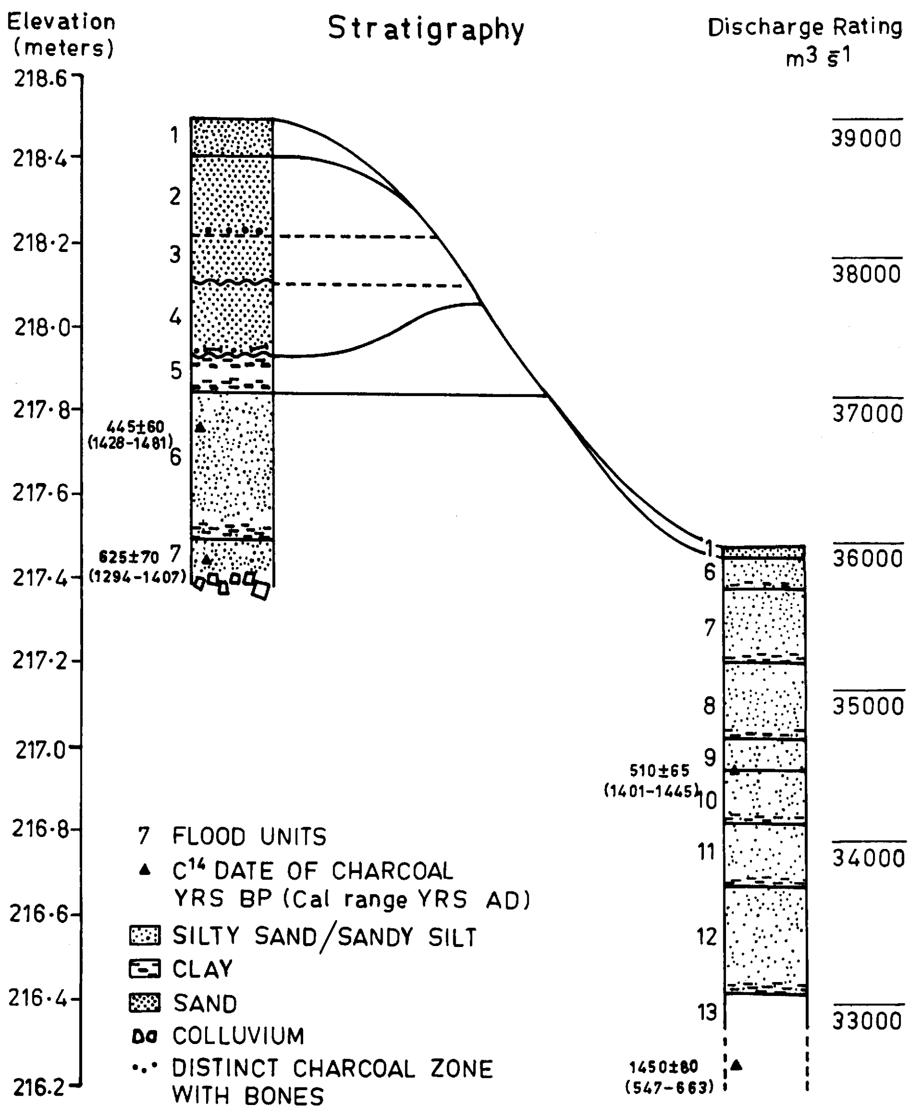

Fig. 2. Stratigraphic section of the Punasa site on the Narmada River (modified from Baker *et al.* 1995).

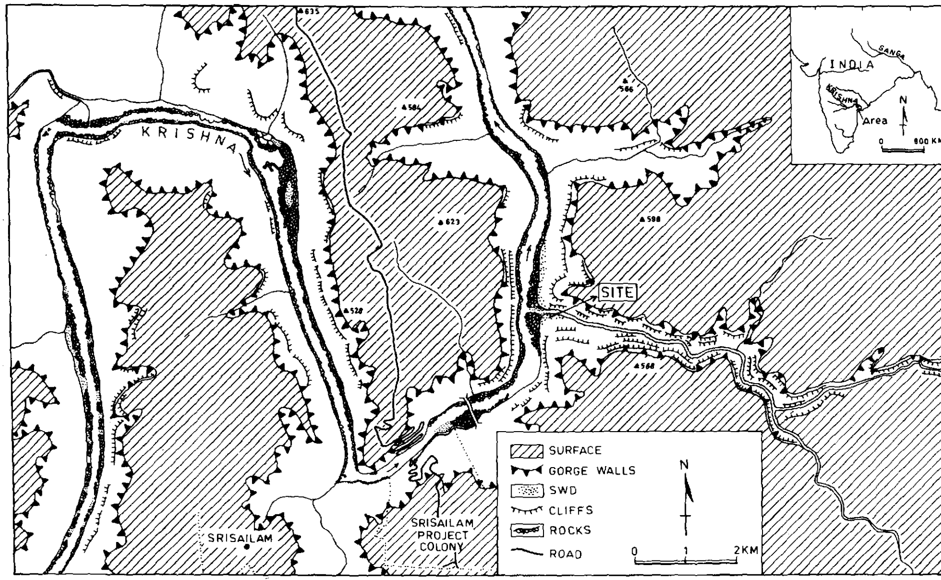

Fig. 3. Geomorphic map of the Srisailam Gorge on the Krishna River, based on IRS image interpretation and field data.

The appropriate reach was selected according to criteria described by O'Connor & Webb (1988). Results indicate that the highest SWD–PSI evidence is associated with flood discharges close to 60 000 $m^3 s^{-1}$ (Kale *et al.* 1992, 1994). These results suggest that the estimated discharges are lower or comparable to the largest modern floods. Other reaches on the Narmada, near Mandala, at Punghat and near Dhadgoan, have been found to have records of multiple floods. Similarly, several SWD sites have been found on the Tapi River, downstream of Dhanora (Fig. 1).

In the Eastern Ghats several sites have been identified on the Krishna and Godavari Rivers. In the Papi Gorge, upstream of Pochavaram (Fig. 1), 3–4 m thick slackwater flood deposits were observed in the lower order tributaries of Godavari. An interesting 10 m SWD section with nearly 60 individual flood units was found at the mouth of Nekkanti Vagu and the Krishna River near Srisailam Dam (Fig. 3). In the upper Krishna Basin, eighteenth century pottery has been discovered in the slackwater deposits. Similarly, radiocarbon dates from Annapur (770 ± 80 years BP) and Bibi (180 ± 90 years BP) represent high-magnitude floods in the upper Krishna Basin. At Manjri, on the Krishna River (Fig. 1), human skeletons have been discovered below 8 m flood deposits of about 4000 years in age. Further, sedimentological evidence of high magnitude floods in the Chalcolithic period (*c.* 3000 years BP) has been discovered at Daimabad in Godavari Valley (Rajaguru 1986).

At most sites on the peninsular rivers, the modern floods have been recorded to be higher than the top of a well-developed sequence of late Holocene slackwater sediments. On the Narmada and Tapi the deposits emplaced by the modern floods are coarser than those of the late Holocene slackwater deposits (Fig. 2). This is consistent with the modern (post-1950) events being larger than events in the past millennium (Baker *et al.* 1994; Ely *et al.* 1993, in press).

Monsoonal palaeoclimate

High-magnitude floods on the peninsular rivers in this century were invariably associated with monsoon depressions and cyclones originating in the Bay of Bengal and the Arabian Sea (Ramaswamy 1985; Kale *et al.* 1994). It is therefore logical to conclude that the palaeofloods provide direct evidence of individual, intense tropical storms during the monsoon season. However, in the absence of reliable and continuous historical data on the intensity and/or frequency of tropical storms it is difficult to interpret the hydroclimatic conditions. It is also necessary to verify whether the relationship between floods and the Bay depressions holds for the historical period.

Atmospheric teleconnections, such as the links between the Indian monsoon and the El Niño and Southern Oscillation (ENSO), may explain climatic aberrations. The major hydrological manifestation of ENSO events are anomalous periods of flood or drought (Baker 1991). Analysis for the Narmada Basin reveals that, out of 17 large flood events between 1891 and 1991, six were associated with years dominated by El Niño conditions, and five events occurred in the year following an El Niño year. The data given in Table 2 also show that many of the historical flood events coincided with El Niño/ENSO events.

Analysis of the climatic records in India indicate that, on a decadal scale, there have been four major climatic epochs: the periods 1885–1920 and 1950–1990 characterized by frequent droughts and the periods 1871–1884 and 1921–1950 being practically drought free (Mooley & Parthasarthy 1984; Kumar 1993). It is interesting to note that some of the largest floods on some peninsular rivers occurred during periods of rainfall deficiency. As suggested by Gordon *et al.* (1992), the increased flood activity caused by increased intensity of tropical storms can also be attributed to significant warming of 0.6°C/100 years in annual maximum temperature over India during the period 1901–1987 (Kumar 1993).

In recent years many attempts have been made to reconstruct long series of rainfall for the past few centuries. Climate reconstruction of the Indian monsoon, using tree ring data for the period 1602–1870, indicates that there was an excess rainfall epoch during the period 1610–1635 and a period of frequent droughts during a major part of the nineteenth century (Pant *et al.* 1988). Variation of monsoon rainfall by decades over the past millennium was interpreted by Bryson & Swain (1981) from pollen data in a Rajasthan palaeolake to indicate a period of relatively high monsoon rainfall between AD 1000 and 1200 and a period of low monsoonal rainfall *c.* AD 1600–1900. On the basis of foraminiferal studies of cores from the shelf region off Karwar, Nigam (1993) identified periods of high monsoon precipitation around 280, 840, 1610 and 2030 years BP and significantly dry climatic episodes around 420, 910 and 1680 years BP. The study also reveals significant variations in monsoonal precipitations and the existence of cyclicity of approximately 77 years. Using chemical tracers, such as organic matter, U/Al and Mn/Al ratios, Somayajulu *et al.* (1994) have inferred that between 200 and 700 years BP the monsoon was stronger and there was high fresh water input from Narmada and Tapi Rivers. The above studies, as well as the historical and palaeoflood data presented earlier, indicate that flood-generating monsoonal rainfall are temporally clustered but vary spatially. Further, the data show that the largest floods do not necessarily coincide with the periods of increased monsoon intensity. Clearly, more research needs to be done to evaluate how the spatio-temporal variations in modern, historical, and palaeofloods might represent long-term fluctuations in the monsoon.

Conclusions

Recent palaeoflood hydrological work on Indian Peninsula rivers, summarized in this brief survey, indicates a great potential for understanding long-term variability in monsoon and tropical storm activity. These phenomena are of global importance for understanding future environmental change (Baker 1994, 1995). Excellent sites for preservation of slackwater deposits and palaeostage indicators occur in gorges of the central Indian rivers, which are characterized by floods that achieve world maxima for a given drainage area. Some work has been accomplished on the Narmada River palaeoflood hydrology (Kale *et al.* 1994; Ely *et al.* in press), and much more opportunity exists to generate long and detailed palaeoflood records on rivers described above.

Preliminary indications from historical data (Table 2) and from preliminary surveys of palaeoflood sites indicate a possible modern increase in the incidence of large flood-producing storms. This is an initial observation that will need to be confirmed as various palaeoflood and historical datasets from the region become more complete. These datasets should indicate long-term variations in the spatio-temporal distribution of large floods in relation to regional and global phenomena. The data will also prove important in regard to modeling of the changing hydroclimate (Baker 1995).

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A preliminary palaeohydraulic model applied to late Quaternary gravel dunes: Altai Mountains, Siberia

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Abstract: The morphology and granulometry of large 'fossil' late-Quaternary gravel dunes at Kuray in Siberia are used to estimate the hydraulic conditions associated with dune building and stabilization. To accomplish this, an algorithm is tested which provided reasonable predictions of the flow characteristics measured above small gravel dunes in the Toutle River, Washington State. Palaeoflow depths and velocities above the 'fossil' dunes as the bedforms finally stabilized are estimated using a simple force balance for initial motion and the partitioning of the flow resistance between the drag owing to the duneform and grain resistance. Flow parameters for peak discharge conditions when dunes actively developed are assessed with respect to the dune–upper plane bed transition and the entrainment of over-passing blocks. These data, coupled with an assessment of the lateral extent of the flood, provide estimates of the discharge magnitude of the glacial out-break flood which was responsible for the dunefield. Sensitivity analyses indicate the stability of the estimates and demonstrate that the dunes were initiated and finally stabilized when the flood was of the order of $5 \times 10^4\,m^3\,s^{-1}$ and $2 \times 10^4\,m^3\,s^{-1}$ respectively. The flood peak need not have exceeded $750\,000\,m^3\,s^{-1}$ with a maximum depth of a few tens of metres.

Accurate estimates of the flow characteristics of Quaternary palaeofloods in part depend on the correct interpretation of preserved flood landforms and the stratigraphic sequences in diluvial deposits (e.g. Smith 1993). A number of palaeoflow reconstructions of late Quaternary floods from glacial Lake Missoula in North America and glacial Lake Kuray–Chuja in Siberia have been made using the altitude of flood bars as a control on water levels in computer-based simulations of flood waves (e.g. Baker & Bunker 1985; O'Connor & Baker 1992; Baker *et al.* 1993).

In a similar vein, other relict flood bedforms may provide clues as to local palaeohydraulic conditions. The interpretation of gravel dunes has previously proved useful in one investigation, where high-water marks constrained a slope-area analysis of flow above Missoula dunefields (Baker 1973). Recently, Carling (1996) has detailed the morphology and internal sedimentary structure of Late Glacial large gravel dunes found at six localities in the Katun River basin of South-Central Siberia (Fig. 1). These dunes were created by outbreak floods from glacial Lake Kuray–Chuja formed when local mountain ice sheets impounded the drainage of intermontane basins (Rudoy & Baker 1993). It is tempting to directly associate the development of the dunes to peak discharges as determined by step-backwater

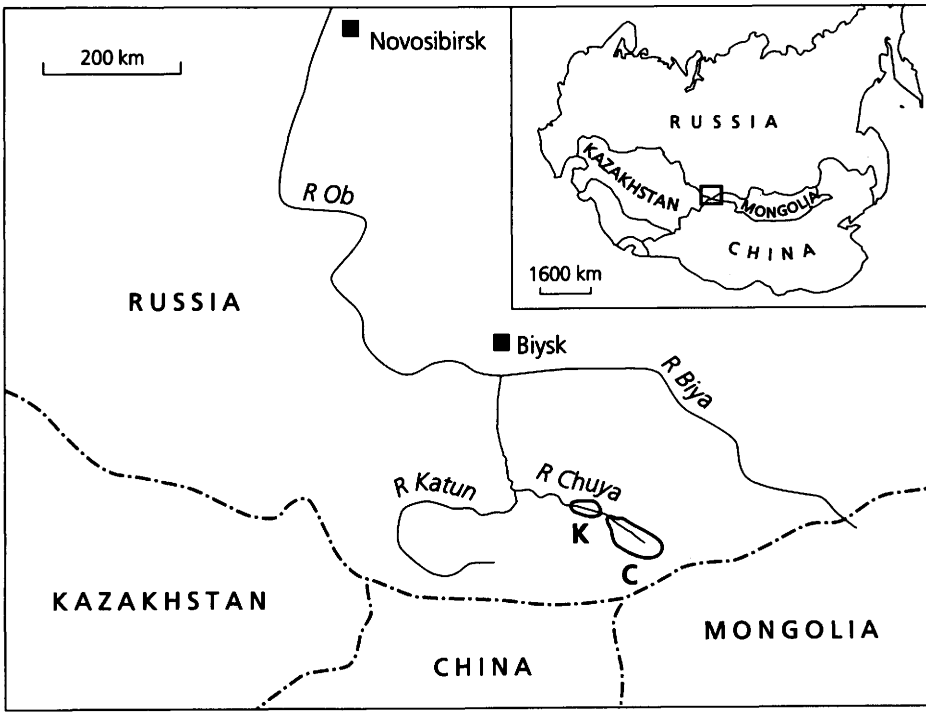

Fig. 1. Location of the Katun River basin in the Russian Federation. K, Kuray Basin; C, Chuja Basin.

modelling (Baker *et al.* 1993), but some means of estimating discharge from an appreciation of dune dynamics would be desirable.

Dunes in Siberia include local three-dimensional cuspate and lunate shapes (*sensu* Allen 1982) together with more common and extensive simple two-dimensional transverse structures. The varieties of form and bedding reflect differences in the interaction of flow and sediment transport and can provide clues as to the nature of the palaeofloods. At present it is not possible to model the fine detail of the flow because understanding of the coupling of flow and bedform generation in coarse material is poorly understood. Nevertheless, for geological interpretation, it is usually only necessary to determine the approximate scale of palaeoflows rather than absolute values. Consequently, simple models might provide an indication of the probable depth, mean velocity, discharge and the degree of uncertainty of the estimates. This specification requires a model which is physically based but which does not require data which are not attainable from 'fossil' dunefields.

In this paper, a simple model is outlined, based on partitioning form- and grain-friction effects of flow over the back of a steep (equilibrium) two-dimensional dune. The influence of the wake of the dune on the development of the dune-train downstream is neglected but it is assumed an identical contiguous dune exists upstream; the wake of which extends over the model dune. The strength of the model is that it gives reasonable estimates of the observed flow field over small transverse gravel dunes in the Toutle River reported by Dinehart (1992*a*, *b*). When applied to the Siberian dunes at Kuray, sensitivity analyses demonstrated that only a limited range of palaeo-discharge estimates is possible given a physically reasonable parameter range. Further testing and refinement of the procedure is planned so it is hoped a full description will be published in due course. In the section immediately following, the controls derived from field data are outlined as a precursor to constraining simulations.

Rationale

The Siberian dunefields are distinctive topographic features which consist either of series of two-dimensional (2D) transverse structures, or three-dimensional (3D) lunate and cuspate bedforms (Carling 1996). The former find their most impressive expression in the Kuray Basin where a dunefield is 3.3 m in streamwise extent with continuous individual crestlines perpendicular to the palaeoflow up to 2.4 km in span. Three-dimensional dunes have shorter spans up to *c.* 70 m. Many dunes are steep and approach equilibrium as defined below, whilst others are of low height compared with their length and are termed 'diminished' dunes. Many of the steep dunes have a distinct humpback long profile (Fig. 2) with a lower stoss slope typically greater than 10° reducing steadily to only 1° or 2° close to the crest. Lee-slope angles tend to be monotonic and less than the angle of repose.

Carling (1996) hypothesized that the dynamics and resultant morphology of equilibrium gravel dunes in a palaeoflow would be similar to that of equilibrium dunes developed in coarse sand by fluvial bedload transport without suspension. The latter case is well known both from field, laboratory and theoretical study. Four propositions logically follow which require support from field data.

(1) If Siberian gravel dunes consistently attain a maximum steepness then this might reflect an equilibrium morphological response to the palaeoflow.
(2) If grain size of sediment composing dunes is not a constraint on duneform then (given sufficient water depths and time for morphological adjustment) gravel dunes should develop similar height to length ratios as equilibrium fluvial sand dunes (Ashley 1990).
(3) The internal structure of the gravel dunes must be interpretable as owing to dune migration.
(4) If the above propositions hold true, as a first approximation the flow structure above a gravel dunefield would be analogous to that over equilibrium dunes

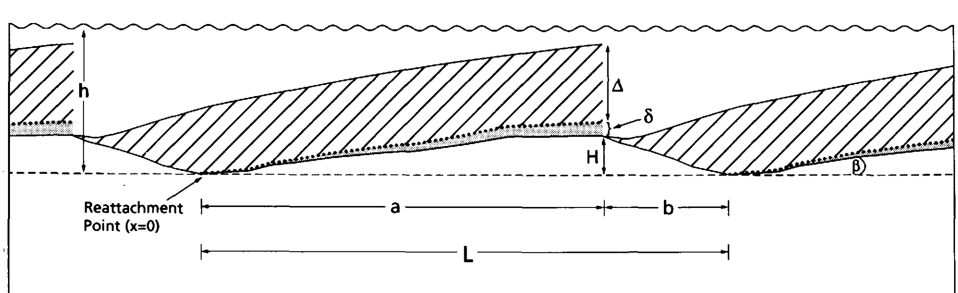

Fig. 2. Definition diagram of separated flow over two-dimensional gravel dune at Kuray. Length = 200 m and height = 16 m.

with bedload transport of coarse sand albeit with adjustment for grain roughness. Equations developed to derive bulk flow parameters over coarse sand might then be modified and applied to estimate the flow properties over gravel dunefields.

Dune morphology

To test propositions (1) and (2) a scatter plot of gravel dune height and length data was developed. Distinctly 2D steep dunes may be separated from discrete steep 3D dunes (Fig. 3) by the provisional discriminator:

$$H = 0.0073L^{1.5} \tag{1}$$

which is curve b in Fig. 3. Also shown are data for Missoula dunes (Baker 1973) which mainly represent steep pebble or cobble bedforms in relatively 'deep water'. The scatter in the data in part reflects the fact that many of the longer dunes in particular may have been depth limited or were bedforms not in equilibrium with the incident flow. Interest however rests in (a) whether data approach an upper limit of steepness, which may be regarded as representing equilibrium conditions, and (b) whether 3D dunes are steeper than 2D dunes (cf. Best 1996) and (c) whether

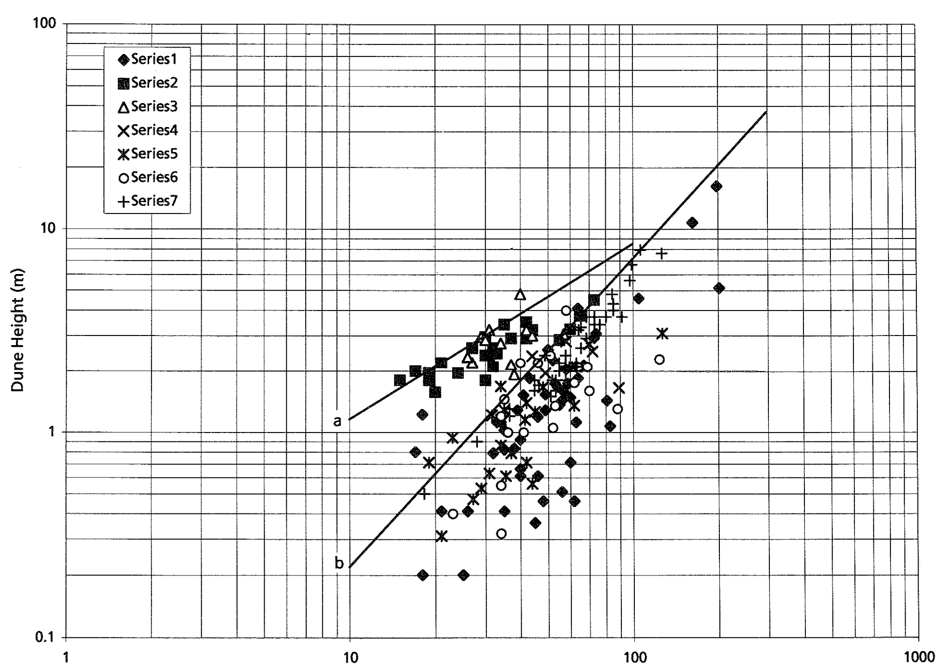

Fig. 3. Height–length data for a variety of two-dimensional and three-dimensional gravel dunes in the Katun River Basin (Russian Federation) and Missoula (USA). *Primarily two-dimensional dunes*: S1, Kuray; S4, Platovo; S5, Lower Kara Kjöl; S6, Akturu; S7, Missoula. *Primarily three-dimensional dunes*: S2, Upper Kara Kjöl; S3, Little Jaloman. Curve **a** after Ashley (1990): $H = 0.16L^{0.84}$. Curve **b** after Carling (1996): $H = 0.0073L^{1.5}$.

steepness is similar to that noted in modern fluvial sand dunes. The only clear exceptions to the discriminator are a few small 2D dunes at Kuray. Equation (1) is not presented as an absolute delimitation but provides a useful bound to data for presence purposes.

In a similar vein, the upper curve purports to delimit the steepest equilibrium sub-aqueous dunes formed in sand (Ashley 1990). Ashley's function (Curve a in Fig. 3) essentially defines the range of steepness ratios: $H/L = 0.1–0.08$; indicative of an equilibrium steep duneform (e.g. Allen 1982; Yalin 1977); the 3D gravel dunes and the largest of the 2D dunes approach this limit. Steepness is important as it largely determines the form wake-induced roughness of dunefields in deep flows (Engelund 1966). Usually longer sand-dunes are relatively less high (e.g. Allen 1968) as is evident from the exponent (0.84) in Ashley's function. The data for the Siberian 3D gravel dunes do not controvert this trend whilst Thiel (1932) and Galay (1967) indicate that 3D gravel dunes lengthen in progressively shallowing flow.

This progressive flattening may be shown more clearly by rewriting Ashley's relationship to define steepness as a function of length:

$$H/L = 0.16L^{-0.16} \tag{2}$$

The small negative exponent in Equation (2) may reflect interaction of the bedform with the boundary layer developing over large sand dunes or a depth limitation influencing sediment transport. Although there are insufficient data to draw a firm conclusion, an upper bound to the gravel 3D dunes in Fig. 3 conveniently may be represented by changing the constant in Equation (2) from 0.16 to 0.18. In contrast, the height of 2D gravel dunes clearly increases as dunes lengthen, which may indicate development in deepening flow. However there are insufficient data to determine if the discriminator should become asymptotic to Ashley's function in the case of very large dunes. It is proposed that for the present purposes the steepness of 2D equilibrium gravel dunes may be predicted by:

$$H/L = 0.0073L^{0.5} \tag{3}$$

It was concluded that propositions (1) and (2) are supported by the data in Fig. 3.

Dune sedimentology

All the dunes consist essentially of simple decimetre-thick tabular gravel foresets dipping down current to meet a planar substratum at an acute angle (cf. Rubin 1987, fig. 8). Individual distinct cross-beds of cobbles grade upwards into (or are conformably replaced by) well-sorted pebble beds. The angle of the beds (e.g. 12° to 25°) is below the angle of repose (*c.* 32°). All these thin layers are interpreted as being deposited from thin bedload sheets periodically passing over the dune crests (Carling & Glaister 1987). Dynamic sorting within discrete bedload sheets typically results in coarse particles concentrated at the leading edge and progressive fining upstream (Whiting *et al.* 1988) so the pulsed input leads to alternate cobble–pebble bedding on lee slopes (Siegenthaler & Huggenberger 1993). Whereas the surface fabric may only reflect hydraulic climate as the dunes stabilized (e.g. Byrne 1963) clast orientation measured in sub-surface exposures gives some clue as to the nature of active bedload transport. Clast orientation was extremely variable (with only strongly elongate particles aligned flow parallel) which is typical of coarse bar fronts in high velocity flows (Johansson 1963; Rust 1972; Carling & Glaister 1987) wherein clasts roll and saltate in a flow with little suspended load.

In each dunefield over-passing 'erratic' blocks ($d_i = 0.2$–3 m) were commonly located near dune crests or were stalled on the stoss slopes. Often these were imbricated and associated with a horse-shoe vortex scour hollow in the cobble-gravel. The mechanics allowing these blocks to slide or to stall on the backs of dunes provide clues as to the maximum entrainment forces during active dune migration. These blocks are also indicators of lower-stage flow regime. Fahnestock & Haushild (1962) noted from flume studies that the progressive imbrication of out-size blocks in smaller bed material was associated with horseshoe vortex scour and lower-stage flow regime whilst upper-stage regime promoted plane beds, dislodged blocks downstream and infilled scour hollows.

These observations and interpretations (developed more fully by Carling 1996) are consistent with proposition 3.

Palaeohydraulic model

Constraints on model development. Proposition (4) cannot be fully tested owing to the lack of data concerning gravel dune dynamics. Nevertheless, a logical argument is constructed which produces palaeoflow data which can (i) be compared to know constraints on fossil dune dynamics and (ii) be tested against available field data on modern gravel dune dynamics (Dinehart 1992*a*, *b*). Calculations were derived for the case of the Kuray dunefield and then a limit test made using Dinehart's data (1992*a*, *b*).

Logical constraints with respect to (i) include the following. During vertical growth in dune form, predicted velocities on stoss and crest must be sufficient to entrain coarse gravel but should not be so great as to induce crestal flattening and upper-stage plane beds. Predicted depths must be greater than dune height and at least of the order of the height (40 m) above the dunefield of flood-deposited boulders on flanking hills. A consideration of both the predicted depths and velocities must preclude upper-stage plane beds or the development of antidunes. Predicted velocities over the region of the crest of equilibrium dunes must also be in accord with calculations for transport of 'erratic' blocks over the crestal platform.

Only one kind of dunefield is envisaged for modelling purposes which is scaled by data from Kuray (Table 1). The model consists of a contiguous series of morphologically identical large two-dimensional dunes of height (H) and length (L) giving steepness values of *c.* 0.08– 0.1 in equilibrium with flow of depth h (Fig. 2). It is also assumed that a wake develops downstream of the crest of such steep dunes such that the form drag increases with the onset of separated flow. Dinehart (1992*a*) for example estimated the form drag over small gravel dunes in shallow flow increased from less than 10% to 50% as dunes steepened and consequently the flow separated or failed to recover fully downstream of the dune brinkpoints. However, Smith & McLean (1977) and Nelson *et al.* (1993) noted that the absence of flow separation over low and sparse sand dunes caused the form drag to increase to values greater than associated with high, closely spaced bedforms where separation was

Table 1. *Notation and basic field-derived input data required for calculations pertinent to the Kuray dunefield*

b	Leeside length – 60 m
d_{50}	50% grainsize – 32 mm
d_{84}	84% grainsize – 250 mm
d_b	Intermediate axis of overpassing block – 3 m
d_c	Short axis of overpassing block – 3 m
H	Dune height – 16 m
L	Dune length – 200 m
W	Span of dunefield – 2400 m
β	Bedslope at locations over dune stoss and crest
a_2	Long axis of reference ellipsoidal particle
b_2	Intermediate axis of reference ellipsoidal particle
c_2	Short axis of reference ellipsoidal particle
ρ	Density of flood water – initially assumed 1 g cm^{-3} but variable in sensitivity tests
ρ_s	Density of reference particle – 2.65 g cm^{-3}

present. Consequently, the diminished gravel dunes in the Katun Basin will require further consideration and are not modelled here. A further assumption of the model is that the hydraulic conditions over the crests of dunes of maximum steepness in fast flow are close to the upper-stage plane bed transition. Yalin (1977), Yalin & Karahan (1979) and Allen (1978) amongst others deduced that H/L and H/h are bell-shaped functions of the bed shear stress, with dunes reaching their maximum steepness close to the transition. This assumption was latterly shown to be reasonable from a consideration of initial motion of 'over-passing' erratics.

Governing equations. In the following text the fundamental equations are introduced in the logical order necessary to describe model formulation. However, this is not the order in which equations are necessarily utilized for calculations. The basic procedure is to use grain-size data to determine the bed roughness and shear velocity at threshold of motion above the dune crest. This information is used to construct a velocity profile in the internal boundary layer and subsequently in the wake layer over the dune crest. Velocities characteristic of the depth-averaged flow, say at 0.4 of the thickness of the boundary layer, or at a given height within the wake layer are then calculated for three conditions which for succinct argument are considered out of chronological order: (1) when the dune finally stabilized on a waning hydrograph; (2) when it initially grew from a plane bed and; (3) for active growth when the dune approached the upper-stage plane bed transition. Associated water depths are also estimated for each condition. Given an estimate of the lateral extent of the dune field, the product of mean velocity, depth and width gives an estimate of the order of magnitude of the palaeoflow at each of the three stages.

Although Wang (1994) has considered the additional effect of dune asymmetry, Engelund (1966) amongst others determined that the form-induced drag (f'') of large dunes is primarily proportional to the relative depth of flow (H/h) and the steepness (H/L), such that in deep flows over steep bedforms the relative-depth term is slight compared with the drag imposed by the stoss slope:

Theoretical consideration by Kennedy (1963) and Yalin (1977) indicate that equilibrium dune length is proportional to water depth, and so:

$$f'' = 4(H/h)(H/L) \tag{4}$$

$$L = 2\pi h \tag{5}$$

However, field observations have shown that there is some variability such that van Rijn (1982) amongst others suggested an approximate length scale for equilibrium fluvial sand dunes:

$$L = \alpha 2\pi h \tag{6}$$

with α being assigned a narrow range of values typically between 0.6 and 1.1. Support for a similar approach in the case of gravel dunes is provided by an analysis by the author of the data of Dinehart (1992*b*) which shows that for developing (rather than diminished) gravel dunes $a \approx 0.7$ (s.e. 0.05) falling to *c.* 0.4 for the steepest dune.

Combining Equations (4) and (6) gives general relationships for form drag, i.e.:

$$f'' = \alpha 8\pi (H/L)^2 \tag{7}$$

or

$$f'' = \frac{2}{\alpha\pi}(H/h)^2 \tag{8}$$

Convergence and divergence of separating flow over steep bedforms leads to development of compound velocity profiles consisting of an quasi-logarithmic internal boundary layer (of thickness $-\delta$) developing over the stoss-slope beneath the wake-layer (Δ in Fig. 2). The wake layer occurs due to flow separation at the crest of the bedform immediately upstream of the study dune (e.g. Dyer 1986). Above the flat crest a log-profile:

$$U_z = 5.75 u_* \log(30z/k_s) \tag{9}$$

provides a description of variation in velocity (U_z) of the internal boundary layer up to a reference level z_* where the velocity is denoted, U_{z^*} (Nelson *et al.* 1993; McLean *et al.* 1994). In the absence of data concerning flow structure over gravel dunes an approximate reference level is determined from an empirical relationship applicable to flow over steep sand dunes (Smith & McLean 1977; Nelson *et al.* 1993).

$$\frac{z_*}{z_0} = 0.1\left(\frac{\lambda}{z_0}\right)^{0.8} \tag{10}$$

where λ is the dune chord. In accord with proposition 4, λ is set equal to L in the model and the roughness length selected must be appropriate for gravel instead of sand, i.e.: z_0 for a coarse mobile gravel bed is $0.11 d_{84}$ (e.g. Dinehart 1992*b*) where d_{84} is about 250 mm at Kuray. A velocity characteristic of the wake, for example at 0.4Δ above the crest (cf. Nelson *et al.* 1993) is obtained assuming a further log-profile develops above the reference level. At the reference level the velocity of the top of the internal boundary layer and the base of the wake layer are the same such that the wake layer shear velocity (u_{*z^*}) is determined using Equation (9) with a wake-layer roughness length determined by adding skin and form-drag components (e.g. Engelund 1966; van Rijn 1984). The combined roughness of coarse-sand or gravel duneforms scales approximately as: $k_{s^*} \approx Hb/L \approx 20d_{84}$ (Vanoni & Hwang 1967; Dinehart 1992b). This procedure yields a k_{s^*} of about 5 m for Kuray dunes. Nelson *et al.* (1993), considering flow over fixed laboratory dunes, concluded that such a general approach should predict well the mean flow structure over steep bedforms which is all that is required in a model for geological application.

At its simplest, the threshold of initial grain motion and the transition for gravel beds from dune to upper-stage plane bed is commonly expressed in terms of the Shields parameter (θ); the non-dimensional shear stress for given grain-size and relative fluid density being given by:

$$\theta = \frac{\tau}{(\rho_s - \rho)gd_{50}} \tag{11}$$

However, by modifying particle shape and projection ratios etc. through the application of Equation (12) (described below) sensitivity analyses can be conducted with respect to the non-dimensional shear stress for initial motion and the plane-bed transition. Similarly the sensitivity of calculations to variation in fluid density owing to suspended sediment can also be considered; although here only the results for clear water are reported.

(1) Estimate of flow when dune stabilized. To estimate the palaeoflow when the dune finally stabilized an estimate of the critical shear stress at threshold of motion (τ_c) and a measure of bed roughness (k_s) are required. Once the dune was fully formed, stabilization for the bulk of the sediment only occurred again during falling stage as the dune ceased to migrate and pebble fractions were deposited as a drape above less mobile cobble fractions (Carling 1996). A characteristic grain size is taken as the d_{50} of the surface once stabilized. As flow shallows over the crest of very large humpback dunes, the surface drag exerted by the internal boundary layer over the extended crestal region is approximately equivalent to that over a flat bed (Raudkivi 1967) although the presence of a wake over the stoss may increase this slightly. The critical shear stress for ellipsoidal particles therefore may be defined as:

$$\tau_c = \frac{2V(\rho_s - \rho)g}{\psi C_d \pi (a_2 b_2 c_2)^{2/3}} \frac{\sin(\phi + \beta)}{\cos\phi} \frac{1}{[5.75\log(30z_p/k_s)]^2} \tag{12}$$

wherein parameters are defined by Carling *et al.* (1992), who also detail typical parameter values for horizontal pebble and cobble beds. The bed slope (β) is zero over the crest and positive over the stoss, so an increase in the critical shear stress associated with flat bed conditions will occur as β is increased. This is because the pivoting angle (ϕ'') through which a grain must rotate to be entrained on the stoss slope is equal to the sum of the flat-bed pivoting angle and the adverse slope ($\phi + \beta$). Given $u_{*_c} = \sqrt{\tau_c/\rho}$ and $U_{*_c}/U_z = \sqrt{f/8}$ for incipient motion on a flat bed, the characteristic velocity of the internal boundary-layer U_z at 0.4δ is given by Equation (9). The shallow depth of flow at incipient motion beneath a compound boundary layer is not known precisely but according to Raudkivi (1967) over an infinitely broad and flat crest (h_H) scales approximately as:

$$\frac{1}{\sqrt{f'}} = 2.03\log\left(\frac{11.09h_H}{k_s}\right) \tag{13}$$

where k_s for a stable pebble gravel is of the order of $3.5d_{50}$ (Thompson 1963). Given $L = 200$ m, $H = 16$ m, $d_{50} = 32$ mm (Table 1) and parameter values listed by Carling *et al.* (1992), the minimum near bed u_{*_c} was determined from Equation (13) to be $0.10\,\mathrm{m\,s^{-1}}$ such that $f' = 0.0313$ and h_H is of the order of 6.15 m. Consequently, total flow depth $h = h_H + H = 22.15$ m. With such a small depth above the crest it is unclear whether the wake layer would be well formed, and so two solutions for velocity can then be obtained from Equation (9) assuming either the internal boundary layer extends to close to the water surface or a wake-layer is present above the reference height z_* such that the reference velocity is at 0.4Δ. In either case, and allowing for some uncertainty in parameter values, a narrow range of bulk flow velocities is obtained: typically of the order of $1.5\,\mathrm{m\,s^{-1}}$. From Equation (7) the maximum form resistance $(f'') = 0.1609$, such that the grain resistance over the crest constitutes some 16% of the total flow resistance ($f = 0.196$). This result is in accord with both the field and laboratory studies of near-equilibrium sand-dunes (referenced above; $H/L = 0.05$ to 0.125) which indicate maximum drag of about 0.2. Further consideration of Equation (12) shows that for positions on the steep lower stoss, grain resistance increases so that cessation of motion would occur locally before the crest stabilized; which accounts in part for the crestal flattening observed in the field. As the dune stabilized, the discharge ($Q = h_H U_z W$) across the full span of the dunefield ($W = 2.4$ km) must have been of the order $2 \times 10^4\,\mathrm{m^3\,s^{-1}}$.

(2) Estimate of flow when dune first developed. This condition is poorly parameterized by field data at Kuray owing to the absence of any gravel plane-beds adjacent to the dune field. However, Dinehart (1992*b*) noted that for coarse gravel of equivalent size to that at Kuray, dunes did not appear on a plane-bed until the Shields parameter exceeded 0.11. Assuming this value is applicable at Kuray, the critical shear stress for incipient dune growth can be calculated from Equation (11) with variation of parameter space explored using Equation (12). Subsequent calculations to determine velocity and depth above a coarse mobile plane-bed ($k_s = 3.5d_{84}$) beneath a logarithmic boundary layer follow as outlined previously. These indicate that the velocity was about $2.8\,\mathrm{m\,s^{-1}}$ in a water body some 8 m deep such that the discharge was of the order of $5 \times 10^4\,\mathrm{m^3\,s^{-1}}$.

(3) Estimate of flow when dune was close to upper-stage transition. During active building of the dunes, the velocity and depth would have been greater than determined in (1) and (2). Peak flows can be ascertained with reference to two conditions; that is, first the threshold for sliding of over-passing blocks and secondly, the threshold for transition from duneform to upper-stage flat bed beyond which the duneform could not have been sustained.

Sufficient conditions necessary to allow the largest block ($d_b = 3$ m) to slide across the crest was evaluated using an expression and parameter values given by Carling & Grodek (1994):

$$U_{\min} = \frac{1}{(1 + 3k_{turb})}\left[\frac{2(\rho_s - \rho)gd_b}{\rho}\,\frac{\mu_f}{[C_L(d_b/d_c)\mu_f] + C_d}\right]^{1/2} \tag{14}$$

from which it can be deduced that the minimum over-passing velocity ($U_{\min}$) acting on the upstream face of the block was $6.5\,\mathrm{m\,s^{-1}}$ which approaches the condition:

$\theta_c = 0.3$ for gravel, a value commonly adopted for the transition to upper-stage plane beds in gravel (Engelund & Hansen 1967; Allen 1970; Bridge 1982; Carling & Glaister 1987; Dinehart 1992*b*). This correspondence of calculated velocity and near upper-stage plane-bed conditions is theoretically and experimentally consistent with the existence of dunes of maximum steepness (Fredsøe 1975; Yalin 1977; Yalin & Karahan 1979). Entry into Equation (9) using the over-passing velocity at a height of $d_c/2 = 1.5$ m as an initial reference to calculate the near-bed and subsequently the wake-velocity profile then indicates that U at 0.4Δ would be about $8\,\mathrm{m\,s^{-1}}$ in water typically 40 m deep (Equation (6)). However, the greatest uncertainty in any dune model is the associated water depth and so this is considered further.

For two-dimensional coarse-sand dunes of maximum steepness ($H/L \approx 0.084$) during active transport, Vanoni & Hwang (1967) indicated that the associated crestal depth is given by:

$$h_c = \left[\frac{1}{L}\left[H^2 10^{\left(\frac{\frac{1}{\sqrt{\eta f''}}+2.3}{3.3}\right)}\right]\right] - 0.5H \qquad (15)$$

where the curve-fitting parameter η decreases to between 0.6 and 0.8 for coarse-grained dunes of maximum steepness. The uncertainty in η indicates that the crestal depth at Kuray was between 50 m and 40 m with a relative depth H/h greater than 0.3. These depths are consistent with the presence of flood boulders on hillsides 40 m above the dune field as noted above. However it is unlikely that the hill-side boulders would have been deposited close to the water surface. Consequently, it is probable that at some time during the flood the water depth was greater than calculated but this condition need not equate to the either the period of maximum velocity or the existence of dunes of maximum steepness.

Raudkivi (1967) and Allen (1978, 1982) amongst many others, observed that the relative height of steady-state equilibrium dunes in coarse sand increased as a parabolic function of excess shear stress to a maximum of typically 0.3, before declining close to the transition. Dinehart (1992*b*) presents field data for coarse-gravel dunes which is consistent with Allen's (1978) analysis. Dinehart noted that as the shear stress increased the relative height exceeded 0.25 before diminishing again as the dunes achieved maximum steepness close to transition. Therefore a relative depth of about 0.3 seems reasonable, giving a peak Kuray discharge of about $750\,000\,\mathrm{m^3\,s^{-1}}$. The selected relative depth has to be preferred, as not only is it dynamically consistent, but it is in accord with Equation (6) and the observed maximum length of the Kuray dunes (180– 200 m).

Sensitivity analysis

Space limitations preclude a full report on sensitivity tests but variability owing to suspended load increasing the effective fluid density (Lee 1969) and decreasing the total friction (Raudkivi 1967) has been considered. At the threshold of motion bedload transport is vanishingly small, but at other transport conditions the influence of saltation on bulk flow was assessed by systematically varying the value of bed-roughness parameters (e.g. Smith & McLean 1977; Wiberg & Rubin 1989) using the data of Dinehart (1992*b*) for bed roughness variation in the presence of slight to intense bedload transport. Typically mean boundary-layer velocity varied by less than 13% in the presence of suspended load and saltating grains.

Model verification

The best test of the model is to compare model estimates with measurements of active gravel dunes. The most comprehensive data are for coarse gravels ($d_{50} = 33$ mm; $d_{84} = 60$ mm) in the Toutle River, USA (Dinehart 1992*b*). Dinehart observed that low-amplitude dunes developed first when the depth-averaged velocity equalled 1.89 m s^{-1} and water depth was 1.86 m. Dunes approached equilibrium with a maximum height of 0.45 m and steepness of 0.06. A maximum water depth of 2.44 m was associated with local velocities up to 3.3 m s^{-1} but depth-averaged velocity was 2.0 m s^{-1}. The flat stable-bed friction factor was 0.09 with $k_s = 3.5d_{84}$ and the total roughness associated with a dune-bed ($k_s = 20d_{84}$) was typically 0.2 implying a form roughness of 0.11. The dunes then diminished in height as the water depth in the dune troughs fell to 1.4 m and the lowest shear velocity recorded before measurements were terminated was 0.24 m s^{-1}.

From Equation (12) the critical shear velocity when the highest Toutle dune crests would stabilize is 0.14 m s^{-1} at which time the bulk velocity was about 1.4 m s^{-1} and flow depth above the crest was estimated as 0.83 m, giving a maximum depth of 1.28 m in the dune trough. Such shallow high-velocity flow over the dune crests immediately prior to stabilization explains the rapid evolution of diminish dunes during low flows as observed by Dinehart. Assuming the steepest dunes reached equilibrium and occupied at least 0.2 of the depth then maximum depth during transport was *c.* 2.25 m. Alternatively, Equation 15 with $f'' = 0.11$ and $\eta = 0.6$ gives a water depth of 2 m, but generally tends to under-predict the observed depths. The largest particles in bedload samples overpassing the dunes ranged between 100 mm and 150 mm. Solution of Equations (9), (11) and (14) indicate velocities during dune growth would range between 2.7 m s^{-1} and 3.7 m s^{-1}.

The calculations of velocity, although not exact, are reasonably consistent with Dinehart's (1992*b*) measurements. Depths are less well predicted using Equations (13) and (15) unless specific field data are utilized concerning the division of grain and form roughness. The model would have performed better if the Toutle dunes had been steeper. However, the rider should be added that this is the only test currently possible and the model is largely based on a consideration of the dynamics of equilibrium dunes developed in coarse sand and not gravel.

Discussion

The close agreement between the maximum steepness of 3D gravel dunes and sand dunes (Fig. 3) could conceivably be owing to data limitations. However, the conformity more probably is indicative of dynamic similitude across a broad range of sediment grain-size and hydraulic conditions. If substantiated this is an important conclusion indicating that in principle gravel and sand dunes might be modelled similarly where downstream translation is dominated by a bedload. The process similarity however must diverge where the interaction of a suspended load and the wake is critical in determining dune dynamics.

Although there is a degree of uncertainty with respect to the discharge associated with the building of the Kuray dunefield, exploration of a reasonable parameter range indicated that only a relatively narrow range of estimated values reasonably can be accepted. Modifying parameters within acceptable limits does not alter discharge estimates wildly. The procedure allows the calculation of three critical discharges during the flood hydrograph. That is the discharge when dunes first grew from a plane bed; secondly the 'peak' discharge close to the upper plane bed transition and finally; the minimal discharge at which stage the dunes stabilized. Evidently further research is required to extend flow competence relationships – established primarily through a consideration of the applied stress on the bed and pivoting models of entrainment – to determine discharge, bulk velocity and water depth.

Conclusions

A simple algorithm is proposed for palaeoflow analysis of 'fossil' dunes which in tests replicates, to a reasonable degree, velocities measured over gravel dunes in a modern river. Water depths are less well specified unless site-specific information is available concerning the division of grain and form roughness. Although the model only provides a crude approximation of real three-dimensional turbulent flows over complex dune forms it encapsulates the essential variation of forces responsible for bedform development without requiring data inputs and assumptions which cannot be derived from 'fossil' gravel dunefields and the known behaviour of modern coarse sand-dune systems. Specifically, dune height, length and grain-size data are required although information on the lee slope angle is useful to determine if flow separation occurred with a well-developed wake layer. It should be emphasized that the model indicates the probable scale of palaeoflows and not absolute values. Nevertheless, even allowing for parameter uncertainty, it appears that the discharge over the Kuray dunefield was probably about $750\,000\,m^3\,s^{-1}$ and need not have exceeded $1 \times 10^6\,m^3\,s^{-1}$ with a water depth of only a few tens of metres. If correct, this is an important conclusion indicating that the joint occurrence of exceptional depths and velocities are not a precondition for dune development in very coarse gravel. Further development of the model probably will lead to revision of these initial estimates and tests against modern gravel dune data from both laboratory and field would be desirable. Following these developments a similar model could be applied to other fossil dunefields such as those associated with Missoula (Baker 1973).

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Late Quaternary intra-continental river palaeohydrology and polycyclic terrace formation: the example of south Siberian river valleys

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Abstract: The subject of intra-continental river regime and related polycyclic terrace formation has been investigated in south and central Siberia. The intra-continental river regime in the past has differed from the regimes of regions with both marine dominated climates and marine/continental transitional climates. The main feature of the intra-continental paleohydrological river regime was extremely high increases in river level. These increases in level were caused by periodic increases of river discharge during the Pleistocene and Holocene which occurred during periods of valley relief instability caused by climate cooling. The simultaneous accumulation of alluvium on surfaces of different heights occurred during high floods and caused the formation of polycyclic terraces. These are the type terraces of the intra-continental regions.

Rhythmic changes of river regime have been identified in the intra-continental regions of Siberia (Fig. 1). The floods which occurred during cold periods followed phases where the river regime was relatively stable. Increases in river level during the Glacial stages were a result of the displacement of cyclones which caused an extreme continental climate, widespread occurrence of permafrost and the formation of ice dams. The simultaneous accumulation of alluvium on terraces of different heights (polycyclic terrace formation) occurred during the extremely high increases in river level.

Geology and geomorphology of the valleys of southern and central Siberia

The differences in morphology of the valleys of southern and central Siberia are a result of neotectonic movements and changes in hydro-climatic processes. The large Siberian rivers originate in the Southern Siberian mountains where the relief comprises high mountain ridges which alternate with inter-montane basins. The river valleys have a beaded structure, whereby the valley structure alternates between broad basins and narrow canyons. Antecedent river valleys with steep slopes and low terrace structures have formed along the mountain sections (Fig. 2). The accumulation of alluvium occurs mainly within the inter-montane basins. The river valleys are broad and have a well developed terrace structure. Island formation is typical for these parts of the rivers (Fig. 3). In Central Siberia rivers flow across the continental shield. Here river valleys also have a beaded structure caused by the differential tectonic movements of ancient structures. The morphology of the river

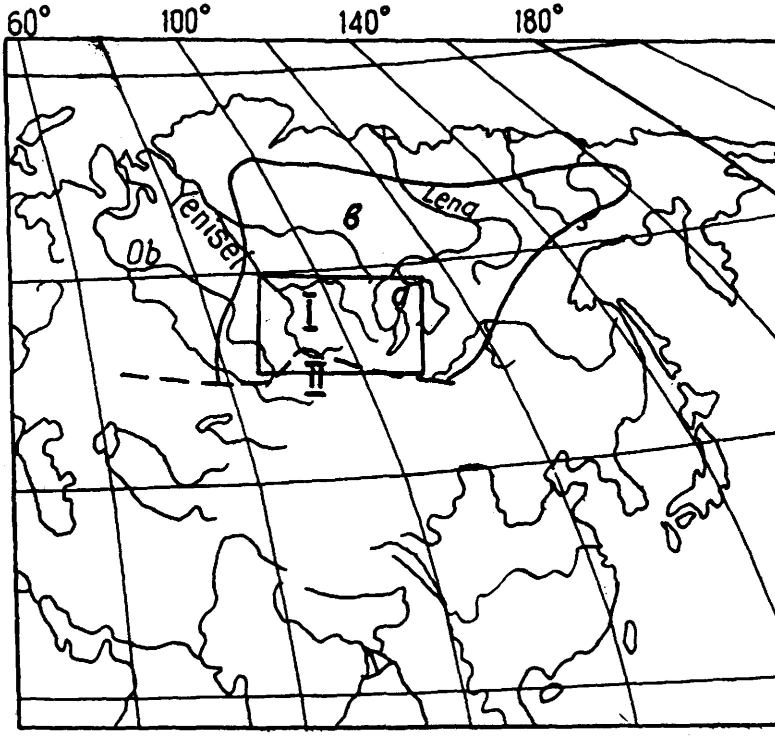

Fig. 1. The location of the study region on the map of Asia (**a**); region of widespread polycyclic terraces and high amplitude of the river level fluctuations in Siberia (**b**).

Fig. 2. Antecedent valley of the river Yenisei in the West Sayan mountains.

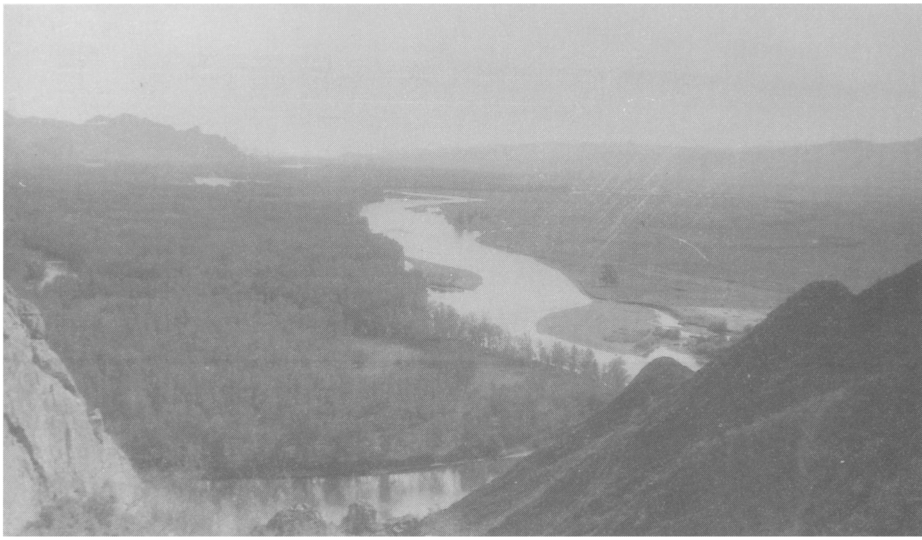

Fig. 3. Broad valley of the river Yenisei with the well developed terrace structure in the Tuva intermontane depression. Climate belts: I, cryosemihumid; II, cryoarid.

valleys was modified by Pleistocene glaciation, and trough valleys were formed in the Southern mountain glacial region. Valleys with low terraces and extensive areas of glacial landforms are located in Northern Siberia where there was widespread glaciation during the Pleistocene.

The river valleys of the former periglacial zone have the most complex terrace structure. Here terrace formation has been influenced by tectonic movements and rhythmic changes of temperature and precipitation which correspond to Glacial/Interglacial cycles.

Within the periglacial zone between 9 and 11 terraces and several levels of floodplain are evident. Three cycles of deep incision have been identified of Early, Middle and Late Pleistocene age. Basal alluvial levels were formed during these incisions and terrace complexes are located on each of the basal alluvial levels. Terraces higher than 70 m belong to the high terrace complex, the middle terrace complex consists of terraces 35–70 m high and terraces lower than 35 m comprise the low terrace complex. Each terrace complex has several levels (Fig. 4). These terrace levels are most distinct within the middle and low terrace complexes.

Sedimentation during the Late Pleistocene provides an example of the formation of polycyclic terraces in the Siberian intra-continental river valleys. This type of terrace formation was caused by the simultaneous accumulation of alluvium on the surface of terraces of different heights. For example, the low terrace complex (Ladeyskaya, Krasnoyarskaya and Berezovsko–Kalinivskaya) and the upper part of the middle terrace strata (Lagernaya and Batojskaya) were formed during the Late Pleistocene. These were formed during extremely high increases in river level (up to 40–50 m) which occurred during Glacial periods.

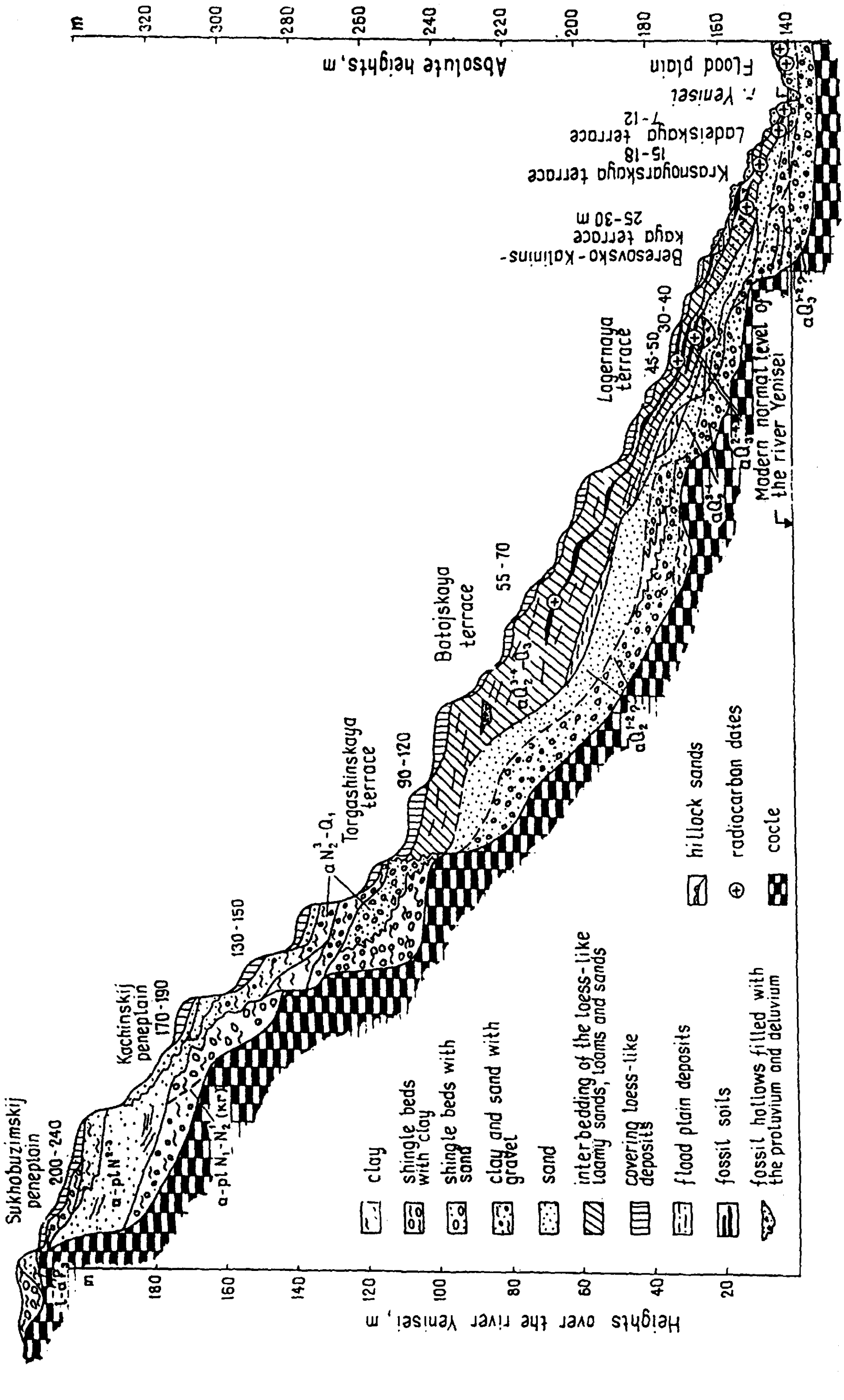

Fig. 4. Structure of the river Yenisei valley in the former periglacial region. Polycyclic terrace formation.

The formation of the low terrace complex on the common alluvial bottom is the main distinctive characteristic of their structure. In the inter-montane basins the erosion surfaces are 10–15 m lower than the modern river level. The climatically-caused incisions in the common alluvial strata occurred after the main erosion-accumulation cycle. Polycyclic alluvial sequences were formed during these incisions.

Fossil soil complexes and cultural horizons within archaeological sites indicate that the accumulation of alluvium was interrupted during Interstadials. Stabilization occurred simultaneously on the terrace surfaces (sections at Oya, Kurtak and Krasnoyarsk) and on the valley slopes (sections at Berezovskoe and Salba) (Fig. 5).

The deposits of the polycyclic terraces were formed during several cycles of sedimentation. For example, the Late Interglacial alluvium comprises the lower part of the Batojskaya and Lagernaya terraces (middle complex) and the Late Glacial periglacial alluvium is in the upper part of the sections. The modern floodplain in

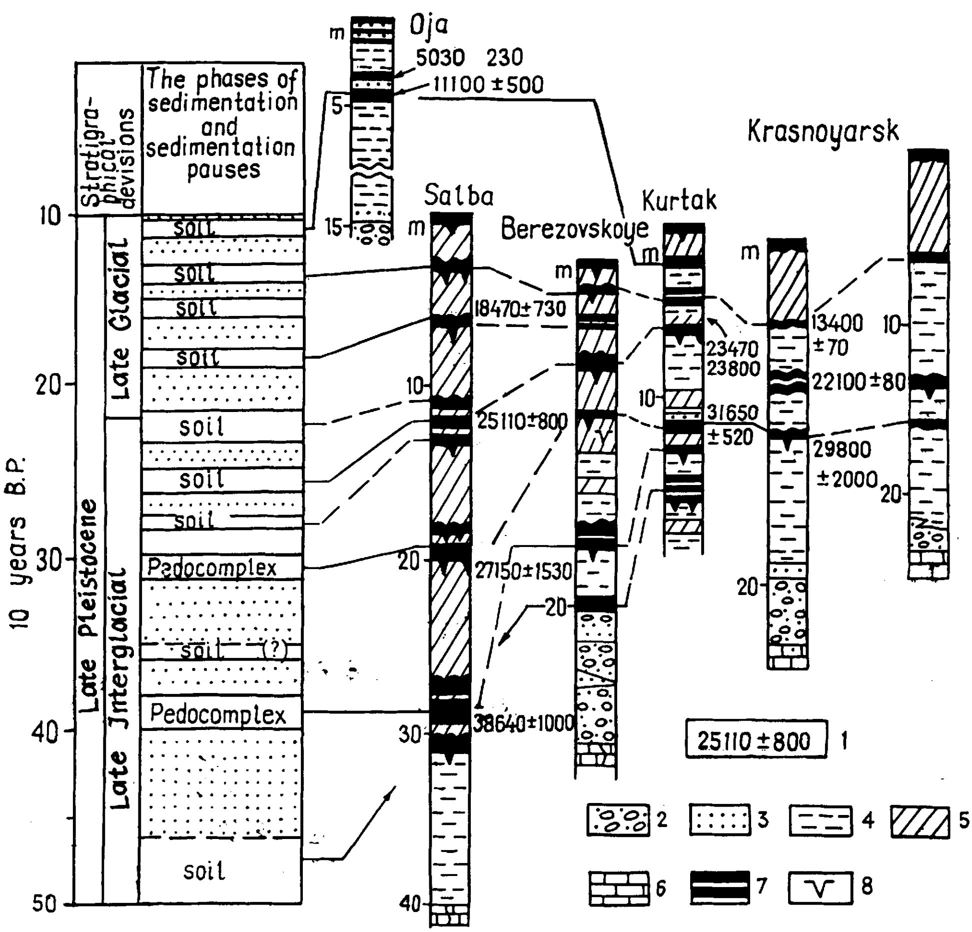

Fig. 5. Correlation of Late Interglacial and Glacial deposits of the Central Siberian river valleys. 1, ^{14}C dates; 2, normal/Channel alluvium; 3, flood-plain alluvium; 4, periglacial alluvium & flood-basin deposits; 5, slope deposits; 6, rock; 7, soil; 8, ice wedges. Arrows indicate phases of sedimentation.

intra-continental Siberia has been formed during the last 5000–6000 years. The longest interruptions in the accumulation of alluvium on the floodplain occurred between 4500–4000 and 500–400 years BP. The formation of some of the alluvial sediments and soils corresponds to the formation of the same sequences in the upper and middle parts of the first terrace strata.

Changes of the intra-continental Siberian paleohyrological river regime

The cyclicity of changes in the climatic, hydrological and erosion/accumulation processes are well expressed in the intra-continental regions of Siberia (Fig. 6). The Late Interstadial/Late Glacial river regimes have been reconstructed on the basis of the accumulation of alluvium on surfaces of different heights. The largest increase in river level occurred during the Late Glacial and was enhanced by the widespread

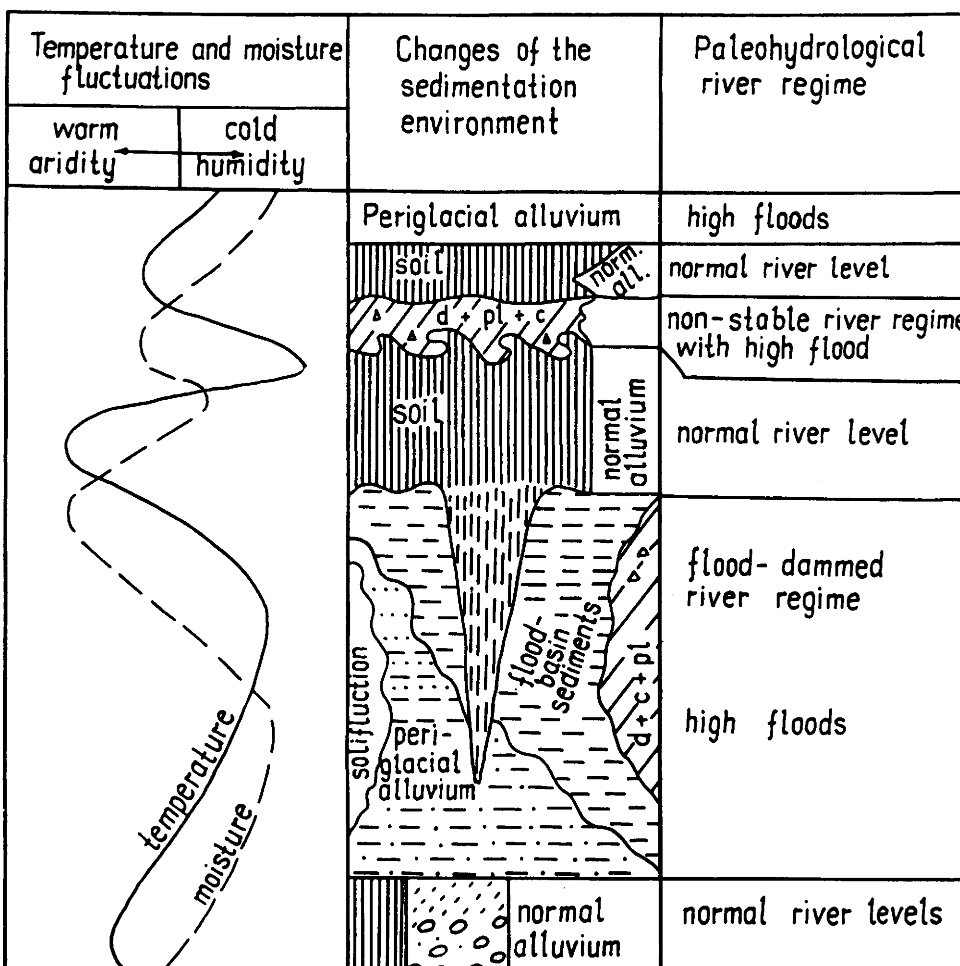

Fig. 6. The complex cyclite formation during one rhythm of the palaeoclimatic fluctuations of the Late Pleistocene.

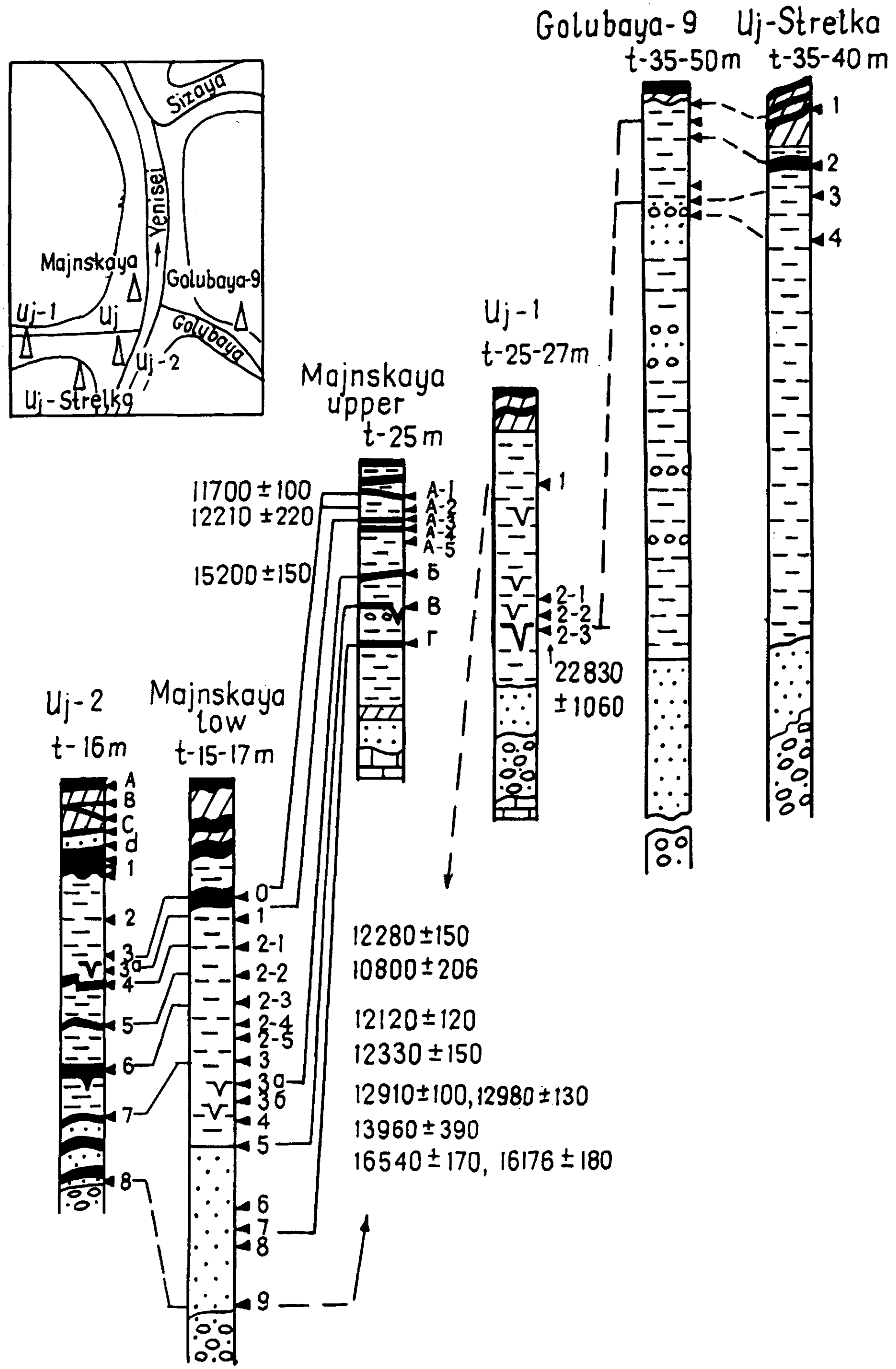

Fig. 7. Cultural layers of the multi-layer archaeological sites within the polycyclic terrace sediments in the river Yenisei valley, Majninskaya group of archaeological sites. Legend as Fig. 5. Solid arrows indicate locations as a number of cultural layers.

occurrence of permafrost and ice dams. Catastrophic increases in river levels occurred during the outflow of water from the dammed lakes.

Interstadial/Stadial climatic fluctuations were one of the factors that caused changes in Pleistocene river runoff. The fluctuations of the meridional temperature gradient caused changes in the zonal atmosphere circulation. These changes resulted in the periodic increase of precipitation in the intra-continental regions (Lamb 1974; Brison & Murray 1977; Budyko 1980). The large amount of melted snow and permafrost in the periglacial zone caused the prevalence of the spring-summer river runoff with the frequently catastrophic rises in river level. These floods continued during the whole warm period.

The most intensive erosion/accumulation processes occurred during the floods. The high energy discharges removed a large proportion of the alluvial, glacial and other types of sediment. The valley bottoms, floodplains and terraces were flooded and on some occasions the direction of these flood streams was different from the direction of the low-water channels.

Palynological data confirm that cyclic sedimentation in the periglacial zone of the Southern Siberian river valleys was caused by cyclic climatic changes. The cyclic changes of the tree pollen content indicate climatic cyclicity; increases in the amount of cold-steppe pollen indicates a cooling climate. Some terrace levels were formed coincident with the 12 cycles of change in spore and pollen content.

Extremely high floods caused the movement of people from one terrace to another during the Palaeolithic and Neolithic periods in southern and central Siberia (Yamskikh 1992). Figure 7 illustrates these movements with reference to the Majninskaya group of archaeological sites which is located in the piedmont of the West Sayan mountains. Human occupation occurred on the low terraces during the periods of valley stabilization (i.e. during Interstadials), whilst during stages of high fluctuations in river level the people moved to the high terrace surfaces.

The highest increases in river level during the Pleistocene occurred before 43.0, 34.0–31.0, 28.0–26.0, 24.0–23.0, 22.0–19.0, 18.0–17.0, 15.5–13.5 and 11.0–10.5 ka BP. Low levels of flooding were typical about 43.0, 35.0 (36.0)–33.0, 29.5–28.0, 25.5, *c.* 22.0, 18.5, 16.5–15.5, 12.5–11.0 and 10.5–10.2 ka BP (Fig. 8).

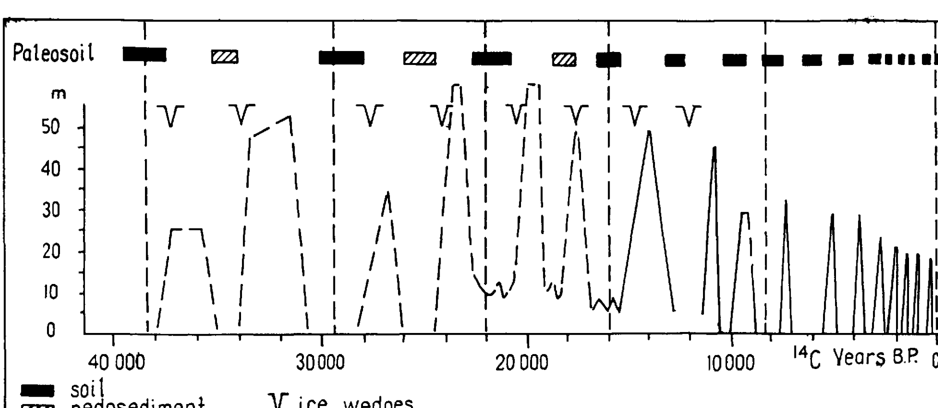

Fig. 8. Changes of the river Yenisei level fluctuations in the periglacial region during the last 40 000 years. The main phases of the levelling of the river flow and climatic warming are shown by vertical broken lines.

Normal, periglacial, floodplain and slope deposits formed during the climate changes during the Interstadial–Stadial transition. These climatic changes corresponded to fluctuations in river level. The dynamics of the erosion–accumulation processes during the Interglacial–Glacial macrorhythm warming–cooling and humidization–aridization have been characterized into five phases.

(1) Warm and humid phase – occurrence of river incision and accumulation of coarse alluvium in bedded and perstrative facies.
(2) Cryohygrotic phase (cold and humid climate) – increase of the amplitude of river level fluctuations, high spring/summer floods and significant river runoff.
(3) Cryoarid phase – characterized by low precipitation, long-term cold season and widespread permafrost. The most intensive accumulation occurred during the floods caused by the breakage of ice dams.
(4) Cryophase phase – the most severe phase of the Interglacial-Glacial macrorhythmn. High amplitudes of river level fluctuations and well-developed ice dammed features were typical of this phase.
(5) Thermal phase – the final phase in the macrorhythm. The rhythmic formation of alluvial strata continued during this time. Temperatures were higher than in the previous phase and the amplitude of fluctuations in river level were lower. In the last macrorhythm this phase corresponds to the accumulation of alluvium on the first terrace during the Holocene. Extremely high floods during the Holocene occurred at 10.2–8.7, 8.0–7.0, 5.2–5.1, 4.0–3.5, 3.3–2.7, 2.4–2.25, 1.7–1.5, 1.3–0.8, *c.* 0.6 and 0.4–0.3 ka BP. Lower river level fluctuations occurred at 8.7–8.0, 7.0–6.0, 5.45–5.2, 5.1–4.75, 4.45–4.3, 3.5–3.3, 2.7–2.4, 2.25–1.76, 1.5–1.3, 0.45 and 0.28–0 ka BP (Fig. 8). This cyclicity caused the formation of the 8–12 m high terrace and the multi-level floodplain.

Fluctuations in river levels changed with the phases of the extremely high floods and were typical within southern, central, northern and northeastern Siberia. Radiocarbon dates confirm the hypothesis that there was simultaneous accumulation of periglacial alluvium on the different height terraces of this region during the Late Interglacial–Late Glacial. Periods of the valley relief stabilization were repeated rhythmically every 7000–8000 and 1700–2000 years. Rhythms which repeated every 900–1000 and 400–500 years were less pronounced.

Fluctuations of the modern river regime are spatially and temporally varied (Afanasiev 1967; Kuzin 1970). This instability has caused different intensities of the accumulation of alluvium on the different height terrace levels and erosion of the floodplain landforms.

Conclusion

The intra-continental location of southern and central Siberia has caused an extra-continental climate in these regions. Irregular river discharges were caused by the occurrence of permafrost and water released from melting snow. These conditions were favourable for the formation of ice dams on the rivers and caused the periodic occurrence of extremely high magnitude floods. This is one of the processes by which polycyclic terraces were formed in the intra-continental regions of Siberia.

Metachronous changes of precipitation and temperature also caused exceptionally high increases in river level. The increase of precipitation usually corresponded with the cold stages (Yamskikh 1993). This was in converse to regions with marine climates and marine–continental transitional climates where increases of precipitation usually corresponded to climate warming (Klimanov 1989). The highest amplitudes of river level fluctuations occurred during the Late Glacial and caused the formation of the greatest number of the different height terrace surfaces.

Terraces became stable on a number of occasions during their formation, when the accumulation of alluvium was interrupted. The simultaneous accumulation of alluvium on the different height surfaces caused the formation of polycyclic terraces. These terraces are typical both within the intra-continental regions and within regions with marine and marine–continental transitional climates.

The regularities of the Pleistocene and Holocene river regime and polycyclic terrace formation are important for a number of reasons. Among these are; assisting palaeoecological reconstructions of the Palaeolithic and Neolithic environment, stratigraphic subdivision, informing the restoration of anthropogenically damaged regions, stratigraphic subdivisions and the prediction and retrodiction of hydrological processes.

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River responses to decadal-scale changes in discharge regime: the Gila River, SE Arizona

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Abstract: Rapid and large responses of semi-arid fluvial systems to alterations in discharge are well documented. Some of these changes have been sudden and have resulted in 'metamorphosis' of the channel. The Gila River in SE Arizona, USA, was identified as behaving in this way in the early twentieth century. Changes of the Gila River channel since 1920 have been investigated, and in particular, the last three relatively wet decades have been compared with the drier ones of the 1920s to 1950s. The research has shown that the morphological response to high flow events depends on sequences of events and critical combinations of conditions. Feedback effects of morphology can induce a ramped response and morphology can be further complicated by interactive effects with vegetation growth. Morphology may therefore not show a simple relationship to size of event or wetness of a period and therefore care should be taken when making palaeohydrological inferences from morphology.

Analysis of changes in river channel morphology in response to known hydrological and environmental conditions is important for retrodiction of discharges, for making palaeohydrological inferences and for prediction of fluvial impacts and land and water management implications. It is becoming clear that climatic fluctuations of sufficient magnitude to produce major changes in fluvial morphology can take place on the timescale of decades. This is a timescale on which it is possible to compile detailed morphological evidence and to obtain long-term discharges, climatic and even land-use records. Analysis of case studies of changes where such data exist therefore provides opportunity to identify and quantify the type and scale of changes in morphology and the relationship to hydrological conditions, including to individual events, sequences of events and longer-term fluctuations.

Retrodiction of discharges in palaeohydrology is based on the premise of equilibrium and quantifiable relations between form and discharge, derived from present-day channels. Since the 1960s and 70s, when such relations were primarily established, their applicability, even to contemporary channels, has been increasingly questioned. Even at the time, there was much debate as to the representative or appropriate discharge parameter to be used. Nevertheless, the approach was used in classic studies such as that of Schumm (1968) in SE Australia. The general notion that the evidence of a set of palaeochannels with very different sizes and shapes from those of the present in an area must indicate a change in regime is accepted. Rotnicki (1983) asserted that retrodiction of discharges for braided streams has not progressed far and this is still arguably the case in spite of much work on processes and forms (Best & Bristow 1993). Rotnicki (1991) has made some progress in application of steady uniform flow formulae using evidence from palaeomeanders. Dury (1985) assessed errors in retrodiction of palaeodischarges from channel dimensions and showed cross-section was better than width which in turn was better than wavelength in relation to meandering channels. Thornes (1987) was of the opinion that development of non-linear models and of integrated climate–biosphere–hydrosphere interactions are of major significance in palaeohydrology but arguably these developments have not been absorbed into palaeohydrological practice as fully as they might. This paper exemplifies the variability and complexity of relations which can arise even on a timescale of decades.

The high variability of channel morphology in semi-arid fluvial systems is well known (e.g. Thornes 1981). It has been recognized that ramped responses can occur because of the occurrence of high-magnitude events without intervening, smaller events which tend to accomplish reconstruction of channels (Wolman & Gerson 1978). Sudden transformations of channel morphology have also occurred, in a process termed 'channel metamorphosis', and have been identified in various semi-arid and sub-humid regions, such as the western USA (Schumm & Lichty 1963; Nadler & Schumm 1981) and Australia (Brizga & Finlayson 1990). One of the most cited examples of such a change is that of the Gila River in SE Arizona, documented in detail by Burkham (1972). That study was part of a major US Geological Survey project on the hydrological effects of phreatophytes (Culler 1970).

Burkham (1972) showed that the Gila River in Safford Valley, Arizona, had a narrow, stable, meandering channel in the late nineteenth/early twentieth century. In the period 1905–1920 the channel was transformed to become very wide, unstable and braided, occupying much of the valley floor. Gradually during the 1930s, 40s and 50s the floodplain was reconstructed and the channel became narrower and more stable once more. Burkham (1972) suggested that the destabilization early this century took place as a result of a massive flood in 1905 and that this pushed the channel system across a morphological threshold. The middle part of this century was a period of few high flows and a lack of heavy winter rains. A wetter period, including several major floods, has ensued since the completion of Burkham's study and this provides opportunity to examine whether conditions for destabilisation have been repeated and what the response of the system has been to these succeeding hydrological variations.

Physical setting of the Gila River

The Gila River flows westwards from the mountains of southern New Mexico, across southern Arizona (Fig. 1). Safford Valley is a wider portion between mountain ranges in this Basin and Range physiographic province. Where the river flows out of the canyon at the Head of Safford Valley it has a catchment area of 20 451 km^2. The valley floor broadens rapidly and attains its maximum width of approximately 6 km near the town of Safford. Safford Valley, the study section, is about 90 km in length and extends to the San Carlos reservoir, impounded by the Coolidge dam in 1928.

The valley has a mean annual rainfall of *c.* 313 mm and naturally supports desert scrub vegetation at lower elevations. The flow of the river is perennial in most years, fed by precipitation in the mountains of New Mexico. Water from the river and from groundwater in the underlying alluvial gravels is used for irrigation of cotton, grown

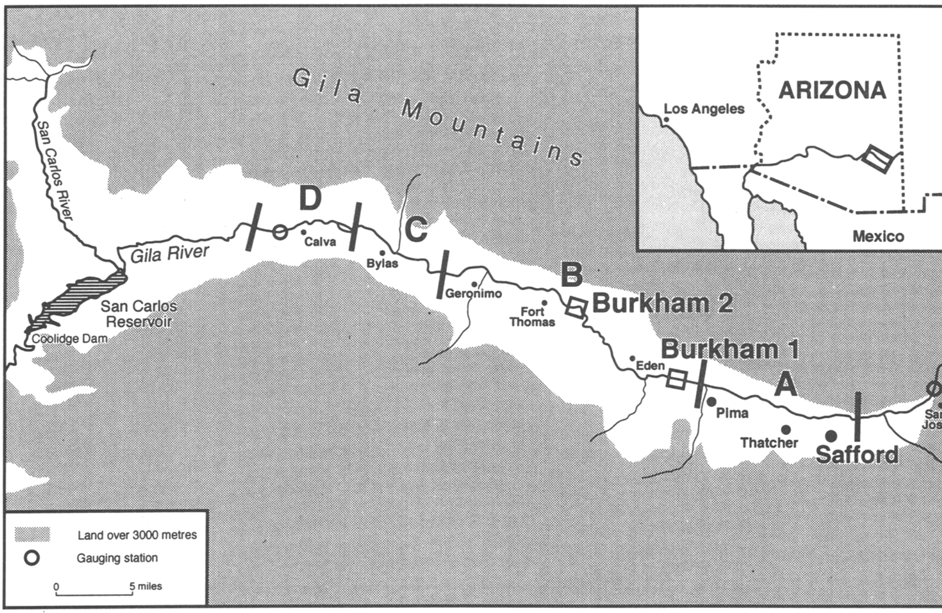

Fig. 1. Map of Safford Valley with sections and study sites.

intensively over much of the valley floor. The lower end of Safford Valley falls within the San Carlos Apache Indian reservation.

Discharge has been gauged continuously at the Head of Safford Valley since 1920 (peak flows since 1914) and at the lower end, at Calva, since 1936. Of the 20 highest annual peak flows since 1914, nine have occurred since 1972 and none of the 10 highest peak flows occurred in the period 1917–1965 (Fig. 2, Tables 1 & 2). The

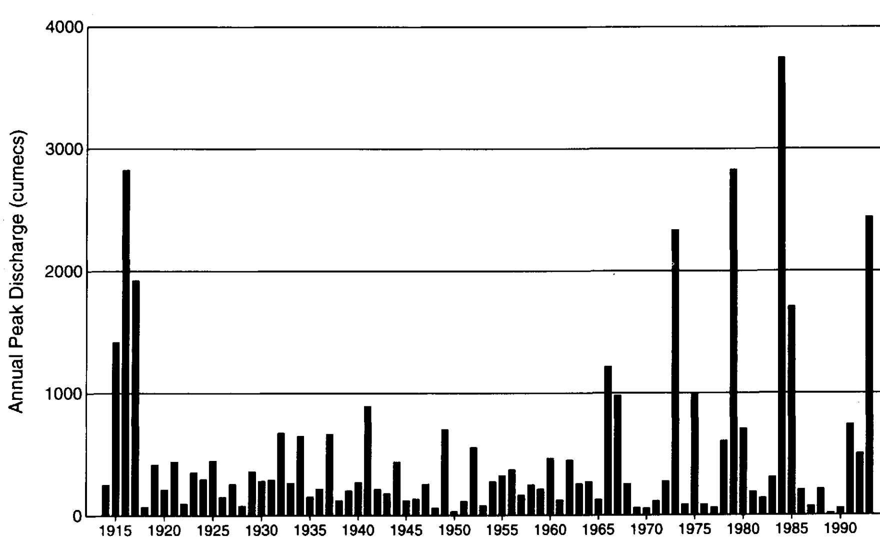

Fig. 2. Annual instantaneous peak flows of Gila River at Head of Safford Valley, gauge No. 4485, 1914–93.

Table 1. *Annual discharge data ($m^3 s^{-1}$), River Gila Station 4485, Head of Stafford Valley, 1914–1993*

Date (water year}	Mean daily	Annual max daily	Annual peak	Date (water year)	Mean daily	Annual max daily	Annual peak
1914			254.88	1954	6.95	214.67	278.95
1915			1416.00	1955	6.28	89.77	331.34
1916			2832.00	1956	2.88	132.54	376.66
1917			1922.93	1957	8.31	213.25	169.35
1918			76.46	1958	17.61	216.93	256.58
1919			424.80	1959	6.89	95.72	222.60
1920			215.80	1960	12.47	328.51	472.94
1921	11.78	270.46	444.62	1961	3.54	42.76	135.94
1922	5.32	48.99	107.05	1962	17.39	233.64	455.95
1923	12.92	177.57	356.83	1963	11.20	100.25	264.79
1924	15.30	254.03	300.19	1964	6.37	99.12	279.80
1925	8.30	322.85	450.29	1965	7.19	54.09	135.94
1926	13.14	124.89	160.29	1966	25.16	872.26	1217.76
1927	11.55	129.71	263.94	1967	8.54	489.94	985.54
1928	6.76	39.65	91.47	1968	27.77	176.72	262.81
1929	9.26	136.79	359.66	1969	4.77	28.89	69.67
1930	9.80	156.04	286.03	1970	4.33	28.60	63.72
1931	13.04	227.41	297.36	1971	3.81	42.76	127.72
1932	19.76	328.51	679.68	1972	9.83	210.70	288.86
1933	9.74	121.49	271.87	1973	40.01	1277.23	2333.57
1934	5.66	359.66	651.36	1974	3.61	44.18	92.89
1935	8.92	82.13	157.18	1975	13.65	421.97	991.20
1936	8.52	139.62	226.56	1976	6.34	70.23	96.29
1937	16.52	373.82	671.18	1977	5.02	53.52	71.93
1938	6.69	75.05	132.82	1978	12.95	492.77	611.71
1939	5.80	66.27	208.72	1979	47.66	1775.66	2832.00
1940	9.01	148.68	278.67	1980	17.44	484.27	716.50
1941	34.73	623.04	903.41	1981	5.60	52.68	198.24
1942	13.33	238.17	218.91	1982	8.55	57.49	148.40
1943	6.00	88.64	189.18	1983	30.32	368.16	320.02
1944	5.40	76.75	447.46	1984	28.22	2548.80	3738.24
1945	8.85	71.65	136.50	1985	35.87	985.54	1704.86
1946	4.58	72.22	144.43	1986	16.16	184.93	217.95
1947	4.44	66.27	261.96	1987	15.17	60.04	85.53
1948	4.96	22.83	71.93	1988	16.72	171.62	221.46
1949	22.73	472.94	713.66	1989	6.29	19.60	25.23
1950	3.74	19.65	35.12	1990	4.85	40.21	62.73
1951	2.86	41.35	120.08	1991	24.67	526.75	741.98
1952	12.89	402.14	557.90	1992	31.63	373.82	506.93
1953	3.54	32.85	86.09	1993	–	270.73	2441.18

highest recorded peak, in October 1983, reached an instantaneous peak of $3738\,m^3s^{-1}$ and is calculated to be very nearly as high as the 1905 flood attributed with destabilising the channel. Particular attention is therefore paid to the effects of this 1983 flood. Records of mean and total flows also exhibit marked fluctuations this century, with the period of lower flows in the middle part of the century producing significant cumulative deficits from the average. Analysis of the discharge variations in relation to both rainfall and water usage indicates a close relationship

Table 2. *Averages of annual discharges of the Gila River*

		Gauge No. 4485		
Start year	End year	Annual mean daily ($m^3 s^{-1}$)	Annual maximum daily ($m^3 s^{-1}$)	Annual total flow Coolidge Dam site ($m^3 \times 10^6$)
1874	1891			729.75
1892	1904			174.52
1905	1920			811.76
1921	1945	11.04	187	306.68
1946	1965	8.35	148	187.36
1966	1992	16.85	439	513.27

between mean annual discharge and winter precipitation (Hooke 1994) and that most of the high flows are generated by winter storms of various types (Hooke 1994; Burkham 1970; Hirschboek 1991; Webb & Betancourt 1992). The climatic influence far overwhelms that of human activities on discharge and the incidence of storms, floods and wet periods in this region has now been related to major climatic oscillations and El Niño events (Webb & Betancourt 1992).

Methods of morphological analysis

Burkham (1972) provided quantitative data on channel widths at various dates, averaged for four reaches within Safford Valley (Fig. 1), and more detailed evidence in the form of historic maps and aerial photographs for two sections (marked Burkham 1 and Burkham 2 on Fig. 1). In this study, a baseline mapping has been obtained from colour aerial photographs at 1:7000 scale taken in 1982, which therefore allows assessment of changes resulting from the floods of the 1960s and 70s and documentation of the situation prior to the 1983 flood. The post-1983 flood channel was mapped from USGS Reports and Soil Conservation Service aerial

Table 3. *Classification of areas used in mapping Gila Valley floor*

Zone	
1	(Dark brown and water) active, moist low-flow channel
1a	(Dark green) seepage zones; wet, with low plants
2	(Light brown/yellow) fresh bars and unoccupied (dry), active channels
3	(Grey) slightly older, active bars with bare sediment, scattered bushes
4	Scrubland, young phreatophytes, scattered bushes, old channels visible
5	Old abandoned channels
6	Dense phreatophytes
7	Structures, dykes, levées, etc.
8	Workings, gravel pits, anthropogenic features
9	Abandoned, rough fields
10	Cultivation
11	Upland

Fig. 3. (**a**). Aerial photograph of Gila River channel and floodplain in Safford Valley, near Pima, 1982. (**b**). Ground photograph of Gila River channel and floodplain in Safford Valley, near Pima, 1992.

surveys. Sample areas were resurveyed by field mapping in 1992 and a cross-section profile published by Burkham (1972) was remeasured. Additional information on flood impacts was gleaned from post-flood appraisals published by the US Army Corps of Engineers and USGS after major floods (Aldridge & Hales 1984; Aldridge & Eychaner 1984; US Army Corps of Engineers 1973, 1987).

The channel and floodplain throughout the length of Safford Valley were mapped from the 1982 aerial photographs by categorising areas into 11 types classified according to morphology, sediment and vegetation cover (Table 3); these were then combined to identify three major zones in the floodplain: the low flow channels, the active zone of gravel and sand bars, and the phreatophyte zone, occupied mainly by dense tamarisk bushes. The nature of the channel and floodplain can be seen from the examples of an aerial photograph (Fig. 3a) and ground photograph (Fig. 3b).

Morphological changes

Average widths for Burkham's (1972) four reaches have been calculated from measurements of sample cross-sections every kilometre on the 1982 maps. Active width is that zone reworked by the channel and appears to be comparable with widths used by Burkham. The 1982 figures have been added to Burkham's (1972) original graph in Fig. 4. This shows how width increased in the period 1905–20 but then gradually declined. It can be seen that the floods of the 1960s and 70s widened the channel again such that, by 1982, it was wider than in the 1940s and 50s but not comparable with the metamorphosed channel of 1905–20.

It has not been possible to derive accurate measurements for the whole length of the valley for the post-1983 period but more detailed evidence of responses can be examined at the section scale. Burkham 2 is a section in which the channel has a dominantly meandering pattern, abutting the highland of the valley sides in places and bordered by dense, mature tamarisk growth on both margins of the floodplain. The sequence of pattern changes derived from Burkham (1972), the 1982 survey, 1983 flood aerial photographs, the post-1983 aerial survey (Garrett *et al.* 1986) and 1992 field mapping is shown in Fig. 5.

Again, it is evident that major changes in channel position and pattern took place prior to 1983, presumably in response to high flows in 1972, 1974 and 1979 (water

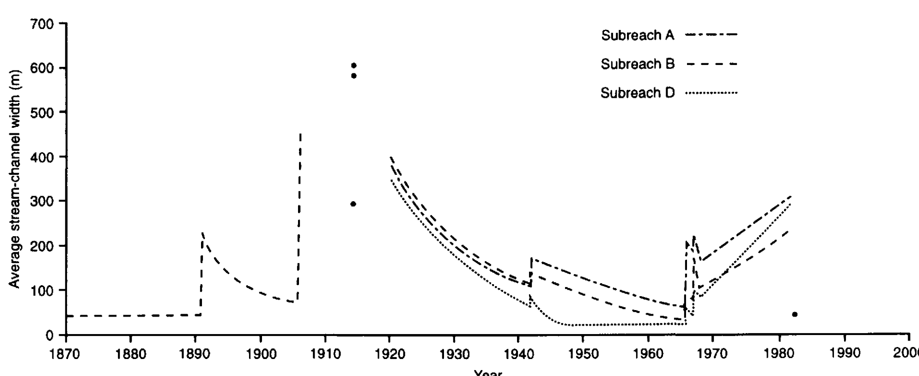

Fig. 4. Mean widths of channel in reaches of Gila River, Safford Valley (after Burkham 1972).

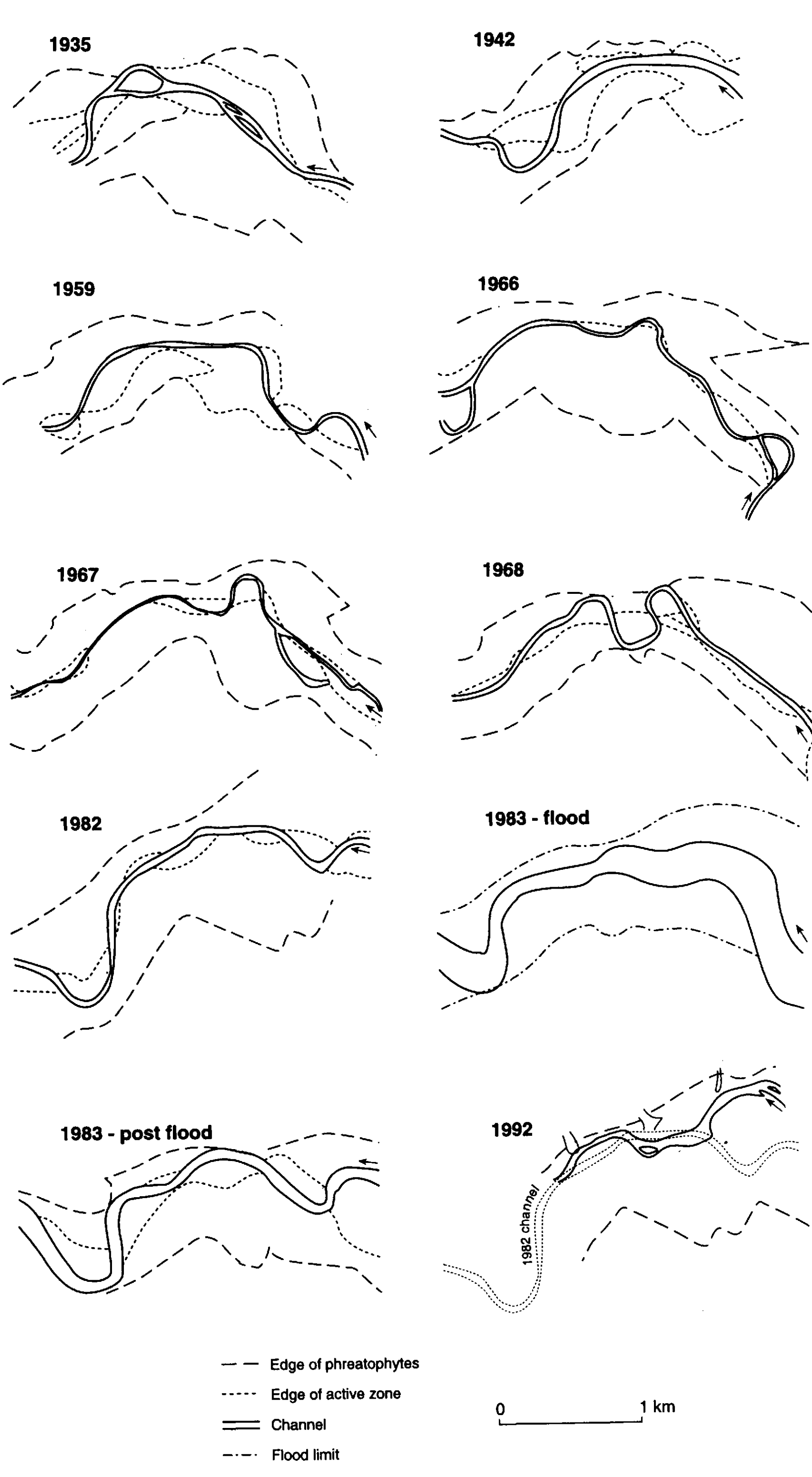

Fig. 5. Variations in width and pattern over time within section Burkham 2 near Fort Thomas, Safford Valley.

years). Indeed, it can be seen that the pattern changed between 1966, 1967 and 1968 a period in which two annual peak floods of around $1000\,m^3\,s^{-1}$ occurred. Similarly, large changes in position and pattern took place between 1983 and 1992, a period in which three annual instantaneous peaks exceeded $500\,m^3\,s^{-1}$.

The change in morphology of the channel through this century is confirmed by repeat surveys of a cross-profile within section Burkham 2. In 1915 and again in 1935 the profile was very even, without a distinctive channel within the floodplain. By 1943 a more distinctive channel had formed, bordered by aggraded floodplain at a higher level. Elevation of the floodplain increased in the period to 1967 and the depth of the channel increased. In 1992 the channel was slightly more incised and some material had been removed from within the active zones. The outer, phreatophyte zones were at comparable height to the situation in the 1960s. Since the channel is more defined and the bed several metres below the floodplain surface now, this means that, in a flood, the flow is more confined, the velocities are higher and thus there is a positive feedback effect of morphology.

The nature of changes differs in detail in other parts of the valley, depending mainly on the type of pattern and degree of confinement but the broad trends are similar. Several conclusions can be drawn from this evidence. Although the 1983 flood was nearly comparable in magnitude to that of 1905, the valley floor was not transformed in the same way. Other, smaller floods, prior to 1983 had produced significant morphological changes and changes appear generally to be roughly proportional to size of flood, though with the impacts of the 1983 flood being rather less than expected. That lesser floods can produce changes is confirmed by observations immediately after a $507\,m^3\,s^{-1}$ event in 1992 in which large amounts of lateral shifting, bank erosion and degradation and aggradation of bars took place. Since then, an even larger event, fifth in ranking of the 80 annual instantaneous peaks and $2442\,m^3\,s^{-1}$ in size, has occurred; preliminary field observation and mapping evidence suggests that it has caused large shifts in the channel and massive aggradation on some bars. Again it has not 'metamorphosed' the channel but the large impact of this latest event may be related to ramped effects of the succession of floods or to the season of occurrence, with greater effects in winter than in the autumn.

It would, therefore, appear that the simple threshold model of Burkham is not applicable to the changes on the Gila in the latter part of the twentieth century. If the threshold model is correct then either the position of the threshold, i.e. the magnitude of the flood, needed for metamorphosis has increased or it is not a simple, single flood magnitude threshold. Supporting evidence that particular combinations or sequences of conditions are critical for 'metamorphosis' comes from other work in the SW United States, such as that of Balling & Wells (1990). The 1905 flood occurred after the severest drought for hundreds of years and it is suggested that such a sequence of severe drought and very high magnitude floods had not occurred in 1000 years (Jacoby 1987). The effects of drought on vegetation and therefore resistance to erosion are particularly important. The greater definition and confinement of the present channel confirms the idea that it would need a higher threshold for the channel morphology to be metamorphosed to that of 1905–20.

The last three decades have been a much wetter period than the preceding three decades but not an exceptional period in the long-term record. This recent period is not quite as wet in terms of total rainfall and total or mean flows as that of 1905–20 but comparable in terms of flood incidence (Hooke 1994). However, the channel morphology in the recent period is different from that of 1905–20. If the morphological evidence alone was used, as so often has to be the case for past periods, and conventional palaeohydrological analysis applied then it would be concluded that the channel forms of 1905–20 and 1965–94 originate from very different sorts of periods, whereas, in the overall pattern of fluctuations, both are relatively wet.

Very rarely has the combination of events or patterns of sequences of events been examined or tested in detail, though the importance of relative magnitude and frequency, has of course, been recognized (Wolman & Gerson 1978). Not only should averages for periods or magnitudes of events be analysed and compared but the sequence and order of events. This is difficult to do statistically but the flow data for the Gila were partitioned into periods relating to the major fluctuations and runs tests applied. These produced no significant patterns, nor did cross-correlation between comparable periods. Some progress may be made by modelling outcomes of combinations of conditions, given adequate understanding of the component processes.

The conclusions to be drawn from this case study and the model which may be constructed are further complicated by the other environmental changes which have taken place this century, and the question of whether the situation has been changed so that the channel system is not responding in the same way as 80–90 years ago. Major changes which have taken place include the influx of tamarisk and modifications of the sediment load due to land-use changes.

The extent of cause and effect in the interaction with tamarisk is difficult to determine. Tamarisk grows quickly and has a dense aerial structure as well as a long tap root. It easily colonizes moist bars but large floods can inhibit spread of the plant (Hereford 1984). It is not easily washed out and can survive some flooding once established. Sedimentation takes place around tamarisk bushes (Burkham 1976) as evidenced by the ridges and sediment wakes commonly associated with the plants. Graf (1978, 1981) suggests that the presence of tamarisk causes aggradation, that the channel becomes poised near the surface of the floodplain and that major channel changes then take place by avulsion. Thus, zones of dense tamarisk growth are zones of high channel instability. More usually, in fluvial systems, vegetation is regarded as having a stabilizing effect, increasing resistance of bars. Because the channel of the Gila is now relatively incised, the avulsion process is not a common mechanism of channel change, particularly in the tamarisk zone. Much more important now is lateral erosion and switching within the active zone due to deposition of coarse gravel bars. Lateral erosion of the mature tamarisk does take place, though the extent to which resistance is increased and rates decreased compared with unvegetated zones has not been quantified. Further work is needed on the interactions of vegetation and channel change in floodplains but the evidence here indicates important feedback effects between vegetation and morphology that further complicate effects of climate upon river systems.

Discussion and implications

The temporal variation of channel and floodplain morphology and the complexity of relations to discharge events has been demonstrated in relation to the Gila River. This example is from what is usually regarded as a sensitive environment. However, given that major variations in climate on a timescale of decades are increasingly being identified from different parts of the world and the periodicity and interconnectedness of climatic and oceanic teleconnections is increasingly being recognised then the results may have wider implications, even for more humid areas.

In Australia major variations in discharge regime, termed flood-dominated and drought-dominated regimes, of about 40 years duration have been identified (Pickup 1976; Riley 1981; Erskine & Bell 1982; Warner 1987). These have been related to climate, with the magnitude of fluctuations again quite overwhelming effects of human activities such as impoundment, even on the eastern seaboard of New South Wales. The alteration from one discharge regime to another has been found to produce major changes in channel morphology; for example sinuosity on the Hunter River was reduced from 3.86 to 2.66 between 1870 and 1899 coinciding with a flood-dominated period and increases in mean annual rainfall of 30%. Further decreases in sinuosity took place in a second flood-dominated period, 1949–1978.

Major variations on the timescale of decades are not as well established either for discharge or climate in Britain, though evidence of some periodicity of activity is beginning to emerge (Rumsby & Macklin 1994; Ibsen 1994). However, even in the shorter term and in this less sensitive, humid climate some effects are observed and complex interactions may be operating. Bank erosion on active meandering streams commonly occurs in moderate to high events; rates of erosion and of lateral movement tend to be closely related to incidence of such events. Finer deposition will tend to take place in smaller events and thus bar and floodplain construction may take place in a relatively low flow period. Once vegetation colonises bars, whether due to the fine deposition or additional effects of temperature and moisture variations, then, as in more marginal climates for growth, it will take larger events to destabilize those bars. Few channel systems in Britain may be near the threshold for 'metamorphosis' of morphology but climate variations on the timescale of decades may be sufficient to induce different levels of activity and modify parameters of morphology such as width. Some such impacts are being detected from monitoring of an active river over a period of a decade or more (Hooke 1990).

The implications of these findings for palaeohydrology are important. The sensitivity of systems, and reaches within systems, to changes needs to be quantified in more detail and the thresholds within systems need to be identified. However, the complexity of threshold conditions needs to be recognised and the fact that interactive feedback effects occur means that responses to discharge events are not simple. The impact of an event or even a series of events may depend on the position within longer-term fluctuations. The example of the Gila River shows that morphology may be markedly different even in periods which, on a long-term perspective and with lack of good hydrological records (as for most of the past), would appear to be comparable. Great care must therefore be taken in making simple palaeohydrological inferences from morphological evidence.

There is a need for much more sophisticated modelling of sequences of events and trends in conditions, with the feedback effects of morphology incorporated; proposed future research will develop this. One of the problems is testing such models and collecting adequate data on outcomes of different combinations and sequences of events. Such knowledge is vital for satisfactory understanding of fluvial responses and for retrodiction and prediction of channel changes. To accomplish this, more data on past channel changes for which discharge data are also available need to be collated. However, there is still a problem of comparability of different sites, even within the same type of environment. Extension even farther backwards at particular sites is usually limited by poor discharge and morphological data. Thus analysis of historical data from a large number of sites should be complemented by adequate monitoring of sites where good data do already exist .

The other implications are for predictions of discharge and channel variations and for land and water management in the future. If climatic fluctuations of the magnitude identified here do commonly occur on a timescale of a few decades then the characteristics of each period and the differences in extremes need to be considered in land and water resource planning. If water resource planning is based on data derived from a wetter period without it being recognized as such, then this could cause severe problems through a sustained drier period in the future. Indeed, in the Gila Valley the impoundment of the San Carlos reservoir was planned during the wetter period of the 1920s and was then succeeded by the dry period of the 30s, 40s and 50s in which the reservoir never filled. In recent years it has been brimful and attention is now being paid to the effects of floods on the land. During the low flow period, farmers began to encroach on the floodplain and cultivate the margins of the Gila 'active' zone. The floods in the period since the 1960s have caused much damage but perhaps it now has to be accepted that this recent higher frequency of floods is just as much a 'normal' period as that of the middle part of the century.

Effects of global warming may cause even greater problems because they are likely to be superimposed on the kind of fluctuations identified here. Most of the predictions under global warming scenarios are for mid-latitude semi-arid regions to become warmer and drier with an increasing threat of 'desertification'. This may, therefore, mean that the conditions in the drier phases of the fluctuations already identified become predominant but presumably the same kind of period and scale of fluctuations will be superimposed on this and thus the dry extremes will be greater. (It is also possible from the scenarios that period of variability will shorten.) It has already been shown that interaction of flows with vegetation cover and availability of moisture to sustain vegetation appears to be a crucial factor influencing morphological impacts and thus the seasonality and rainfall regime will influence the impact of global warming. Again modelling of impacts under these climatic scenarios will provide some indication of likely changes. However, given the present morphology of the Gila river, it will take a great deal of aggradation and therefore increased supply of sediment to produce the morphological conditions comparable with those under which the previous metamorphosis took place. This might arise if vegetation cover is reduced by increased aridity but equally some large flow events are needed to produce the channel changes otherwise the stabilization that took place in the middle part of the century is likely to occur. If, on the other hand conditions produce a greater incidence of large flows without additional sediment influx then the present active zone is likely to continue to widen so that a less confined channel may once more be created but at a lower level. However, it would appear that the actual combination of conditions that gave rise to the metamorphosis of 1905 was so rare that it will still have a low probability of occurrence. Of course, spatial feedbacks are likely to operate such that increased widening will supply sediment downstream. Such complex response has still to be identified in much of the data on historical impacts. What does seem likely is that at least the same scale of fluctuations of flow and of morphology should be allowed for under conditions of global warming as have occurred in the recent historical past and effects could be enhanced.

Conclusions

Marked variation in the incidence of high flows, related to climatic fluctuations, can take place on a timescale of decades. Stream channels in semi-arid areas are particularly sensitive to such fluctuations but the morphological response depends on sequences of events and critical combinations of conditions. Feedback effects of morphology can induce ramped responses. Morphology can be further complicated by interactive effects with vegetation growth whether induced by climatic influences or direct or indirect effects of human activity. Morphology may not show a simple relationship to size of event or wetness of a period and therefore care should be taken in making palaeohydrological inferences from morphology.

The magnitude and timescale of variations in climate, discharge and channel and floodplain morphology should also be taken into account in future planning of land and water resources, not only in semi-arid zones but also more humid areas. Data on various combinations and sequences of events are needed in order to be able to model channel responses adequately so detailed documentation on the timescale of decades from a range of environments is essential for progress in this field.

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Climatic or anthropogenic alluviation in Central European valleys during the Holocene?

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Abstract: Traces of increased river activity during the Late Glacial and Holocene can be identified in the river valleys of central Europe. This activity includes changes of river channel pattern and changes in the type and rate of floodplain sedimentation. A model of the sequence of events within a phase of increased river activity has been constructed as part of a detailed study of the Vistula valley. This analysis indicated that the evolution of the Vistula river valley has resulted primarily from climatic factors, although local factors (including human modification) contributed to the magnitude of changes which had already developed as a result of climatic controls.

Traces of increased river activity during the Late Glacial and Holocene have been identified in the river valleys of central Europe. This activity includes changes of river channel pattern (palaeomeander cutoffs, avulsions), changes of type and rate of sedimentation on the floodplain (peat growth, covering of peats by muds, buried soil) and accumulation of a large number of tree trunks within the alluvia. These processes may have been caused by changes in flood frequency and/or magnitude. Evidence shows that these phases of increased activity occurred within specific certain periods of time (Becker 1982; Schirmer 1983; Starkel 1983; Kalicki 1988, 1991*a*); Younger Dryas, 9800–9300, 8800–8000, 6600–6000, 5500–4800, 4500–4000, 3200–3000, 2700–2600, 2200–1800 BP, fifth–sixth, ninth–eleventh, fourteenth–eighteenth centuries (Kalicki 1991*a*; Krąpiec 1992). These periods are also in agreement with periods of increased climate cooling and humidity (Starkel 1985), when there was glacial advance, landsliding, lake level rise, accumulation of travertines and vegetation changes (Patzelt 1977; Bortenschlager 1982; Ralska-Jasiewiczowa & Starkel 1988; Kalicki 1991*a*; Magny 1993).

With the exception of some sources (e.g. Buch 1988; Buch & Heine 1988), most authors have related increased fluvial activity to climatic changes during the Holocene. These phases were identified on rivers originating in the mountains and were attributed to the increase of summer precipitation (Starkel 1983). Other studies (e.g. Florek 1982; Turkowska 1988; Kamiński, 1993) showed that phases of increased fluvial activity occurred on rivers of various sizes and regime type (Kalicki 1991*b*, 1991*c*). Certain authors maintain that rivers were active only during these phases and in the periods between fluvial accumulation did not occur (Brunnacker *et al.* 1976; Brunnacker 1978; Schirmer 1983; Schellmann 1990). It is probably more appropriate to state, however, that fluvial processes occurred continuously throughout the Late Glacial and Holocene, but were increased during flood phases (Starkel 1983; Kalicki 1991*a*).

Model of a flood phase

Detailed studies of the Vistula valley downstream of Cracow (Starkel & Kalicki 1984; Kalicki 1991*a, b*, 1992*a, b*; Starkel *et al.* 1991; Kalicki *et al.* 1996) have enabled a model of the sequence of events within one phase of increased river activity to be constructed. The following four stages have been distinguished in the wide valley.

(1) Accumulation of silty-sandy muds along the active channel with simultaneous claying of peats on the far banks of the valley and accumulation of clayey muds in depressions, e.g. palaeochannels (resulting from flooding of the whole valley bottom).

(2) Channel straightening caused by meander cut-offs, bank erosion and lateral migration. Bank erosion and tree trunk deposition in the sediments, washing out the older series and removal of the older tree trunks ('black oaks') led to accumulation and blocking of the channel which promoted the formation of meander cut-offs and avulsion. There is a considerable difference in the processes of this stage between the early and mid Holocene and the Subatlantic period. In the early Holocene, single tree trunks or separate generations can be found which indicate channel stability and slow lateral migration (Kalicki & Krąpiec 1996; Kalicki *et al.* 1996). During the Subatlantic the number of tree trunks within the alluvia rapidly increases (Becker 1982; Krąpiec 1992). This indicates an increase in the lateral migration of channels. Changes of bank erosion have been reconstructed using dendrochronological techniques (Kalicki 1991*a*; Kalicki & Krąpiec in press).

(3) Channel avulsion (although this did not always occur).

(4) Accumulation of clayey muds towards the end of the phase in a different part of the valley to where the channel relocated, beginning with the accumulation of peats in abandoned palaeochannels.

The change in the rate of accretion of overbank deposits, causing the buried soils to develop, occurred in the narrow valleys with stable channels in the periods of increased river activity (e.g. Glasko & Folomeev 1981; Kalicki & Sanko 1992; Pożaryski & Kalicki 1995).

Climate and local factors

In previous studies there has been a considerable scattering of data (often well documented but consisting of only single profiles in various valleys), and the use of diverse research methodologies has made it difficult to interpret the factors that indicate an increase in fluvial activity. It has not been possible to determine the influence of individual components or processes responsible for particular features. That is why, for example, the covering of peats by muds has been interpreted as evidence of high flood levels (e.g Starkel 1983; Ralska-Jasiewiczowa 1988). Detailed research near Cracow has since shown that in wide valleys changes in floodplain sedimentation have been caused by the avulsion of the river channel (Kalicki 1991*a*, 1992*b*).

A large data set has been collected during detailed research of the Vistula valley near Cracow which includes *c.* 100 radiocarbon and more than 500 dendrochronological datings and *c.* 1000 grain-size analyses. From these data it has been possible to distinguish new phases of increased river activity, e.g. 3200–3000 BP (Kalicki 1991*a*), which can also be found in other river valleys of Central Europe (Kalicki 1991*a, b, c*; Kalicki & Krąpiec 1991*a*). Additionally, knowledge of the geomorphological characteristics,

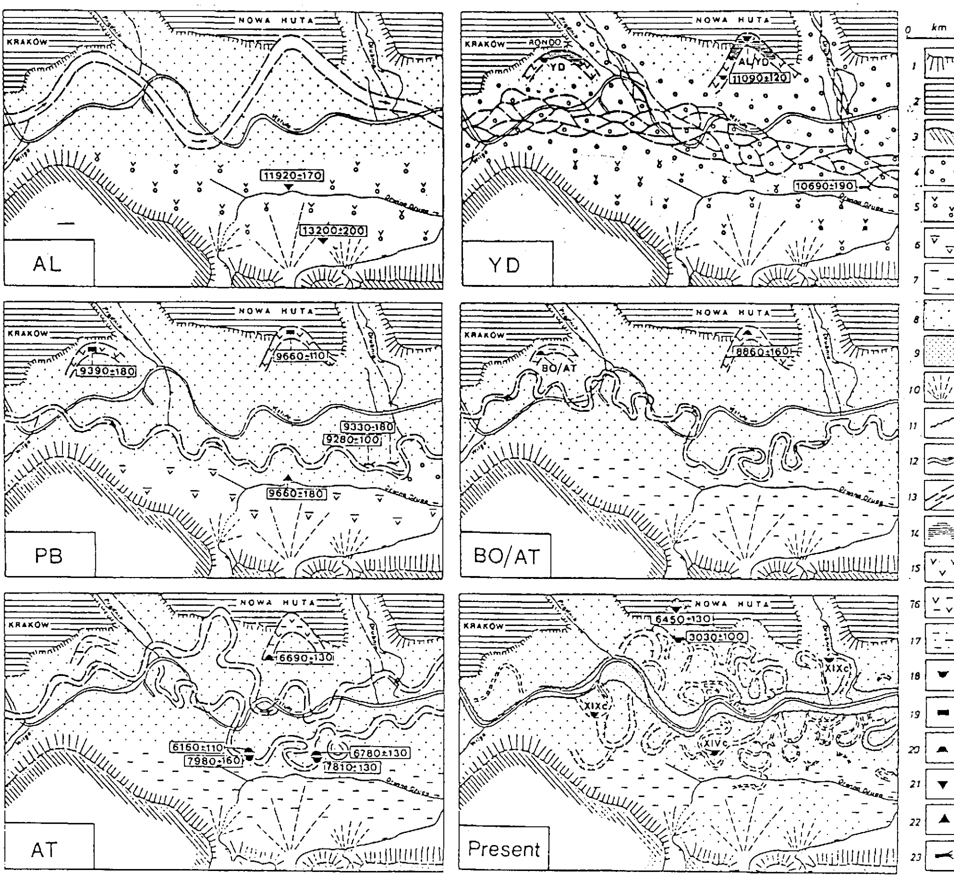

Fig. 1. Palaeogeographical reconstruction of the Vistula river valley near Nowa Huta and present situation. 1, Morphological edges; 2, overflood terraces of the Vistula river; 3, Gdów Divide; 4, active braided alluvial plain; 5, older braided alluvial plain overgrown by peat; 6, peat-bogs on the older braided alluvial plain covered by muds; 7, braided alluvial plain covered by muds; 8, flood plain; 9, modern flood plain (inter-dyke area); 10, alluvial fans; 11, dykes; 12, present Vistula channel; 13, reconstruction of the Vistula channel; 14, oxbow lakes; 15, peats; 16, clayey peats; 17, sandy muds; 18, cutoff radiocarbon, palynological (indicated by letters), historical and cartographical datings; 19, peat initiation in abandoned channels radiocarbon datings; 20, superposition of the muds and clayey peats on peats in abandoned channels radiocarbon and palynological (indicated by letters) datings; 21, peat initiation on braided alluvial plain radiocarbon datings; 22, superposition of the muds on peats on braided alluvial plain radiocarbon datings; 23, channel deposits radiocarbon datings. AL, Allerød; YD, Younger Dryas; PB, Preboreal; BO, Boreal; AT, Atlantic.

and dating of a number of profiles to characterize the type of sedimentation at various periods in different parts of the floodplain, enable the recognition of changes which are related with climatic variation and flood frequency, and which were a result of changes in local conditions such its location and distance from the active Vistula river channel. Analysis of palaeochannel and backswamp fills have provided a sedimentological and palynological record of floodplain evolution which is described below.

During the Allerød the Vistula river flowed in large meanders in the northern part of the valley, cutting through the Pleistocene terraces and alluvial fans of the Dłubnia and Prądnik rivers (Kalicki 1991*a*). Towards the end of this period, in connection with cooling of the Younger Dryas, the Vistula river abandoned a number of its meanders (Mamakowa 1970; Kalicki 1987, 1992*a*, *b*; Kalicki & Krąpiec 1991*a*; Kalicki & Zernickaya 1995) and changed into a braided or anastomosing channel with a tendency to aggradation (Fig. 1). Floodplain alluviation occurred in the southern part of the valley (Kalicki 1991*a*, 1992*a*, *b*). At the beginning of the Holocene the Vistula once again became a meandering channel (Kalicki 1991*a*). Silty muds ($M_z = 6.5$–7.0ϕ) were accumulated locally near the channel in the backswamps (Kalicki 1992*a*). At the Boreal–Atlantic transition a phase of increased activity in the Vistula commenced (Kalicki 1991*a*) and peats became covered by sandy muds in the palaeochannel located close to the active channel (Mamakowa 1970). Frequent flooding of the whole valley bottom resulted in an increase of the clay content in the peats of the palaeomeanders situated further from the river (after 8860 BP) (Kalicki 1992*b*). At the next stage the meander near Rybitwy became cut off, and later, around 8000 years BP there was channel avulsion to the north (Kalicki 1991*b*, 1992*b*) and an increase in the clay content of peats in the Nowa Huta palaeomeander. Simultaneously, on the other side of the valley, accumulation of gyttia and growth of peats commenced in the Boreal abandoned meander. This occurred at the end of the phase of increased activity (7980–7810 BP) when the Vistula was situated far from these palaeochannels and the connection between the active channel and the abandoned channel was broken (Kalicki 1991*b*, 1992*b*).

During the period that the Vistula flowed through the Allerød palaeochannel in the Nowa Huta area was active, clayey peats developed in which the proportion of minerals was inversely proportional to the distance from the active channel (Kalicki 1992*b*). The short distance from the Vistula and frequent flooding connected with it meant that peat bogs from the beginning of the Atlantic were often covered. Thus, after the Vistula avulsion from the south, the growth of peats was disturbed throughout the Mezo- and Neoholocene (Kalicki & Zernickaya 1995).

At the beginning of the next phase of increased river activity, a higher flood frequency resulted in the peats in the Nowa Huta palaeochannel (6690 years BP) being covered by sandy muds ($M_z = 4.5$–5.0ϕ) (Kalicki 1992*b*). This situation occurred only in the immediate vicinity of the Vistula and could have resulted either from the vicinity of the channel or from a large flood magnitude. Further from the active channel in the same palaeomeander strong clayey and sandy peats developed. Simultaneously, on the other side of the valley in the Boreal palaeomeanders organic sediments (6780 BP) were covered by clayey muds ($M_z = 8.0$–9.0ϕ) which indicates that floods covered the whole valley bottom. During the next stage (6450–6250 BP) the meanders were cut off (Kalicki 1991*b*; Wasylikowa *et al.* 1985) and the river channel moved to the south. This resulted in the covering of organic sediments by clayey muds ($M_z = 8.0$–9.0ϕ) in the next Boreal palaeomeander in the southern part of the valley (6160 BP) (Kalicki 1991*b*).

The above description shows that the evolution of the Vistula river valley has resulted from the interaction of climatic and local factors. In terms of sediment formation the local conditions could either magnify the climatic impulse or be dominant and completely blur it. In the first case, at the Boreal–Atlantic transition claying of peats occurred in the palaeomeander in Nowa Huta which resulted from a higher frequency of flooding in the valley bottom. Claying of peat increased after the avulsion of the Vistula channel to the north in the vicinity of the palaeochannel. In the second case, throughout the Meso- and Neoholocene, the vicinity of the active channel caused claying of peats and probably interrupted its growth which made it impossible to determine periods of more frequent flooding from the channel profile.

Flood phases during the Neoholocene

In the Neoholocene several phases of increased fluvial activity can be distinguished. The cause of these phases is disputable. According to archaeological evidence the development of agriculture has influenced the environment (Needham & Macklin 1992), and because of this many authors have stated that human activity was responsible for changes in fluvial activity during the younger Holocene (e.g. Mensching 1951; Jäger 1962; Mäckel 1969; Schirmer 1983; Brown 1987; Klimek 1988). The distinction between the climatic influence and the human impact on the fluvial activity is often difficult, if not impossible, to elucidate in many areas since both factors influenced the channel simultaneously (Klimek & Starkel 1974; Macklin *et al.* 1992). This particularly applies to the western part of Central Europe where human interference began in the Neolithic and increased during the Neoholocene. Human interference occurred in Poland some time later, but by the Bronze Age there was already human occupation in the loess-covered uplands and the Carpathians (Valde-Nowak 1988; Kruk 1988). A different situation occurred on the Russian Plain where human occupation occurred later (Chotinski & Starkel 1982). This makes these regions suitable locations to study the relative influence of climatic and anthropogenic factors on Neoholocene alluviation.

The first argument for regarding the climate as the main factor responsible for changes in river activity is the occurrence of changes during the same periods in rivers throughout Central Europe. For example, the period 3500–3000 BP of activity in the Vistula valley near Cracow (Kalicki 1991*a*) is also evident in the stratigraphic record of other river valleys (Kalicki & Krąpiec 1991*a*) from the Danube valley on the west (Buch 1990) up to the Berezina valley on the east (Kalicki 1991*c*). Similarly, earlier in the Holocene there are factors which indicate the increase of humidity and cooling of the climate, for example, the Löbben glacier advance in the Alps (Patzelt 1977; Bortenschlager 1982), increase of lake levels (Ralska-Jasiewiczowa 1989) and changes in the vegetation in the Carpathians (Obidowicz 1988). It cannot be eliminated, however, that some of these changes were caused or increased by volcanic eruptions such as Santorini (3370 years BP) and Hekla (3100 BP) (Hammer *et al.* 1980; Baillie & Munro 1988; Kalicki & Krąpiec 1991*a*). However, the development of Lusatian culture during this period causes Klimek (1988) to relate mud accumulation in the Oświęcim Basin during the last 3000 years BP exclusively with human activity. Detailed analysis of Klimek's mud profiles leads, however, to the identification of several stages of mud accumulation (after 2710, after 2150 and after 1220 BP) which relate to the phases of increased activity identified within Central-European river valleys (Kalicki 1996*b*). This distinction does not contradict the human influence on mud accumulation on floodplains, especially in smaller valleys (e.g. Kosmowska-Suffczynska 1983; Śnieszko 1985; Alexandrowicz 1988; Rutowski 1991). An increase in slope erosion caused fertilization of sites on the floodplain and resulted in the structural change of oak wood from thin ringed (Becker & Frenzel 1977). This change occurred, however, in different periods in Germany and Poland (Becker 1982; Krąpiec 1992).

A better argument for regarding climatic conditioning, rather than human impact, as the cause of increased fluvial activity is the phase which occurred around 1000 years BP (Kalicki 1993). This phase is marked very clearly in the Vistula valley near Cracow by aggradation, mud accumulation and deposition of tree trunks within the alluvia (Radwański 1972; Kalicki 1991*a*; Kalicki & Krąpiec 1991*b*, 1995). Traces of this phase can also be found in other valleys of Central Europe (e.g. Alexandrowicz *et al.* 1981; Becker 1982; Havlicek 1983; Niedziałkowska *et al.* 1985; Sokołowski 1987). Also during this period changes of sediment type and vegetation occurred in the valleys on the Russian Plain (Zolotokrylin *et al.* 1986; Klimanov & Serebrannaya 1986), which indicate that climate cooling occurred (Zernickaya & Kożarinov 1988). This phase was evident both in the unmodified, forested, Berezyna drainage basin, which has widespread peat accumulation, (Kalicki 1991*c*) and in the Dnieper drainage basin which had been extensively modified by human activity (Kalicki & Sanko 1992). In the wide Berezyna valley the organic fill of the palaeochannel cutoff (dated at 3120 ± 40 BP) was covered by muds around 1000 ± 50 BP. This was probably caused by the change of the river channel (Kalicki 1991*c*). In the narrow

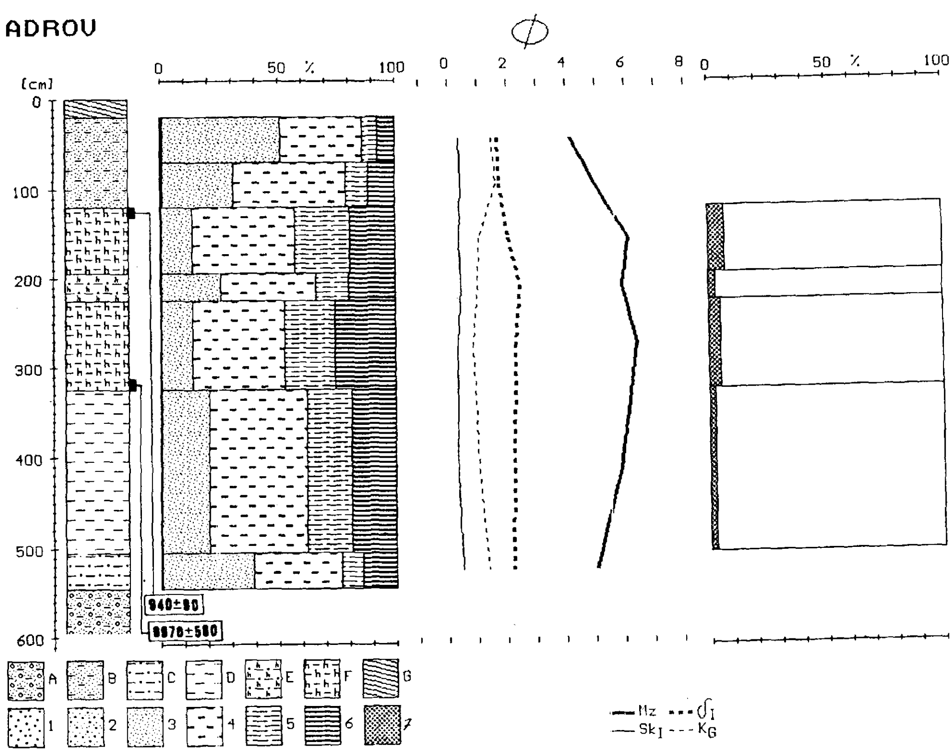

Fig. 2. Grain-size change of overbank deposits at Adrov site in the upper Dnieper valley (with Folk–Ward parameters). A, silty sands with gravels; B, silty sands; C, sandy silts; D, silts; E, organic sandy silts; F, organic silts; G, soil: Fractions: 1, coarse sand (−1 to 1ϕ); 2, medium sand (1–2ϕ); 3, fine sand (2–4ϕ), 4, coarse and medium dust (4–6ϕ); 5, fine dust (6–8ϕ); 6, clay (above 8ϕ); 7, organic matter content.

Dnieper valley, soil developing on silty muds ($M_z = 6.0\phi$) was covered by sandy muds ($M_z = 4.0\phi$) around 940 ± 90 years BP (Fig. 2). This indicates an increase in both the magnitude and frequency of flooding. Thus, a phase of increased fluvial activity occurred both in the natural drainage basin and in the basin that had been modified by human activity. The difference between the basins is that in the Dnieper basin the climatic impulse was stronger and resulted in changes of not only the frequency but also the magnitude of the floods (evident in the change from silty to sandy muds and flooding of higher levels of the floodplain) (Kalicki 1993).

Dendrochronological analysis can provide a more precise correlation between river valleys. In both the early and late Holocene there is a good correlation between the phases of fallen trees in various Central European river valleys (Becker 1982; Delorme & Leuschner 1983; Kalicki & Krąpiec 1995) which shows the domination of the climate in causing the increase of bank erosion. In the Neoholocene the influence of the climate could, however, be stimulated to a greater degree by human activity. Deforestation of floodplains contributed to the lateral migration of river channels and deposition of tree trunks in the alluvia (Kalicki 1991*a*). The occurrence of deforestation probably explains why the phases of erosion of different intensities are evident in the Roman and Early medieval times on various Central European rivers (Kalicki & Krąpiec 1996).

Conclusion

The above discussion indicates that phases of increased fluvial activity during the Neoholocene were connected with climatic changes. The impact of human activity merely contributed to the magnitude of changes which had already developed as a result of climatic controls (Kalicki 1993). Deforestation on the floodplain contributed to the lateral migration of channels and to the increase of tree trunk deposition in the alluvia (Kalicki 1991*a*). In the upper Dnieper drainage basin deforestation caused both the magnitude and frequency of flooding to increase and this resulted in changes in the content of muds and inundation of higher levels of the floodplain. An analogous process of thickening of muds connected with human activity can be identified in the Vistula basin, although it occurred at different times in different sections (Klimek & Starkel 1974; Alexandrowicz *et al.* 1981; Kalicki 1991*c*; Pożaryski & Kalicki 1995; Kalicki 1996*b*).

Deforestation and agriculture caused destabilisation of the fluvial system and contributed to both minor climatic changes and also single events in the river valleys. This may have resulted in the 'concentration' of Subatlantic phases (a large number of phases in a short time period) which can be identified in the stratigraphical sequence of a number of valleys (e.g. Schirmer 1983; Schellmann 1990) which compliments the better distribution of the more recent periods derived from historical sources (Petts *et al.* 1989; Strasser 1992).

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River response to the last neoglacial (the 'Little Ice Age') in northern, western and central Europe

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Abstract: Climate changes since AD 1200 have been of high magnitude. Significant lowering of temperatures occurred during the neoglacial ('Little Ice Age'), between AD 1200–1400 and AD 1600–1800 with maximum cooling in the mid-late eighteenth century. At this time many European valley/cirque glaciers reached their maximum extent since the late Pleistocene. Neoglaciation was followed by an overall warming trend, although with significant reversals superimposed. Alongside these temperature changes were variations in the nature and amount of precipitation, and in consequence, river basins in north, west and central Europe experienced enhanced fluvial activity between 1250 and 1550 and particularly between 1750 and 1900. These phases coincide with periods of climatic transition; cooling after the Medieval optimum and warming during the latter stages of the Little Ice Age respectively. In contrast, the intervening period (1550–1750), which corresponds with the most severe phases of the last neoglacial, was associated with lower rates of fluvial activity.

A number of recent studies, including several reported in this volume, have highlighted the variability of flood frequency over historical timescales in a range of river environments in Europe, North America and Australia (e.g. Knox 1984; Warner 1987; Bravard 1989; Balling & Wells 1990; Passmore *et al.* 1993; Rumsby & Macklin 1994; Baker this volume; Benito *et al.* this volume; Hooke this volume). In particular, they identify alternating periods (10^0–10^2 years in length) of enhanced and diminished flood frequency, that closely correspond with abrupt short-term climatic changes, and parallel changes in channel and floodplain stability. These findings have several important ramifications for river basin engineers and managers. First, non-random fluctuations in flood frequency invalidate traditional assumptions of stationarity in flood series over time periods of 10^2 years or less. Second, in the context of predicted future climatic changes, the links between short-term climatic variation and river stability are crucial. To date, most attempts at modelling the impact of future climatic change on river basins have focused on rainfall-runoff relationships and not considered channel and floodplain activity in any detail (Chang *et al.* 1992; Leavesley 1994). Clearly, historical studies of river response have much to offer in this respect, providing information on the rate and scale of fluvial response to high-frequency climatic change. However, we now need to move beyond the study of single reaches and basins towards an examination of the range of variability, and relative susceptibility, of river systems at a regional or sub-continental scale in order to develop a generally applicable model of river response.

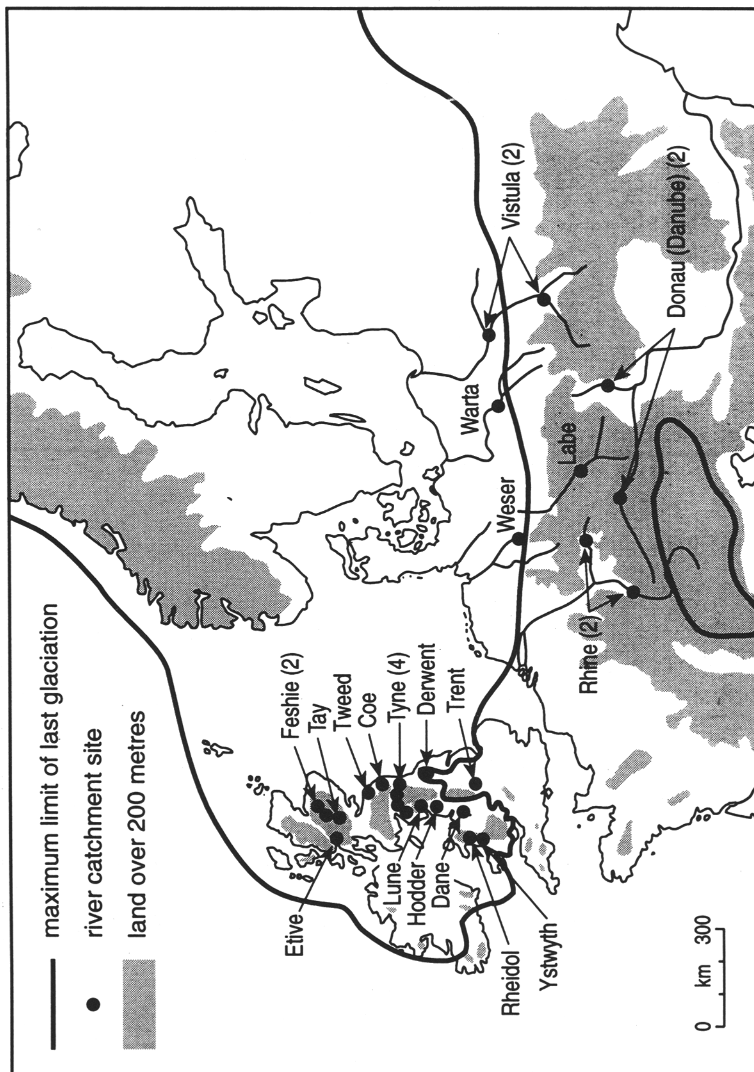

Fig. 1. Map showing location of study basins. Upland areas (>200 m) and limits of the last glacial maximum are also indicated.

This paper reviews published literature on the timing and nature of channel and floodplain activity in north, west and central European river basins (between latitude 45° and 60° N and longitude 5° W and 25° E; Fig. 1) for the period *c.* AD 1300–1900 and examines the impact of short-term changes in climate and flood frequency on river behaviour. The time frame encompasses the 'Little Ice Age', the most recent Holocene neoglaciation, which was characterised by reduced annual temperatures, mountain glacier advance and increased geomorphological activity (Grove 1988). The magnitude of temperature and rainfall variations since *c.* AD 1300 are of the same order as those predicted for the next 40 years or so (temperature ±1–2°C, precipitation ±10–20%, Knox 1993) and may therefore provide a useful analogue for river response to future climate change. All of the study basins are within the warm temperate climate region and have similar meteorological causes and potential for flooding (Hayden 1988). Flooding results from frontal cyclones occuring all year round, although most frequent in winter, and rain on snow, with snowmelt important in mountainous areas. Thus, a winter or spring runoff maximum occurs. Basins draining the Mediterranean region of southern Europe are excluded from the study as the hydroclimatic causation of floods is different. Here summer thunderstorms are the principal means of flooding; winter fronts and cyclones produce only moderate precipitation.

Timing and nature of the 'Little Ice Age'

The 'Little Ice Age' is the most recent Holocene neoglaciation (period of glacier advance, Porter & Denton 1979) and may have been 'the most significant of all the periods of glacier expansion that have occurred since the last Ice Age' in Europe (Bradley & Jones 1995, p. 2). It was first recognised and defined in the European Alps and Scandinavia, where there is widespread evidence that valley glaciers had reached positions well beyond their twentieth century limits during the recent past (Grove 1988). Subsequently, a large body of evidence accumulated to support a Europe-wide 'Little Ice Age' during the sixteenth to nineteenth centuries, including glacier advance in Scandinavia, widespread flooding of rivers in central Europe, expansion of sea ice in the North Atlantic (Grove 1966, 1979, 1988).

More recently, a number of high-resolution reconstructions of climate, based on a range of proxy and documentary sources have demonstrated greater complexity than originally envisaged, in two key respects: (i) recognition that the 'Little Ice Age' was not a single cooling phase, but rather comprised a series of 'complex climate anomalies with both warm and cold episodes' (Bradley & Jones 1995, p. 659), with the most widespread cool period occurring during the seventeenth and nineteenth centuries (Bradley & Jones 1995); (ii) identification of 'Little Ice Age' glacier advances outside of Europe, including North America (Luckman 1993; Wiles & Calkin 1994), and the Himalayas (Hughes 1995) and increasing evidence of climate (temperature and rainfall) fluctuations of similar magnitude in semi-arid and subtropical regions (D Arrigo & Jacoby 1991; Grove & Conterio 1995). Most recently, isotope evidence from Antarctica and the Quelccaya ice cap, Peru indicates southern hemisphere cooling phases between the sixteenth and nineteenth centuries (Mosely-Thompson *et al.* 1993; Thompson 1995). Hence, the 'Little Ice Age' comprised a series of widespread, episodic, cooling phases, possibly global in extent but not necessarily synchronous.

The question of the extent and timing of 'Little Ice Age' climate changes is vital in the context of identifying potential causal mechanisms. Three forcing factors have received most attention: variations in the volcanic aerosol loading of the atmosphere, variations in solar output and changes in coupling of the ocean–atmosphere system (e.g. variations in rates of North Atlantic Deep Water (NADW) formation). Rind & Overpeck (1993) have attempted to differentiate the spatial and temporal signatures of each of the potential mechanisms using GCM experiments. They suggest that solar forcing (e.g. sunspot minima) is the most viable explanation if the cooling episodes were global in extent. If, however, cooling was not globally synchronous then regional or hemispheric forcings such as volcanic eruptions or changes in ocean circulation are possible. The cooling effect of volcanic aerosols has a rather short timescale, lasting up to 4 years depending on the nature of the eruption, location of the volcano and atmospheric conditions (Bradley 1988; Stuiver *et al.* 1995) and the chronology of large explosive eruptions, although incomplete, suggests they do not occur with sufficient frequency to generate decadal-length cooling episodes (Rind & Overpeck 1993). The role of ocean circulation changes on Holocene climates is unclear at present and it may be that they operate over longer (century to millennia) periods (Stuiver *et al.* 1995). It is possible that several forcing mechanisms acted in combination to produce 'Little Ice Age' climatic changes.

Characteristics of the European 'Little Ice Age'

The timing and magnitude of European climate fluctuations over the last 700 years are now known in some detail. Instrumental records of temperature and rainfall are available for many locations since the late seventeenth century and, prior to that, a range of high resolution proxy records has been established, primarily based on tree ring, documentary and phenological evidence (see Bradley & Jones 1995 for a good summary of sources). Several general trends in temperature and precipitation are apparent (Fig. 2). The first onset of cooling occurred across northern, western and central Europe in the second half of the sixteenth century and continued through the seventeenth century. A warmer phase in the early eighteenth century was followed by a second cooling in the late eighteenth–early nineteenth century. The nineteenth century appears to have been characterized by a cool–warm–cool oscillation in many locations, but with overall amelioration into the twentieth century.

The availability of daily records of precipitation and temperature since the late seventeenth century allows annual and seasonal trends to be examined more closely (selected records are illustrated in Fig. 2a). Marked seasonal contrasts in temperature and precipitation are clear during the seventeenth to nineteenth centuries. Winter and spring temperatures were generally well below mean values across Europe, although Berlin experienced warmer winters 1740–70. Summer and autumn months, however, were relatively warm, especially in the second half of the eighteenth century, indicating more extreme seasonality than present. The England and Wales precipitation series was characterized by above average annual rainfall in the late eighteenth and late nineteenth centuries, with the intervening period close to the long term mean. Again, there are seasonal differences, with relatively dry winters and wetter summers 1750–1900. However, there are problems linking average annual rainfall totals and peak storm rainfall and the relationship between rainfall and river flow is not straightforward. The proportion of runoff to rainfall is likely to have been

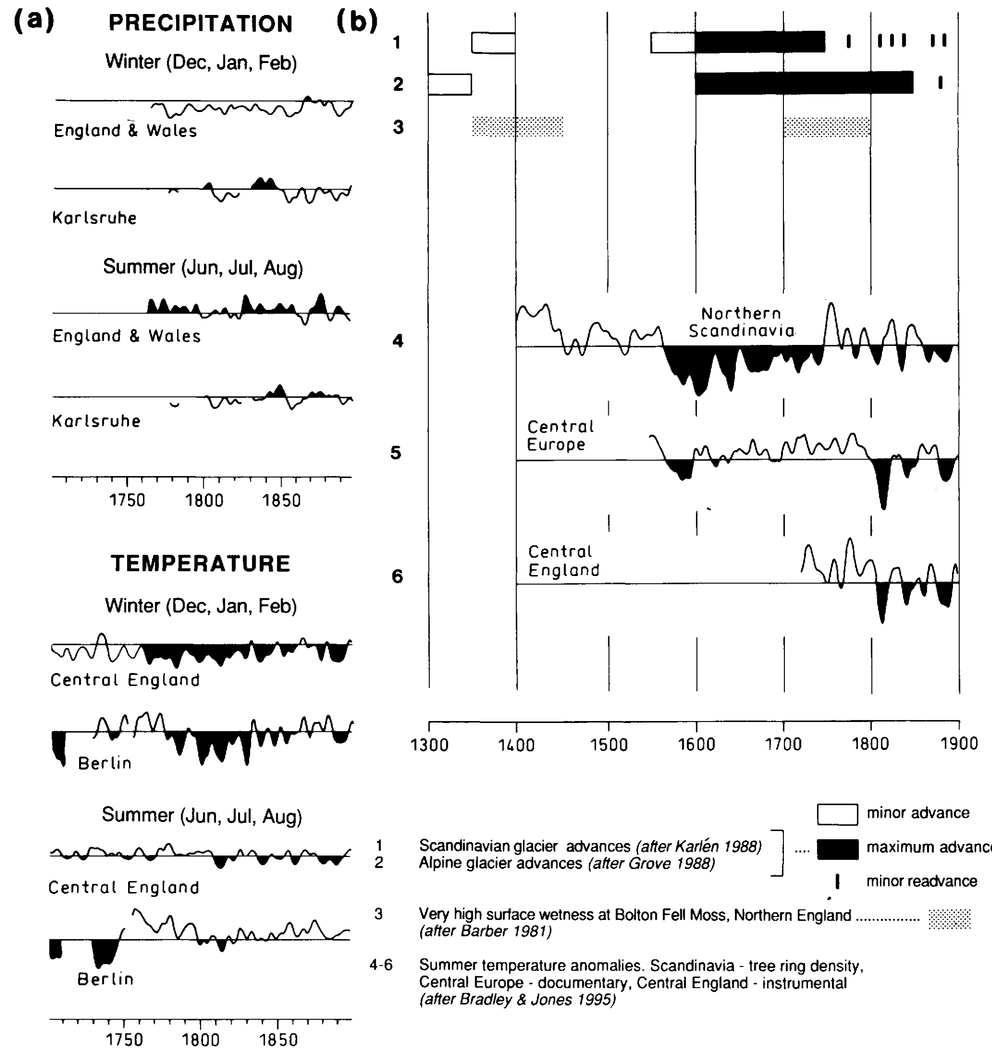

Fig. 2. (**a**) Instrumental records of summer and winter temperature and precipitation at selected European locations, AD 1700–1900 (after Jones & Bradley 1995), (**b**) various proxy records of temperature and precipitation, AD 1300–1900.

higher in cooler periods with reduced evapo-transpiration rates (saturated catchments) and frozen ground. In addition, increased snowfall contributions, particularly between 1690–1705, 1745–90, 1810–15 and 1845–50 (Manley 1969), would lead to increased snowmelt enhancement of floods (e.g. Archer 1992).

High-resolution proxy evidence of climate variations is available for several hundred years before AD 1500. However, temperature and rainfall estimates need to be treated with care as most proxy data is seasonally-specific and may respond to a combination of climatic parameters. For example, tree ring density reflects growing season temperature and/or rainfall and may be unrepresentative of the year as a whole (Jones & Bradley 1995). Figure 2b illustrates several proxy climate reconstructions for the fourteenth to sixteenth centuries. A common feature of many of these records is a period of significantly reduced temperature, and increased wetness, prior to the 'Little Ice Age', centred on AD 1300, corresponding with glacier advance in Scandinavia (Evans *et al.* 1994, Karlén 1988) and the Alps (Grove 1988).

To summarize, the European 'Little Ice Age' between 1550 and 1850, was characterized by a cooler and wetter climate in many locations, with particularly severe conditions in the period 1700–1850, coinciding with maximum advance of valley glaciers and increased flooding. An earlier climatic deterioration, centred on the thirteenth/fourteenth century, was also characterised by glacier advance and increased surface wetness. The intervening period, 1450–1550, appears to have been milder and drier, with especially warm summers in Scandinavia.

River response data

The data on river response included in this review are limited to published sources and to studies with reliable dating control, giving a total of 26 reaches in 18 basins. Table 1

Table 1. *List of study basins, showing sub-catchment, nature of dating control and authors*

Basin	Sub-catchment/ location	*Dating control	Reference(s)
Feshie	Upper	^{14}C, C, S	Brazier & Ballantyne (1989)
	Middle		Robertson-Rintoul (1986)
Etive		^{14}C	Brazier *et al.* (1988)
Tay		C	Gilvear (1993)
Tweed	Bowmont Water	^{14}C, A, C, P	Tipping (1994)
Coe		^{14}C, TM	Macklin *et al.* (1992*a*)
Tyne	Thinhope Burn	^{14}C, C, L, P	Macklin *et al.* (1992*b*), Macklin (1994)
	Featherstone	^{14}C, C, L, PM	Passmore *et al.* (1993)
	Broomhaugh Island	C, TM	Rumsby & Macklin (1994)
	Scotswood	^{14}C, C, TM	Passmore *et al.* (1992)
Lune	Middle Langdale	^{14}C, C	Harvey *et al.* (1981)
Derwent	Jugger Howe Beck	^{14}C	Richards *et al.* (1987)
Hodder	Langden Valley	^{14}C, C	Harvey & Renwick (1987)
Dane		^{14}C, C	Hooke *et al.* (1990)
Trent		^{14}C, A, C	Salisbury *et al.* (1989)
Rheidol		^{14}C, C	Macklin & Lewin (1986)
Ystwyth		C, TM	Lewin *et al.* (1983)
Weser		A, C, D	Bork (1989)
Warta		C	Kozarski & Rotinicki (1977)
			Kozarski (1991)
Vistula		^{14}C, A, C	Starkel (1981, 1985)
			Starkel (1991)
			Falkowsi (1975)
Rhine		^{14}C, A, DC	Schirmer (1988)
Main		^{14}C, A, C, DC	Becker & Schirmer (1977)
			Schirmer (1983, 1988)
Labe		^{14}C, A	Ruzickova & Zeman (1994)
Moravia		A, D, S	Havilicek (1983, 1991)
Donau		^{14}C, A, TM	Buch (1989)

* ^{14}C, radiocarbon; A, archaeology; C, cartography; D, documentary; DC, dendochronology; L, lichenometry; P, pollen; PM, palaeomagnetic; S, soil development; TM, trace metals.

Table 2. *Characteristics of study basins, indicating area, altitude, relief, underlying geology and whether glaciated at the last glacial maximum (where available)*

Basin	Sub-catchment/ location	Area (km^2)	Altitude of study reach (m)	Relative relief (m)	Geology	Glaciated at LGM?
Feshire	Upper	< 1	390	725	Schists	✓
	Middle	240	122	993	Schists	✓
Etive		< 1	160	795	(granites, rhyolite lavas)	✓
Tay		4690	(100)	(1100)	Dalradian schists, grits, slates	✓
Tweed	Bowmont Water	(?)	190	524		✓
Coe		0.75	210	50	Carboniferous sandstones	✓
Tyne	Thinhope Burn	12	272	310	Carboniferous sediments	✓
	Featherstone	322	140	753	Cardoniferous sediments	✓
	Broomhaugh Island	1918	30	863	Carboniferous sediments	✓
	Scotswood	2337	<10	885	Carboniferous sediments	✓
Lune	Middle Langdale		200	480	Silurian siltstones and shales	✓
Derwent	Jugger Howe Beck		120	170	Jurassic sediments	✓
Hodder	Langden valley	14	198	352	Carboniferous sandstones	✓
Dane		150	42–70	510	Triassic and Carboniferous sediments	✓
Trent		20			Triassic Mercia mudstone group	
Rheidol		182	25	727	Silurian siltstones and shales	✓
Ystwyth		116	60	510	Silurian siltstones and shales	✓
Weser		46 136	<200		Mesozoic marls and clays, loess Tertiary sediments, limestone	✓
Warta		25 053	60	402		Lower reaches only
Vistula		198 000	150		Cretaceous limestone and Jurassic	Lower reaches only
Rhine			125		Tertiary sediments	Alipine tributaries only
Main			<200			×
Labe.			178			Lower reaches only
Donau			320			×
Moravia		26 580	139			×

lists the basins, including sub-catchments, authors and dating control. All ^{14}C dates have been calibrated to calendar years (Stuiver & Becker 1986). Nearly every study has used at least two dating methods, with ^{14}C, cartographic, archaeological and documentary methods most common.

Figure 1 shows the location and distribution of river basins included in the study and Table 2 details catchment characteristics. Several common characteristics are

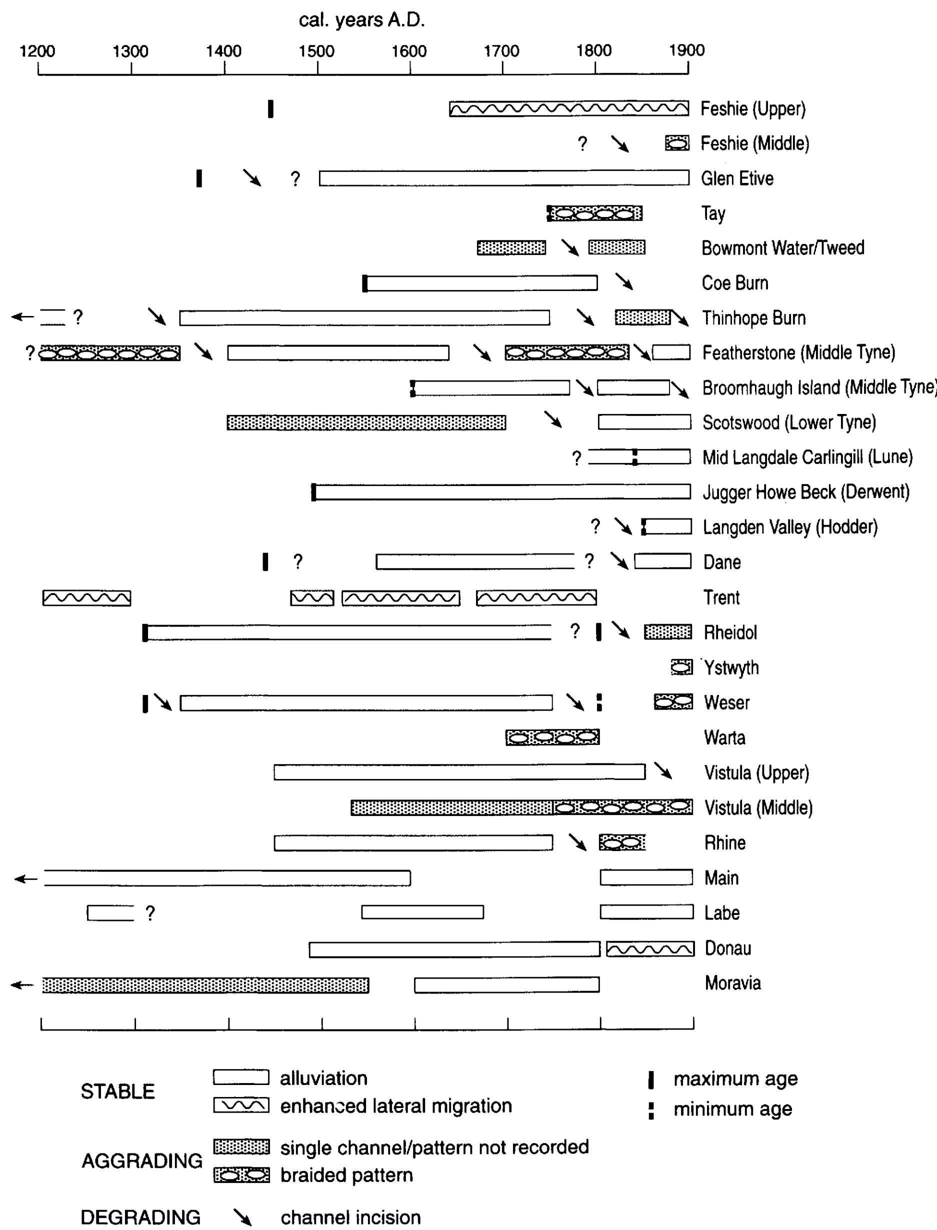

Fig. 3. European river response AD 1200–1900. Basins are listed by latitude, with the most northerly at the top of the diagram.

apparent. Most are located in the middle and upper reaches of river basins, at elevations generally above 200 m and with high relative basin relief, and are within or near the limits of the last glaciation.

River activity between AD 1300 and 1900 is indicated in Fig. 3 whereby basins are listed by latitude, with the more northerly sites at the top of the diagram. Three modes of river activity are differentiated: lateral accretion and overbank sedimentation in vertically stable meandering–wandering river systems, channel aggradation, normally associated with braided or near braided rivers and channel incision. Braiding in many of the catchments discussed in this review appears to have been the consequence of higher rates of sediment supply to trunk rivers from bank erosion, tributaries or valley-side slopes. In some cases (Starkel 1991) it was a response to a short-term increase in erosion and sediment transport following a major flood (or series of floods). In others (Falkowski 1975; Passmore *et al.* 1992) it was the result of long-term alterations in flood regime or catchment land-use change. Incision, on the other hand, usually followed a decrease in sediment supply (Hooke *et al.* 1990) but also occurred where channels were adjusting to large floods especially where trunk rivers were poorly coupled with valley slopes and tributary streams (Macklin *et al.* 1992*b*). Construction of flood protection embankments, dredging of river channels and alluvial aggregate extraction also resulted in local incision. Periods characterized by high rates of lateral channel migration are highlighted in Fig. 3. These probably reflect an increase in flood frequency but not necessarily flood magnitude, as changes in the latter would have been more likely to cause river instability and metamorphosis (e.g. transformation of a single to a multi-thread channel, Passmore *et al.* 1993).

Timing and nature of river response in NW Europe

To enable a more detailed examination of spatial patterns and causal relationships of river response, dates marking changes of river activity are re-plotted in Fig. 4 at 50 year intervals. The basins are again listed by latitude (from north to south). Notwithstanding the likely bias towards preservation of more recent evidence, both sedimentary and documentary, some general patterns are evident. First, an overall increase in activity is recorded over the period AD 1300–1900, especially after the sixteenth century. Second, a contrast in the nature of river response before and after *c.* 1700 with channels generally being vertically stable prior to the seventeenth century, whereas from the eighteenth century river instability was much more widespread with marked changes (both aggradation and incision) of river bed levels. Three phases of river activity can be identified, 1300–1550, 1550–1750 and 1750–1900.

1300–1550. Incision is recorded in three basins (Tyne, Weser and Etive) before 1450, coinciding with the short-lived climatic deterioration in the fourteenth century. In the Tyne basin, for example, there is documentary evidence for extreme flooding during the period, with the 'most disastrous flood' (although not necessarily the largest discharge) on the river occurring in 1339 (Archer 1992). Most rivers during this period, however, appear to have been experiencing little vertical change though many were actively laterally reworking their valley floors.

1550–1750. Changes in river activity are recorded in only nine out of the 19 basins between these dates (Feshie, Bowmont, Coe, Tyne, Dane, Trent, Rheidol, Warta and

	1300 -49	1350 -99	1400-49	1450 -99	1500 -49	1550 -99	1600 -49	1650 -99	1700 -49	1750 -99	1800 -49	1850 -99
Feshie (Upper)							≈					
Feshie (Middle)											↘	⇑
Etive			↘		≈							
Tay										⇑		
Bowmont								↑		↘	↑	
Coe						≡					↘	
Tyne (TH)	↘	≡								↘	↑	↘
Tyne (FS)		↘	≡					↘	⇑		↘	≡
Tyne (BH)							≡			↘	≡	↘
Tyne (SW)			↑							↘	≡	
Lune											≡	
Derwent				≡								
Hodder											↘	≡
Dane						≡					↘	≡
Trent				≈	≈			↑				
Rheidol	≡							≡			↘	↑
Ystwyth												⇑
Weser	↘	≡								↘		⇑
Warta									⇑			
Vistula (Upper)				≡								↘
Vistula (Middle)					↑					⇑		
Rhine				≡						↘	⇑	
Main										≡		
Labe					≡						≡	
Donau				≡							≈	
Moravia							≡					

STABLE	≡	alluviation
	≈	enhanced lateral migration
AGGRADING	↑	single channel/pattern not recorded
	⇑	braided pattern
DEGRADING	↘	channel incision

Fig. 4. Summary diagram indicating changes in river activity for 50 year periods.

Moravia), with lateral accretion (associated with vertically stable channels) and braiding (associated with channel aggradation) being the dominant responses. Climatically, this was the severest part of the neoglacial, with especially cool summers recorded in Scandinavian tree rings and severe winters in NW Europe (Lamb 1977), coinciding with the maximum advance of glaciers in the European Alps, as well as in Scandinavia (Grove 1988). Total annual rainfall appears to have been generally lower than average and low frequency of flooding was recorded (Lamb 1984). Although flood frequency might have been lower the magnitude of some flood events, such as those resulting from snowmelt (e.g. Yorkshire Ouse 1625 and 1626), may have been larger during this period.

1750–1900. A marked response was recorded in nearly every basin, commonly with channel incision especially in the late eighteenth century followed by aggradation or lateral accretion. During the period there is an irregular warming trend, though with cool winters, variable summers and increased frequency of flooding (Lamb 1977).

Climatic causal linkages

Recent work undertaken by the authors in the Tyne basin, northern England offers some insight into possible climatic causal linkages that may be applicable to the European data set (Rumsby & Macklin 1994).

The Tyne Basin, northern England

Full details of catchment characteristics, sources and nature of evidence are given in Rumsby & Macklin (1994). Brief catchment characteristics are given in Table 3 and Fig. 5 summarizes the response of the River Tyne to high frequency (20–30 year) changes in climate and flood regime over the last three hundred years. Since AD 1700 fluvial activity has been characterized by alternating phases of river bed incision and stability coinciding with non-random, decadal-scale fluctuations in flood frequency and hydroclimate. Episodes of widespread channel bed incision in the late eighteenth century, late nineteenth century and mid twentieth century resulted from a higher frequency of large floods (>20 year return period) and cool, wet climate. Phases of more moderate floods (5–20 year return period) in the mid nineteenth century and early twentieth century were characterized by enhanced lateral reworking and sediment transfer in upper reaches of the catchment and channel narrowing and infilling downstream (Rumsby & Macklin 1994).

Trends in flood frequency and magnitude in the Tyne basin correspond with changes in large-scale upper atmospheric circulation patterns, specifically the strength, wavelength and amplitude of the circumpolar vortex. The circumpolar vortex (or jet stream) controls the position of storm tracks, movement of air masses and location of fronts in mid latitudes. Circumpolar upper air waves tend to alternate between two specific forms, zonal and meridional, thought to be controlled by temperature (Kutzbach 1970; Lamb 1977, 1982; Charney & DeVore 1979; Hirschboeck 1988). Under warmer conditions the circumpolar vortex is displaced northwards and flow is strongly zonal (west to east) with low amplitude, widely spaced waves. In cooler periods (including the 'Little Ice Age') the lateral temperature gradient is steepened and circumpolar air masses are shifted southwards, favouring more frequent and enhanced occurrence of meridional (north, east, south) wind patterns in mid latitudes (Lamb 1977, 1982).

Table 3. *Catchment characteristics of the Tyne basin*

	Basin area (km^2)	Gradient (mm^{-1})	Altitude (m OD)	Basin relief (m)
River Tyne	2927		893–0	893
Thinhope Burn	12	<0.01–0.1	620–180	440
Broomhaugh Island	1918	0.019	30	863
Low Purdhoe	2198	0.001	10	883

In the Tyne basin there is a clear relationship between extreme floods and meridional configurations of the circumpolar vortex (Rumsby & Macklin 1994). Two synoptic characteristics associated with meridional configurations favour large floods. First, high amplitude waves in the circumpolar vortex are liable to be

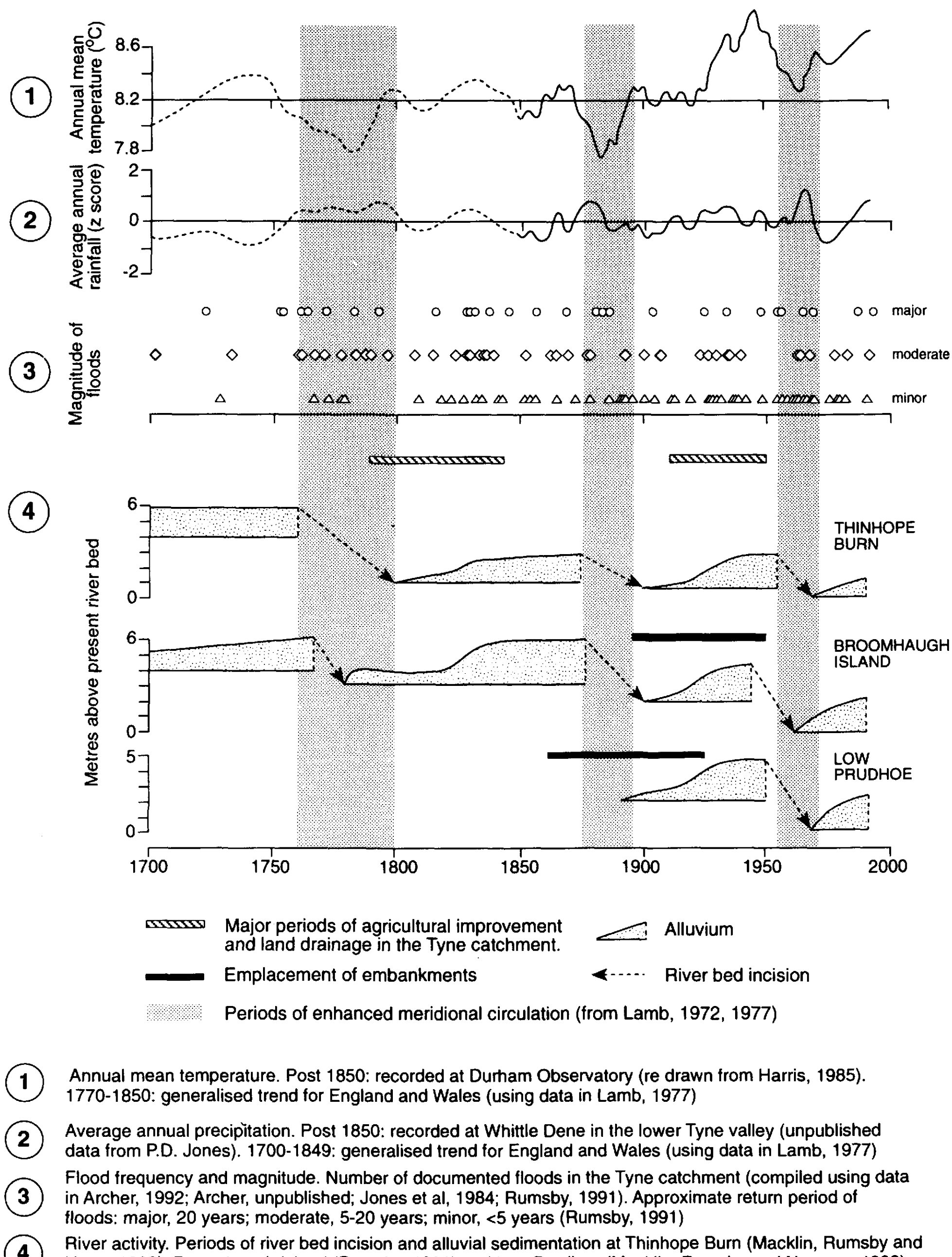

Fig. 5. Climate, flooding and river activity in the Tyne basin, northern England, AD 1700–1900 (modified from Rumsby & Macklin 1994).

associated with stationary blocking situations which can lead to exceptional (high intensity, multiple peak) rainfall events (Rodda 1970; Hirschboeck 1987). Second, low winter temperatures and increased prevalence of northerly and northeasterly (polar) air masses result in low rates of evapotranspiration and higher soil wetness, as well as increased occurrence of snow, all of which promote rapid runoff and large flood peaks. An additional factor, that makes the Tyne particularly sensitive to precipitation from easterly and northerly air masses is its geographical location. The catchment is located to the east of the Pennines, the major rainfall divide in northern England (Wheeler 1990; Wigley *et al.* 1984), and is in the rain-shadow of zonal weather systems travelling west to east, thus meridional weather systems, especially those from the north and northeast that have collected moisture over the North Sea, are more important in terms of precipitation extremes over the area.

Implications for European rivers

The behaviour of the River Tyne over the last 300 years suggests that an important climatic control on flood frequency in mid latitude river basins, at the regional scale, is the configuration and location of circumpolar upper atmospheric air waves which determine air mass source, and pathway, and hence the duration, intensity, distribution and type of precipitation (Rumsby & Macklin 1994). The geographical location of river catchments with respect to orographic barriers to airflow is also another important controlling factor.

Tentatively extrapolating from the Tyne investigations, we would suggest that basins draining to the north or east of major rainfall divides should experience increased frequency and magnitude of flooding during periods of enhanced meridional circulation, i.e. during cool phases. In contrast, those basins draining to the south and west of rainfall divides should experience increased flooding during warmer, zonal phases. The European data set does appear to follow these general trends, particularly for the period 1750–1900. In the late eighteenth century, under highly meridional airflows (Lamb 1977), many basins draining east and north (e.g. Tay, Bowmont, Weser, Rhine) experienced considerable river instability and channel bed incision. Whereas, after 1800, with reinstatement of zonal airflow, basins draining to the west and south became active (e.g. Lune, Hodder, Dane, Rheidol).

The geographical patterns of river response identified above accord with European rainfall trends obtained by Tabony (1981) in a principal component analysis of rainfall series from 182 stations. Superimposed on the dominant component, representing a 'uniform' pattern of annual rainfall variations of similar timing and magnitude across Europe, are secondary patterns with rainfall gradients from north-south and east-west. Strong orographic effects are apparent in northern Britain and adjacent to the Alps, particularly in winter months.

Conclusions

River basins in north, west and central Europe experienced enhanced rates of fluvial activity between 1250 and 1550 and particularly between 1750 and 1900. These phases coincide with periods of climatic transition, cooling after the Medieval 'optimum', and warming during the latter stages of the 'Little Ice Age' respectively. The intervening period between 1550 and 1750, corresponding with the severest phases of the last neoglacial was by contrast associated with lower rates of fluvial activity. The direction of climatic change (warming or cooling) is an important determinant of the nature and sensitivity of basin response, in combination with basin location (latitude, orientation with respect to orography). In the context of global warming predictions, the finding of this study suggest that we should expect those basins located to the west and south of rainfall divides to experience enhanced flooding and possible channel instability.

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Issues in scientific co-operation on information sharing: the case of palaeohydrology

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Abstract: The hydrological cycle is interactively integral with climate, so that for an understanding of future environmental change it is important to have hydrological data accessible from previous times to use in both hydrological and climate change modelling of past, and also of present and future environments and also for the development and operation of surface water resources. A database has been developed as part of the research undertaken by the INQUA Global Continental Palaeohydrology Project (GLOCOPH) to facilitate the preservation and access of palaeohydrological data within the public domain, and through this the organization, facilitation, association and bisociation of the data.

Although fluvial palaeohydrology is the study of river flows in the past before the commencement of continuous instrumental measurement (Schumm 1967; Gregory 1983, this volume), the data collected from palaeohydrological studies can be of considerable utility for the analysis of environmental and hydrological changes not only in the recent and geological past. Indeed, Baker (1991, this volume) has suggested that palaeohydrology should have a dual role and emphasize both the study of mechanisms of hydrological parameters in the past, and the development of reconstructions of past hydrological conditions to verify and test models to predict future hydrological change.

Palaeohydrological data can be divided into two main types; *primary* (or raw) data, which comprise measurements taken in the field or laboratory (e.g. channel dimensions, sediment size, palaeostage indicators and planform characteristics), and *reconstructed* data (e.g. velocity, discharge, sediment yield) that are derived from the primary data. Wohl & Enzel (1995) provide a review of the sources of data for palaeohydrological analysis. The results from these analyses can be used to document and investigate environmental changes over various timescales. In this sense palaeohydrological data are complementary to other environmental datasets that attempt to document the spatial distribution and timescale of environmental change. The following section discusses the context in which palaeohydrological data can be used for the study of environments in the past, present and future.

Palaeohydrological data and past, present and future environments

In addition to providing reconstructions of hydrological variables at the time at which they occurred (for example, the velocity and discharge of a river at a given date), which is the most widely used application of palaeohydrological analysis, data, particularly from the recent past (e.g. the Holocene), can be used for the study of occurrences further back in geological time. Baker *et al.* (1979, 1988*a*, *b*), for example, have used knowledge of Holocene flood processes in the study of ancient floods in the geological record.

In principle, palaeohydrological data can also act as proxy climate data, allowing assumptions to be made about the climate which was responsible for a particular hydrological regime. For example, the existence of palaeochannels in present-day deserts indicates that conditions wetter than present existed there in the past. In practice, however, this can be a difficult task to achieve since it relies upon possibly unsubstantiated assumptions about the nature of the independent variables in the hydrological system at the date of reconstruction (e.g. Petts 1989; Petts *et al.* 1989; Starkel this volume).

Some components of the present-day landscape cannot be understood without reference to past hydrological events, such as the channels incised by the super-floods which covered vast areas of North America at the end of the last major period of glaciation (e.g. Baker 1981; Baker & Bunker 1985). Similarly, abandoned channels, river terraces and ox-bow lakes all require information relating to past hydrological systems to explain their formation (e.g. Kozarski 1991). Thus many rivers in temperate and arid areas are 'underfit' in relation to the size of the surviving channel; these channels were probably formed during glacial or pluvial periods when discharges were higher and they are now palimpsest elements in the present morphology. Palaeohydrological techniques and data can be used to aid the interpretation of such features.

With respect to future hydrological changes, a current focus of research is the appraisal of the influence that a changing climate might have on water resources. This can be assessed using the 'temporal analogue' approach (e.g. Beran & Arnell 1989; Brown this volume) which uses data from the past to estimate the direction and magnitude of changes in a warmer climate in the future. Maizels & Aitken (1991), for example, argue that analogues from the highland areas of the UK during the Holocene deglaciation may be used as a basis for providing estimates of the magnitude and frequency of processes in regions which may become deglaciated in the future. To carry out this type of analysis it is necessary to have long hydrological records, particularly of river discharge, however, continuous instrumental hydrological records only exist over a period of decades. For example, the average length of record in the UK national flow archive is less than 20 years, and only *c.* 70 stations have a record of up to 30 years (Beran & Arnell 1989). A similar application is the use of palaeohydrological data as a baseline against which to assess the influence of anthropogenic factors in historic times. River regulation, for example, may have commenced before the beginning of instrumental records, so that palaeohydrological interpretation may be necessary to gauge the impact of that regulation (e.g. Gurnell *et al.* 1994).

There is thus a requirement for palaeohydrological data (particulary from the Holocene) which can be used in the analysis of potential future hydrological change, both natural and anthropogenic (Knox 1985; Thornes & Gregory 1991; Starkel 1993). Some authors have used historical weather data, historical documents and sedimentary evidence to reconstruct previous river flows and thus to extend the experimental record back into the past (e.g. Potter 1978; Wright 1978; Jones *et al.* 1984). Similarly, Stedinger & Cohn (1986), Stedinger & Baker (1987), Baker *et al.* (1988*a*, *b*) and Frances *et al.* (1994), among others, have used palaeoflood data to extend the contemporary flood frequency record. Studies of this type have been particularly valuable during the design of reservoirs and when locating nuclear waste repositories and similar structures to ensure they will be stable under the vicissitudes of future hydrological change and that the design meets set probabilities of flood occurrence (Baker 1983).

General circulation models (GCMs) are now also used extensively to inform studies of future environmental change. Calibration and validation of models against palaeohydrological data and reproduction of conditions known to have occurred in the past will be an important test of the ability of such models to analyse future change. In this respect palaeohydrological data can provide a resource which can be used as a basis for testing the past extrapolations by GCMs, and also to calibrate models beyond current environmental conditions towards potential future conditions (Waylen 1995).

The current resource of palaeohydrological data

Many studies have assembled palaeohydrological data which could potentially be used to support the areas of research discussed above. In particular, a large data collection exercise was encouraged during the International Global Correlation Project (IGCP) 'Palaeohydrology of the temperate zone' (e.g. Gregory *et al.* 1987; Church, 1989; Starkel *et al.* 1991). Although the majority of palaeohydrological studies have been focused on relatively small geographical regions due to the intensive sampling required for such analyses, the data themselves have an environmental, scientific, engineering and economic significance to areas wider than the specific location of data collection.

Despite the importance and intrinsic value of the palaeohydrological data, however, they are usually unavailable to the wider scientific community, being not easily disseminated from the countries, organizations and individuals who collected them. Datasets relevant to palaeohydrological research may also be in the domain of many disciplines, e.g. geomorphology, hydrology, geology, sedimentology, archaeology. Thus the cross referencing of many strands of data needs to be drawn together, facilitated by the Global Continental Palaeohydrology Project (GLOCOPH) (see Gregory this volume) or other data archives. Even where data have been discussed in publications the data considered are usually not published in full. This is a situation evident throughout many of the environmental sciences in that 'considering the time, effort and expense that generally goes into sampling and analysis, it is an unfortunate loss to the scientific community that those data are not readily available in a known format with sufficient information to make them useful to others' (Hopke & Massart 1986, p. 193).

A questionnaire of 40 scientists involved in palaeohydrological research has been conducted to determine how data are currently stored. The replies indicate that the majority of data are at present stored on the original paper record sheets (63% of respondents) and the remainder in personal computer packages (24%) or computer printout (13%). Thus, as suggested above the data are not readily available for use by other scientists. None of the respondents to the questionnaire had donated their data to national or international data collection efforts but many expressed an interest in doing so.

Within other disciplines there has been an appreciation that a vast resource of valuable data is collecting dust on researchers' shelves. Thus there has been some progress made in the establishment of national and international data centres and databases to act as central repositories from which scientists can obtain data to provide a basis against which their own data can be analysed. These provide a reserve of data and may also facilitate otherwise impossible types of analysis. The following section discusses the role and operation of some of these data holdings with particular reference to hydrological data.

Hydrological datasets

The collection of palaeohydrological data has many parallels with contemporary hydrological archives. One of the earliest attempts to collate a resource of hydrological data from individuals for more general use was the Vigil network, established during the 1960s (Leopold 1962; Emmett 1965; Leopold & Emmett 1965; Emmett & Hadley 1968; Osterkamp *et al.* 1991). This project encourages hydrologists to collect baseline data from network drainage basins following from the recognition that 'those of us who seek to explain the variations in hydrologic phenomena... are painfully aware of the lack of adequate data' (Leopold 1962, p. 5). An aim of the Vigil project was to afford a nucleus for international cooperation in hydrology by providing a data resource in a standard format for future generations of scientists to use (Emmett & Hadley 1968). Similarly, data on very large floods collected during the UNESCO hydrological decade were compiled into a volume to facilitate continued large-scale access to the data (UNESCO 1976). Whilst databases such as these have been established to stimulate either data collection or data access for contemporary environmental systems, others have been built to facilitate data rescue (e.g. Barry 1988; Barry & Brennan 1993).

The value of assembling datasets collected by a variety of people in different areas on similar or related subjects is thus not new. More recently there has been an increase in the number of databases available in the public domain in a digital, rather than paper, format. Of particular significance has been the introduction of international computer networks, such as the Internet, which connect millions of people worldwide to an 'information superhighway' and have provided an impetus to the development of data sharing. In principle, these networks give people access to a vast resource of digital data and information and provide a potential mechanism for multidisciplinary and international data and information sharing, albeit with some significant organizational, logistic and legal issues to be confronted.

Alongside the growth of computer networks, standard procedures have been established to provide data about datasets using the directory interchange format (DIF). Networks have been specifically established to provide gateways to a number of environmentally related databases, and 'special interest networks', which are a collections of data centres which cooperate to provide a focus for research and information on particular subjects (Green & Stocker 1995), have been established. It has been suggested that whilst such networks have hitherto emphasized contemporary data, they should be established in the future as a nucleus for Quaternary research (Green & Stocker 1995).

Another consequence of the increase in the production and availability of data can be attributed to the requirement of many funding councils that data collected during a funded project must be made available within the public domain. Partly in response to this requirement, and the recognition of the value of complete datasets rather than summaries, a number of journals have joint ventures with the World Data Center-A whereby published data are archived in a digital format (e.g. Miller 1994; Webb *et al.* 1993; Webb 1994). Additionally, some journals now include 'data and analysis notes' which describe datasets or algorithms that are accessible in the public domain (e.g. Hornberger 1994). In this way researchers receive recognition for submitting data into a data centre. Researchers who subsequently use the data mentioned in the note then cite the publication.

International and multidisciplinary data sharing as promoted by these data activities can be a stimulus to development in research. Indeed, Miller (1994) argued that the exchange of data and samples that has occurred as a result of the Ocean Drilling Programme has been an important focus in the development of palaeoceanographic research, and Smol & Last (1995) suggest that progress within palaeoliminology has been stimulated through the integration of computers into the subject.

A database for global continental palaeohydrology

As stated above, there exists a wealth of hydrological and palaeohydrological data, but it is currently unavailable within the public domain. Thus it is not the *lack* of data that is the problem (as with the Vigil network) but rather lack of *accessibility* to the data in an appropriate format. In response to this lack of access, and the recognition of the importance of palaeohydrological data to environmental change research, a fundamental part of the INQUA project on global continental palaeohydrology has been the establishment of a database to collate palaeohydrological data for use by present and future scientists. The database contains data from any part of the world from the period from 20 000 years BP until the time at which continuous instrumental monitoring commenced. This database is critical to the project because, as the research focuses on all land areas and is being conducted in four major regional zones, it is not only necessary to facilitate data availability within zones but also between them.

The development of computers and digital data storage has changed the methods by which data are organised and stored. Thus, although the GLOCOPH database is founded on similar principles to the Vigil network and UNESCO dataset, in that it aims to promote access to data, it is the power of the GLOCOPH software that brings the capabilities of the database in terms of searching and analysis far beyond those of the data collections stored on paper. The database thus facilitates not only the *collation* of data but also its *manipulation* and *analysis* (see below). Table 1 compares the features of the GLOCOPH database and earlier compilations of hydrological data. This illustrates how the implementation of computer technology has improved the access, security and updating capabilities of present-day collections of data.

Table 1. *Comparison between features of the GLOCOPH database and earlier compilations of hydrological data*

Dataset	Data holdings	Storage media	Access	Storage problems	Security	Update procedures	Searching
Vigil[1]	Data from a series of experimental and representative basins, including channel cross sections, channel scour and fill, bank recession rates	Paper files, microfiche	Letter to data manager, data returned on paper/microfiche	Paper size limited by size of the folder	Copies of the files and microfiche kept in USA, Sweden and Israel	Only when site is re-surveyed and if important trend or change is observed	Card indexes show sites for which data is available and the researchers involved in the collection
Catalogue of alluvial river channel regime data[2]	Downstream hydraulic geometry data taken from the literature based on consistent methods	Digital/ paper	Report obtained from library	Limited by storage capacity of individual computer	Copies of volume available in many locations	When new data received	Data dictionary showing rivers, techniques used and sources of data
UNESCO[3]/UK flow data[4]	Data of extreme flood events throughout the world/riverflow data from the UK	Book	Book obtained from library	Limited by shelf space in library	Copies of volume available in many locations	Not possible unless book is re-published	Index and contents pages
GLOCOPH[5]	Palaeohydrological and associated data covering the last 20,000 years. Bibliography of palaeohydrology literature	Digital database	On-line via Internet or from ftp	Limited by storage capacity of computer	Tapes archived daily	Daily	Any field within the database can be queried to find related information

[1] Emmett & Hadley (1968); Osterkamp *et al.* (1991) [2] Church & Rood (1983); [3] UNESCO (1976); [4] Jones *et al.* (1986); [5] Branson (1995); Branson *et al.* (1995)

The general aims of the GLOCOPH database have been discussed in detail in Branson (1995) and Branson *et al.* (1995), and can be summarized as:

(1) to improve access to palaeohydrological data and to prevent data from being lost to the scientific community;
(2) to extend available instrumental records;
(3) to provide data collected from many sources standardized to a common metric;
(4) to improve access to data from unpublished reports and theses;
(5) to provide a nucleus for international information sharing and co-operation.

The initial tasks of the database project entailed designing the database structure and identifying existing relevant datasets from the literature, and there is now an on-going process of data entry and management and ensuring availability to users.

Data types and coverage

The GLOCOPH database has been compiled to include data covering all geographical areas and all attributes for which detailed *dated* palaeohydrological results are already available. Table 2 summarizes the types of data variables that are currently held in the GLOCOPH database. The database aims to be multi-disciplinary, thus embracing the thesis that the future of truly Global Science is by using the accumulated knowledge from a range of specialisms (Faure *et al.* 1993).

Figure 1 shows the spatial extent of data coverage within the database. Although still in its preliminary stages the database now contains data from 86 rivers worldwide and encompasses 193 sites (although at present there is a concentration in the temperate regions studied as part of the IGCP project discussed above). In this sense the database will never be complete, it is an organic, growing resource, with continual input and potentially additional variable storage responsive to new needs and techniques. Most entries have been extracted from publications and unpublished reports, although in the future it is anticipated that most records will be obtained directly from the data collector.

A high level of standardization is required if data collected at various times by different researchers are to be easily compared, and thus all data in the database are formatted to International Standard measurements. Users will not normally have access to the original unprocessed values unless specifically requested, but the calibration equations will be part of the database, and it will therefore be possible to recover the original values.

Roles of the database

It has been argued that large databases can have four distinct roles which assist the *organization*, *facilitation*, *association* and *bisociation* of data (Branson *et al.* 1995). A series of simple examples has been designed to illustrate these components using data within the existing GLOCOPH database.

The *organization* of data is the simplest function of the database, encouraging the efficient and secure storage of data. The GLOCOPH data are stored within the

Table 2. *Types of data currently held in the GLOCOPH database*

Primary data	
Drainage basin characteristics	Valley length Drainage area Floodplain width Valley width Drainage density
Channel plaform	Channel length Channel gradient Sinuosity Braiding index Meander parameters
Channel characteristics	Width Area Radius Depth Width/depth ratio Water level Flow width
Channel sediments	Channel depth/sediment diameter ratio Channel facies thickness Palaeochannel sediment thickness Overbank sediment thickness Ratio of channel overbank thickness Clast diameter
Reconstructions	
	Discharge Velocity Stream power Aggradation rate Shear stress Channel shift Sediment yield Flood frequency Flow depth Energy slope Channel deepening Entrainment values Resistance parameters

Oracle relational database system in which data are held in a series of related tables. The database structure is described in detail in Branson (1995) and Branson *et al.* (1995) and therefore only a brief description is given below

Within the relational structure the data are stored in a hierarchy. Although the structure in which data are held is generally invisible to the user it is important that an efficient data structure is designed to assist the facilitating role of the database

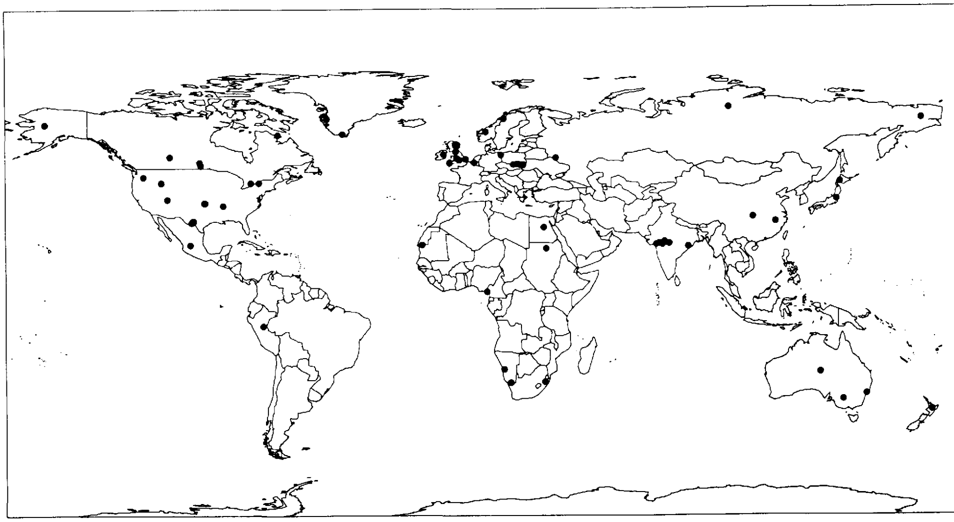

Fig. 1. Map showing the spatial coverage of data in the database.

(discussed below). In the GLOCOPH database, data are first grouped into datasets. A dataset represents an individual donation of data from a researcher or data from related publications by the same author(s). Table 3 shows output from the database which describes an individual dataset. The data from individual rivers within a particular dataset are stored separately within the database, and the river data are further subdivided into reaches and then into sites; a site represents, for example, a single slackwater deposit or channel cross-section. The lowest level of classification is the record which represents a single measurement location (a level within a sediment core, or boulder deposit, for example). At this level all primary data, reconstructions and date information are stored. The database also has facilities to include images, such as site photographs, maps or graphs, and there are routines which format site location data, so that they can be imported directly into site viewing packages currently available in the public domain (e.g. Keltner 1995).

All data are stored within the central database; this centralised approach has several advantages over storing the data in unconnected databases managed by different groups namely:

(i) there is easier access to a variety of data – a form of 'one stop' data shopping;
(ii) a more complete data coverage is possible from the combination of data resources;
(iii) there are lowered overall costs due to reduced duplication of effort;
(iv) data can be updated concurrently.

The *facilitation* role of a database represents the promotion of information about the data and access to them, and this relies heavily upon the efficient organisation of the data. At a simple level the facilitating role of the database enables users to identify the research that has been undertaken in a particular area and the types of data that are available. The next level allows the data of interest to be accessed and extracted. Beyond this retrieval of data the GLOCOPH database also facilitates the modelling

Table 3. *Output of data from the dataset table for an individual dataset*

Name of dataset	River channels changes and palaeodischarge estimates for the Warta River, Poland[1]
Principal investigators	
Name	S. Kozarski and P. Gonera
Address	Adam Mickiewicz University, Poznan, Poland
Data compiler	
Name	J. Branson
Date of compilation	01-10-94
Source of data	Publication
Spatial and temporal coverage of dataset	
River	Warta
Country	Poland
Study location	Reach 1 comprises 6 palaeomeanders between Ksiaz Wielkopolski and Czom Reach 2 comprises 3 terraces, studied between Nowe Miasto and Mosina
Latitude	52.25° N
Longitude	16.56° E
Period of data	27 500 BP – 2490 BP
Summary of dataset	
Techniques	Cores of palaeochannel sediments, dated by radiocarbon methods. Velocity and discharge calculated based on measurement of channel parameters and bed load deposits.
Raw data	Channel width, depth, area and gradient. Sediment size. Meander curvature
Reconstruction	Discharge, velocity, stream power
Summary	Detailed study of river channel changes in the Warta valley. Data includes channel dimensions, channel planform data and reconstructions of discharge, velocity and stream power.
Keywords	Terrace, palaeochannels, discharge, velocity

[1] Source of data: Kozarski (1991).

and analysis of the data (which is an integral part of the association and bisociation roles of the database discussed below).

Access to the data within GLOCOPH database is currently being made available on an interactive basis using the Oracle software or through the Internet. Alternatively, individual datasets can be obtained via file transfer protocols or on floppy disc (at cost of media). When accessed interactively data are viewed through custom-built forms which use a series of menus. There are programmed query structures to allow searches to be made to obtain data corresponding to particular requirements, concerning, for example, an individual river, reach or record, at a particular date, data type, technique or researcher. Information about the data is given in a series of 'metadata' fields and tables, which describe the techniques used to collect the data and errors associated with them. Providing a high level of metadata should ensure that the data are not unknowingly used for analysis outside the context in which they were collected.

One of the strengths of a collection of data obtained from researchers who have worked in different areas throughout the world is that it is possible to search for patterns or coincidences of changes at a range of temporal and spatial scales, and this is categorized as *association* (relationship between variables in a single domain) and *bisociation* (relationship between variables drawn from contrasted temporal, spatial or attribute domains). It is envisaged that the value of the collated resource of data should therefore be greater that the sum of the individual datasets, and that it may be possible to develop explanations that might not be evident if datasets from only one or a limited number of sites were to be analyzed. Within the GLOCOPH database association and bisociation of data are promoted through the provision of routines which can apply a variety of simple hydrological models to the data with which to calculate discharge, velocity and similar parameters.

To demonstrate the potential usage of the global database two examples are provided which graph the results of queries of physical and timeframe variables within the database. Figure 2 comprises palaeodata extracted from the database to indicate the palaeodischarge values for a number of river systems during the last 16 000 years. Figure 3 represents the reconstructed palaeoflood discharge of river systems from 20 000 years BP until present.

Here, it is not appropriate to explore the possible reasons for the correspondence between the data from different regions, but rather to illustrate the potential of combining the research results on a global scale, which can then be used as a basis

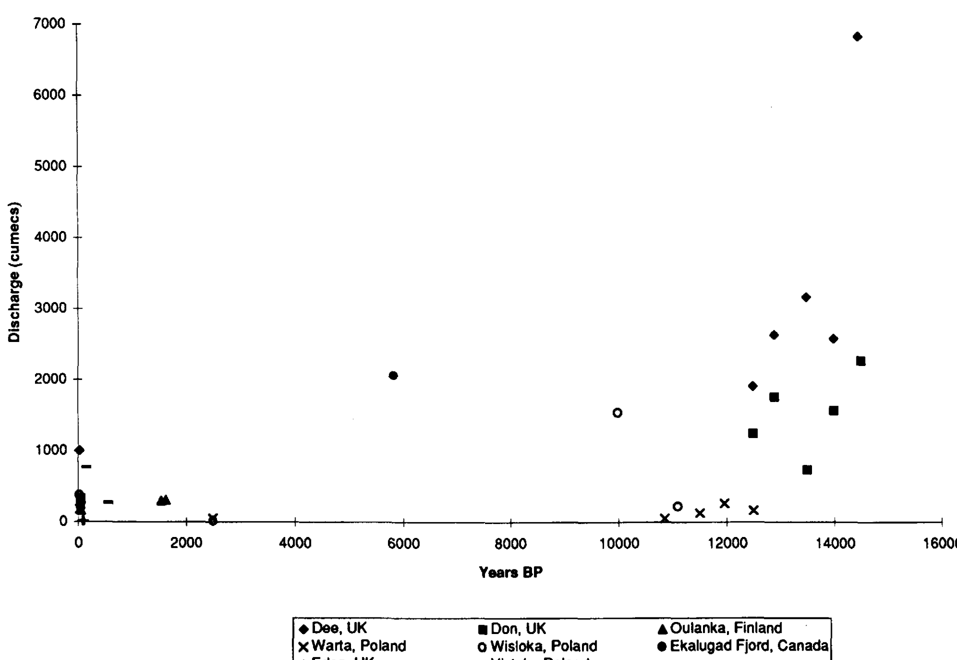

Fig. 2. Palaeodischarge values extracted from the database for selected rivers from 20 000 years BP. Original data from Church (1978), Froehlich *et al.* (1977), Genera & Kozarski (1987), Jones (1984), Klimek (1987), Koutaniemi & Ronkainen (1983), Kozarski (1991), Maizels & Aitken (1991), Mason & Beget (1988).

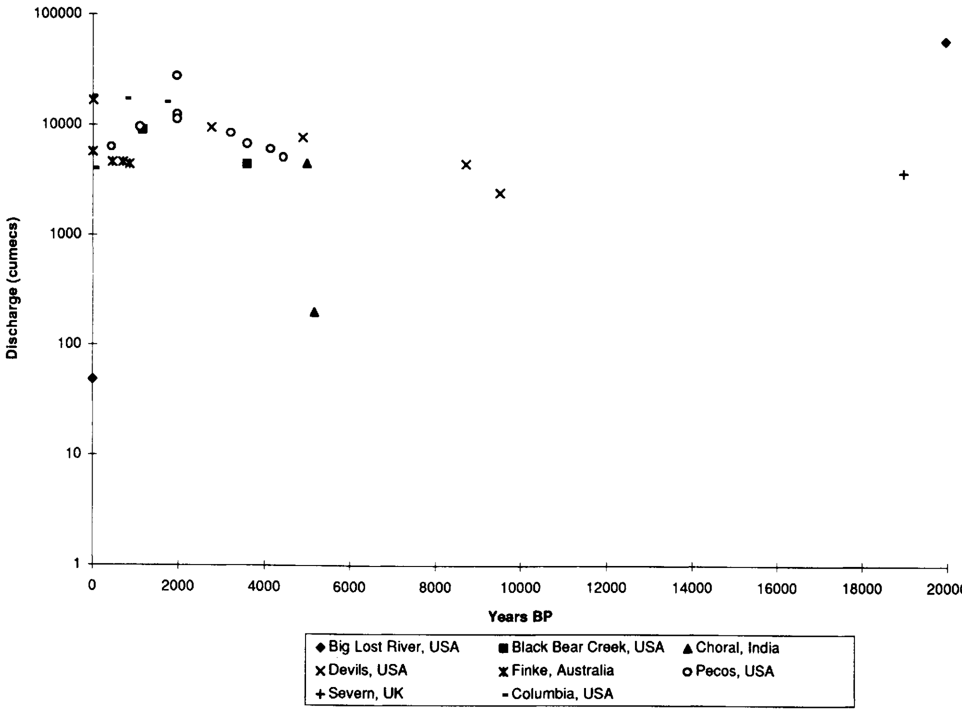

Fig. 3. Palaeoflood discharge values extracted from the GLOCOPH dtabasae for selected rivers. Original data from; Baker *et al.* (1979), Chatters & Hoover (1986), Dawson (1989), Kale *et al.* (1993), Kochel (1981, 1988), McQueen *et al.* (1993), Patton & Dibble (1982), Pickup *et al.* (1988), Rathburn (1993).

for further investigation of the association and bisociation of change both within individual river systems and between river systems.

Conclusions

Palaeohydrological data can be of interest for analyses of past, present and future environmental change. As part of the GLOCOPH project a database has been established to collate palaeohydrological data from researchers worldwide. The database fulfils several roles, and at the simplest level facilitates the collation and storage of data, but beyond this the real strength of the database is that it allows value to be added to individual datasets in terms of analysis and association with other datasets (potentially derived from other palaeoscience databases covering varied environmental subjects).

In many ways the database uses technology to do more rapidly what could have been done using more traditional slower methods. Once created, however, the database can support the INQUA Commission by making it possible to develop derived data and to run and devise other models that would not have been possible without the resource and access to data that are available within the database.

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The Past Global Changes (PAGES) Project

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Abstract: PAGES is a core project of the International Geosphere Biosphere program. Its role is to provide a quantitative understanding of the Earth's past environment and the envelope of natural variability within which we can assess anthropogenic climate changes. The Project is organized into two main time frames: STREAM 1 which covers the last 2000 years and STREAM 2 which covers the last two glacial/interglacial cycles. Within these time frames PAGES has organized a series of Activites and Tasks, the largest of which are three transects from the Arctic to the Antarctic, one through Europe and Africa, one through Siberia, China and Australia and one down the west side of the Americas. In each case PAGES is seeking to co-ordinate the activities of local researchers and bring together scientists from different research backgrounds: tree-rings, corals, deep sea sediments, ice cores, for example, to achieve the scientific goals of the three transects. PAGES is not a funding agency and does not support research, but rather its function is to co-ordinate and facilitate and promote research on a larger scale than any one country or funding agency can support and on topics that go beyond the expertise of any one group of scientists.

PAGES is a Core Project of the IGBP, the International Geosphere–Biosphere Programme. IGBP is itself sponsored by ICSU, the International Committee of Scientific Unions; as are SCAR, SCOR and other global scientific organisations. The objective of IGBP is:

> to describe and understand the interactive physical, chemical and biological processes that regulate the total earth system, the unique environment it provides for life, the changes that are occurring in this system, and the manner in which they are influenced by human actions.

Within the vast field that this remit encloses, IGBP has set priorities in those areas:

> that deal with the key interactions and the significant changes on timescales of decades to centuries;
> that most affect the biosphere;
> that are most susceptive to human perturbation;
> that are most likely to lead to practical, predictive ability.

Note particularly the stress on the biosphere, the timescale and the fact that the results need to be useful.

IGPB has established a number of core projects (Anon, 1994) some of which will now be familiar acronyms such as BAHC (Biological Aspects of the Hydrological Cycle), LOICZ (Land–Ocean Interactions in the Coastal Zone) and PAGES (Past Global Changes).

The role of PAGES within IGBP

Why is there a need for PAGES within IGBP? In order to understand many aspects of the global system we need to study changes on a longer time scale than recent instrumental measurements provide. We also need to understand more about the functioning of the system in times before human interference. Some specific questions that PAGES is focusing on follow.

(1) How has global climate and the Earth's natural environment changed in the past? What factors are responsible for these changes and how does this knowledge enable us to understand future climate and environmental change?

(2) To what extent have the activities of humans modified climate and the global environment? How can we disentangle anthropogenic-induced change from natural responses to external forcing mechanisms and internal system dynamics? What were the initial conditions of the Earth system prior to human intervention?

(3) What are the limits of natural greenhouse gas variation and what are the natural feedbacks to the global climate system? In what sequence, in the course of environmental variation, do changes in greenhouse gasses, surface climate, and ecological systems occur?

(4) What are the important forcing factors that produce climate change on societal time scales? What are the causes for abrupt climatic and environmental events and the rapid transitions between quasi-stable climatic states which occur on decadal to century time scales?

The scientific role of PAGES

PAGES provides leadership and guidance to scientists and provides a framework of collaboration that can enable projects on a larger scale than the individual scientist or small research team can contemplate. PAGES does not tell scientists what to do and it is not a funding agency. PAGES facilitates the international and interdisciplinary scientific efforts that are required to address global-scale questions.

PAGES can:

- provide an international framework for palaeoscience;
- provide visibility for palaeoscience;
- demonstrate a coherent community;
- direct funding attention and resources to palaeoscience activities;
- provide opportunities for collaboration;
- promote planning and identify data gaps;
- encourage consistent analytical techniques.

A framework for research. PAGES can design and co-ordinate projects that go beyond national boundaries. Climate does not respect our political divisions. There is also much science on global scales that is too expensive or too logistically complex for a single nation to attempt.

Visibility and publicity for palaeoscience both in the rest of the global change community and also to the general public. This is important for the political implementation of global change-related policies by governments, but is also important at a more local level in the promotion of conservation of the natural resources that are the life blood of palaeoecological studies. Examples are the preservation of peat bogs in Denmark and Ireland, and the salvage of ancient tree trunks in Ireland, Tasmania, Florida, and California.

A large and coherent research community able to tackle projects the size and complexity of the IGBP core projects which are not possible without a coherent approach. Such projects need the local experience from many countries together with the technical expertise of specialists not necessarily from those countries.

PAGES can influence funding directions. This is an important function as PAGES has no research money of its own. In recognition of the very wide range of international expertise that PAGES can call upon, agencies funding global change programmes in many countries now use the PAGES framework as a guideline for their support of global change research.

PAGES can identify data gaps. By assembling teams of international researchers on a thematic or geographical basis, it is possible to see clearly where the research gaps are. In the PEP (Pole–Equator–Pole) transects described below, these gaps are geographical, however, other PAGES workshops have identified technical and thematic gaps. There are gaps in the availability of data from past research that are being tackled by the data banks (see below and Branson *et al.* this volume). There are also communication gaps, for example at a recent data conference the differences in approach and in recording between the European and Chinese historical data projects were highlighted and moves made to bring the two methodologies onto a compatible basis.

PAGES can improve field and analytical methodologies – not by dictating, but by bringing people together to share. PAGES recently hosted a meeting of those involved in studying stable isotopes in wood and peat and a meeting of those involved with a range of chronometric techniques applicable to PAGES tasks. PAGES published the 'PALE protocols' which attempt to standardise field and laboratory methods for palaeoclimatological work in the Arctic regions.

PAGES Activities

The overall PAGES Core Project is subdivided into five Foci (Table 1), each with various Activities (Eddy 1992; Anon 1994). Within each activity there are specific research tasks. All PAGES research is divided broadly into two time streams.

The objective of temporal Stream I is to reconstruct the detailed history of climatic and environmental change for the entire globe for the period since 2000 years BP, with a temporal resolution that is at least decadal and ideally annual or seasonal. This constitutes the period of man's greatest human impact on the planet and the time of significant overlap between written records and the environmental information stored in natural archives.

Table 1. *PAGES Project organisation*

Focus 1: global palaeoclimate and environmental variability (the PANASH Project)	
Activity 1.1:	PEP 1 The Americas transect
Activity 1.2:	PEP 2 Austral-Asian transect
Activity 1.3:	PEP 3 Afro–European transect
Activity 1.4:	The Oceans
Activity 1.5:	PAGES–CLIVAR interactions
Focus 2: palaeoclimate and environmental variability in polar regions	
Activity 2.1:	Arctic programme
Activity 2.2:	Antarctic programme
Focus 3: human impacts on past environments	
Activity 3.1:	human impacts on fluvial systems
Activity 3.2:	human impacts on terrestrial ecosystems
Focus 4: climate sensitivity and modelling	
Activity 4.1:	climate forcing and feedbacks
Activity 4.2:	climate model – data intercomparisons
Focus 5: cross-project analytical and interpretative activities	
Activity 5.1:	chronological advances
Activity 5.2:	development of new climate or environmental proxies
Activity 5.3:	international palaeodata system (w/WDC-A)
Activity 5.4:	regional, educational and infrastructure efforts with START, IAI

With at least century scale resolution, temporal Stream II focuses on glacial–interglacial cycles of the last several hundred thousand years and concentrates on understanding the dynamics that cause large-scale natural variation. Stream II activities illuminate the interactive feedbacks among various components of the Earth system and their relation to external climatic forcing.

PANASH (Palaeoclimates of the Northern and Southern Hemispheres) was the first theme identified by PAGES. The goals of PANASH are:

to document how climatic records from the two hemispheres are interrelated in amplitude, phase and geographic extent;
to determine the records of potentially important forcing factors which may have affected each hemisphere.

PANASH has crystallized more recently in the establishment of a series of activities known as the Pole–Equator–Pole (PEP) transects. The transects run broadly as shown in Fig. 1. The lines are a guide and will not exclude useful data close to the line e.g. PEP III would include data from a range of European sources, for example the Historical data base and the tree rings for the Stream I timescale and the long French pollen cores and the GRIP and GISP2 Greenland Summit ice cores for Stream II. In Africa the IDEAL (International Decade for East African Lakes) project will generate data mostly on the long timescale. Similarly for PEP II the palaeoecological aspects of the Lake Baikal drilling project will be included. For each transect there has already been a planning meeting, an assessment of what data are now available and of where the key efforts are needed. Grant proposals are now reaching funding agencies with projects that will specifically address the gaps in knowledge.

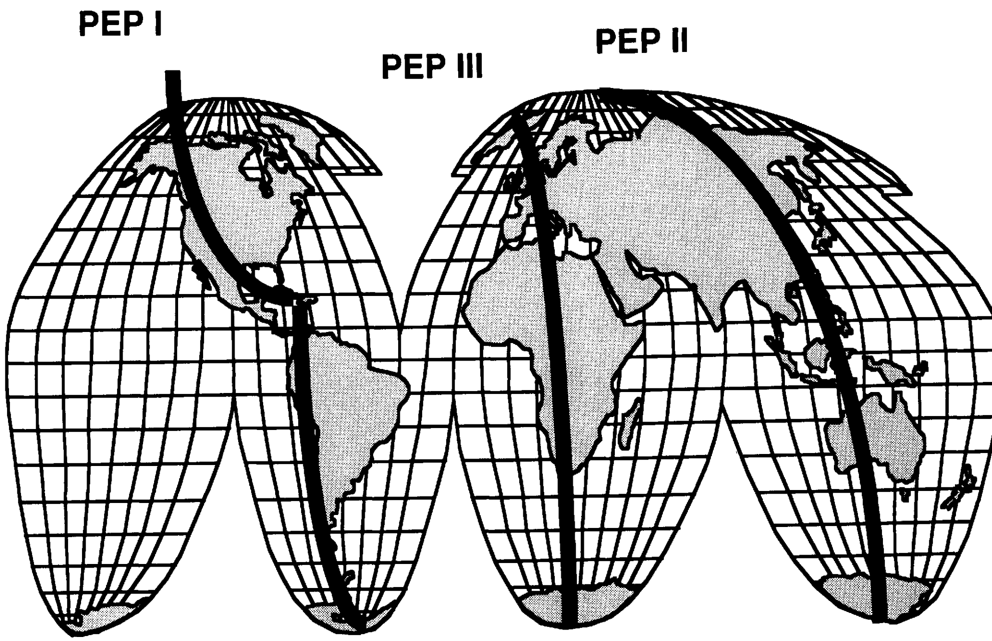

Fig. 1. The three Pole–Equator–Pole transects of PAGES.

FOCUS V Cross project activities

This focus is on the technology and scientific infrastructure required to achieve PAGES goals. As an illustration of one aspect, some of the progress on making palaeo-data available are described below.

The World Data Centers (set up and managed locally for ICSU) are depositories for data that are available to all scientists.

The WDC-A holds data for marine geology and geophysics, solid earth geophysics, solar terrestrial physics, glaciology and palaeoclimatology. At present the palaeoclimate data are held in the World Data Center-A at the National Geophysical Data Center, NOAA, Boulder, Colorado. Data are resources that must be managed in a way that provide the entire global change community with the tools to access easily and manipulate the information needed for a particular investigation. The World Data Center and PAGES are working towards perfecting a system that will co-ordinate data acquisition, management and distribution. This will include browse and visualisation tools that will make access to a huge range of data simple and intuitive.

As a demonstration of the use of this data bank, tree-ring records have been extracted for sites in France and Ireland for a specific time. Equally the ice core results from Greenland or Antarctica, tree-rings from Tibet or Tasmania, pollen or many other records relevant to studies of past climate could have been extracted. Figure 2 shows tree rings from Ireland and France for a few years either side of 1815 when the Tambora eruption in Indonesia affected global climates causing a cold summer in Europe. In the graph we can see how a volcanic event in Indonesia

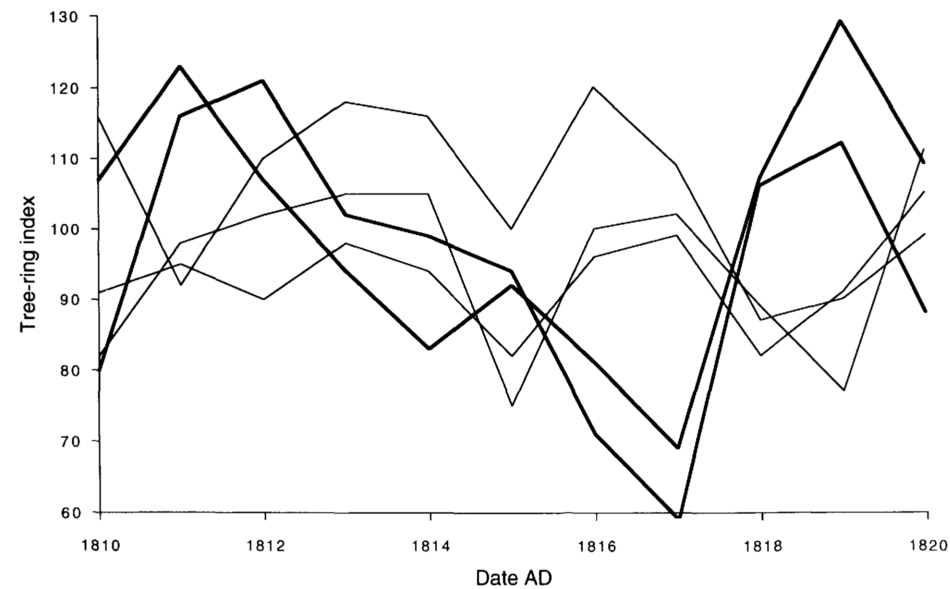

Fig. 2. Widths of oak tree-rings from France and Ireland following the eruption of the volcano Tambora in November 1815. Data from the palaeodata bank at the World Data Center A.

resulted in poor growth in Ireland in 1816 but increased growth in central France where hot dry summer conditions normally limit growth.

Conclusion

PAGES offers a central co-ordinating body for Palaeo science. It does not fund research, but, in many countries, funding bodies are taking notice of PAGES and other IGBP goals and targeting their global change funding in these directions.

For further information and for the PAGES Newsletter contact PAGES Core Project Office at: Bärenplatz 2 CH-3011 Bern, Switzerland.

To access the collections of the World Data Center for Paleoclimatology use:

FTP	ftp.ngdc.noaa.gov
World Wide Web	http://www.ngdc.noaa.gov/ngdc.html
Gopher	gopher.ngdc.noaa.gov

or if these are not available to you contact Bruce Bower at NOAA/NGDC, Mail Code E/GC, 325 Broadway, Boulder, Co 80303-3328, USA.

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Palaeohydrology: prospects and future advances

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Since Stanley Schumm defined the subdiscipline of palaeohydrology in the 1960s and 70s (Schumm 1965, 1968, 1977) it has grown considerably and broadened producing its own sub-disciplines. Palaeohydrology has even become established long enough for both its history to be documented (Patton 1987) and for its conceptual underpinnings to be cogitated upon (Baker 1991). These palaeohydrological subdisciplines include; water cycle modelling, palaeogroundwater studies, regime-based palaeoflow studies (Baker 1991), palaeohydraulics and stable-boundary palaeoflood analyses particularly of slack-water deposits – palaeostage indicators (the SWD-PI of Baker 1974 and Baker *et al.* 1983). In addition wider themes of geomorphological interest such as floodplain formation, alluvial stratigraphy and palaeoecology and the study of alluvial chronologies have all been drawn in to the broad theme of palaeohydrology (Starkel *et al.* 1991; Gregory *et al.* 1987). With the help of various international bodies, such as IGCP (International Geological Correlation Programme) and INQUA (International Union for Quaternary Research), strong links have been forged between palaeoecology, palaeolimnology and fluvial palaeohydrology.

The reasons for doing palaeohydrology are generally implied rather than explicitly stated and it has been assumed that the fundamental aim is the reconstruction of past climates, but this is frequently not the case and we can identify at least five reasons for attempting palaeohydrological reconstructions.

(1) First and foremost in order to reconstruct the water balance and flows of periods prior to the collection of continuous hydrological records.

(2) It is a more rigorous methodology for the interpretation of fluvial sedimentary sequences. The steps made have to be explicit as do the assumptions and at least notional numbers given to sedimentary and hydraulic parameters. The investigation of the sources of error and lessons learnt in the act, are important components of the reflexive nature of scientific work including work in fluvial sedimentology.

(3) The form, magnitude and rates of sedimentary response to past changes in the environment, however caused, can teach us much about the inherent dynamics of the fluvial system and provide indications of the possible responses to future changes.

(4) As part of the reconstruction of past environments palaeohydrology can produce data of value to others interested in the nature of past environments and resources, such as geologists or archaeologists.

(5) The possible validation of climate models or the inference of past hydroclimatic characteristics.

Arnell (this volume) draws attention to the problems of using palaeohydrology to construct analogues for future greenhouse-gas induced climate change but he acknowledges its potential for understanding the links between climate and hydrogeomorphology (3 above) and possible validation of GCMs at the regional scale (5 above), although this requires data on seasonal or annual totals, which are rather more elusive parameters than bankfull discharge or flood magnitude. Given these justifications there are probably four broad avenues along which palaeohydrology is likely to develop during the next two decades. The first is the dating of fluvial events, the second is the role of non-fluvial variables on the thresholds of fluvial change (particularly vegetation) and a third is the application of computer models to palaeohydrology. Fourthly palaeohydrology is likely to figure increasingly in the estimation of probable hydrogeomorphological implications of future climate change scenarios (Starkel 1993).

Dating fluvial events

Palaeohydrology has always been dependent upon the dating of fluvial units since a chronologically floating event of known magnitude cannot be related to causal factors and cannot be assigned a probability of recurrence. Traditionally ^{14}C has provided the chronological backbone of most studies. Whilst this is likely to remain the case for some years to come there are significant deficiencies in radiocarbon-dated chronologies, not withstanding the contributions of accelerator mass spectrometry and tandem accelerator mass spectrometry. Firstly being dependent upon the preservation of organic ^{14}C is both climatically and ecologically biassed to boreal and temperate regions. Suitable material does occur in arid and semi-arid environments but its distribution is highly biased by groundwater conditions and the palaeoclimatic record itself, i.e. dry to wet shift are easier to date than wet to dry shifts. Secondly the wiggly nature of the ^{14}C calibration curve and the existence of plateaus limits precision for some critical periods including the late glacial-Holocene transition (Vandenberghe 1995; Brown 1996*a*). Thirdly the material that is being dated is either debris carried by floods which may be reworked or represent stand-still phases between floods. Either way the ^{14}C determination is a maximum age and does not date the event itself even when the units are sandwiched by dated samples. Fourthly and most obviously, ^{14}C dating is still limited to the last 40–50 ka. For all these reasons several recent studies have used alternative dating methods including palaeomagnetic dating (Brown & Ellis 1996), and luminescence dating.

Palaeomagnetic dating seems to work in two situations, first in relatively uniform sandy silts of overbank origin which have been rapidly accreted and relatively little bioturbated, and where the natural remanent magnetism is carried by detrital magnetic grains. Such sediments have recently been dated from small river valleys north of Rome (Brown & Ellis 1996). The second situation is where palaeochannels have been abandoned and have a silt-clay infill deposited in organic-rich stagnant anoxic conditions where the natural remanent magnetism is carried by iron sulphide minerals probably biogenically precipitated. Magnetic properties such as the anisotropy of susceptibility can also yield a wealth of information on sedimentary conditions and produce estimates of palaeocurrent direction and strength from suitable sediments, generally palaeochannel infills.

Thermoluminescence has been applied to river terrace deposits outside the range of radiocarbon dating (Nanson *et al.* 1991). When first applied to alluvial sediments, doubt was cast over whether they had been sufficiently zeroed to give the date of deposition (Berger 1984) and research is proceeding on techniques to recognise when this has occurred and its magnitude (Rhodes & Pownall 1994). Due to the dynamics of the fluvial system and the generally short period of exposure of the sediment, the thorough penetration of light for all grains cannot be assumed (Bailiff 1992). However, Nanson & Young (1987) have produced thermoluminescence dates from terraces of the Nepean river in New South Wales and from an alluvial sand below the floodplain of Coopers Creek in western Queensland which agree well with other dating evidence. Their conclusion is that, if the dates are from shallow sand flows deposited as sheets on floodplains, where residual thermoluminescence and longterm sediment moisture contents can be accurately estimated, thermoluminescence dating is reliable and can provide an excellent basis for alluvial chronologies. A similar technique, optical stimulation luminescence (OSL), probably has a wider application to alluvial deposits and requires less sophisticated equipment (Huntley *et al.* 1985; Bailiff 1992). Visible light can be used to stimulate luminescence in quartz and infrared radiation used with feldspars. As with thermoluminescence, many factors affect the accuracy of 'apparent' luminescence ages, including mineralogy, light exposure before burialdarkening, overburden, erosion, and the nature of the radiation dose. This last factor is itself affected by sediment water content. Another related chronometric method which has palaeohydrological potential is electron spin resonance (ESR). Further refinement of these techniques on alluvial sediments will hopefully allow the reliable direct dating of flood sediments unbiased by climatic or sedimentary regime.

Causes of variable response

Any attempt to use palaeohydrological reconstructions, especially of flood history, for the determination of climate change faces serious hurdles. One is climatological and stems from our understanding of the links between extreme events and the dynamics of the climate system, whereas the other is fundamentally geomorphological. This stems from the wide variation in catchment conditions (size, geology, relief, vegetation, geomorphic history etc.) which might be expected to produce different non-linear responses and lags, making the direct linkage between extreme events and climate change difficult. In order to be able to overcome these problems and avoid imposing preconceptions upon exceedingly complex data or making statistically and mechanistically dubious teleconnections we will continue to require more sub-regional studies of flood-climate linkages. One of the ways of accomplishing this is through the study of particular internal factors and thresholds whilst other factors are held constant. That important internal thresholds may influence fluvial response has been known for many years (Schumm 1979) and Knox (1993) has demonstrated the non-linearity of flood response to climate change. However, the role of non-fluvial factors in the determination of these thresholds and non-linearities has received relatively little attention and one of the most important factors is vegetation. In addition to catchment influences vegetation will effect fluvial response to climate change through its effect on bank stability and channel roughness, in some cases significantly altering the form and number of channels. Our contemporary engineered and cleansed channels have little in common with channels of the past. As research on rare forested floodplains has shown tree throws and debris dam formation can significantly alter channel form and will therefore alter the processes and rates of channel change and channel sedimentation – the basic data for palaeohydrology. This work has shown that forests can help maintain multiple channel systems (Harwood & Brown 1994; Brown *et al.* 1995), alter channel planform and bank erosion (McDonald *et al.* 1982; Davis & Gregory 1994), control sediment routing (Gurnell & Gregory 1984), initiate bar formation (Nanson 1981) and with the supplementary action of beavers even create flow-through lake systems (Coles 1992). Although it has not been recognised until recently detailed stratigraphic investigations of fluvial sections can reveal evidence of some of these processes (Brown *et al.* 1994). The vegetation of the floodplain is therefore an important variable in palaeohydrological modelling and will need to be explicitly included in future studies.

The palaeohydrological role of vegetation at the catchment scale is also now receiving more attention. New models of river adjustment to the dramatic climatic changes that have characterised isotope stages 4–2 and the transition to the Holocene, stress the effects of vegetation. Vandenberghe (1993) has put forward a model for the periglacial zone based on stratigraphic, dating and palaeoecological work in the Dinkel, Reusel and Mark valleys in the Netherlands. The essentials of the model are as follows.

Beginning of cold stage. Fluvial erosion caused by delayed degradation of vegetation limiting slope erosion and failure despite declines in evapotranspiration and an increase in runoff.

Maximum of cold stage. Gullies aggrade due to high sediment supply and high seasonal runoff.

Transition to warm period. Incision is caused by low evapotranspiration due to the delay in vegetation development and a reduction of sediment supply but an increase in runoff.

In the case of the Dutch valleys this is manifest by cut and fill cycles and a change in river behaviour for meandering to braiding and back to meandering. This model seems applicable across northern Europe and north America with perhaps some accommodation in some areas for the inheritance of form and disequilibrium caused by very rapid climate change as illustrated by the development of anastomosing systems during the early to middle Holocene (Brown 1996*b*). Any models in which vegetation significantly effects the relationship between climate and fluvial change are sensitive to the assumptions or data concerning vegetation climate linkages. This is not straightforward as illustrated by the changing views on the rates and causes of vegetation development during the Holocene in north west Europe (Brown 1996*b*). It will therefore be necessary for palaeohydrologists to work closely with palaeoecologists and preferably to derive new ecological records for the catchments which they intend to model.

The application of models

Computer models of hydrological systems have over the last few years become far more numerous and user friendly. Computer models have applications in both regime-based palaeohydrology (RBP) and slack-water deposits palaeohydrology (SWD-PI) for the estimation of channel forming flows from alluvial channel dimensions and palaeoflood discharges in stable reaches. One of the most commonly used models is HEC-2 step-backwater program (US Army 1990) which computes water surface profiles through channels and stream networks for a steady, gradually varied flow using the standard step method. This is particularly applicable to SWD-PI estimation but has also been used in situations with deformable beds (Keller & Florsheim 1994). User-friendly groundwater models are also becoming increasingly available and they have several potential applications in palaeohydrology (Brown 1995*a*). The first and most obvious is the modelling of open and supposedly closed (Brown 1995*b*) lake response to climate change and particularly the role of lake-groundwater flux. The second is the effluent contribution of groundwater to rivers especially those on permeable lithologies and in semi-arid environments. A third and rather more difficult application is to the preservation of organic sediments and pedogenetic processes within floodplains. The problem here lies in that although the water table height and variation is the most important pedogenetic variable in floodplains its effects on organic matter, iron chemistry and soil characteristics is poorly understood due to the complexity of the soil system and the importance of microbiochemical processes. An example of such a groundwater model is MODFLOW which is a three-dimensional finite-difference groundwater flow model, developed by the US Geological Survey (McDonald & Harbaugh 1988). Separate components of the model simulate recharge, evapotranspiration, flow to wells and drains, and exchanges with rivers. MODFLOW has been used to estimate the conductivity of river beds (Yager 1993) and the effects of wetland creation on floodplain watertables and river flow (Hensel & Miller 1991), both applications are of relevance or applicable to palaeohydrology. Bradley & Brown (1992, 1995) have used MODFLOW to model the hydrology of a small alluvial wetland, one of the aims of this work being to predict the affects of climate change upon floodplain hydrology including effluent flow and ecology. Given a known palaeo-topography and stratigraphy the methodology would be equally applicable to past periods.

A further possibility is the use of models coupling groundwater flow and stream networks. An example is the work by de Vries (1994) on the sandy Pleistocene area of the Netherlands where a zone of groundwater exfiltration leads to the development of a surface drainage network through sapping. The stream network is determined by the discharge capacity that is necessary to transmit surface precipitation through the hydrological system. Another possibility is the combination of groundwater models and digital terrain models (Arnold *et al.* 1993) and a particularly suitable model is TOPMODEL which Durand *et al.* (1992) have used on Mediterranean catchments. The input to these types of model can as easily be past climatic estimates as future estimates.

In theory the modelling of the fluvial response of a catchment several thousand years ago is comparable to the modelling of an ungauged catchment response to recent or contemporary regional climate data. If catchment changes are to be included in models of fluvial response to postulated climate change scenarios, an obvious development is the use of relatively simple non-distributed models of unguaged catchments such as those developed by Pirt & Simpson (1983). Although somewhat speculative (although not more-so than much other scenario-modelling) the use of linked atmosphere-runoff models may be a useful tool in the testing of climatic hypotheses against palaeohydrological data. An example of this class of model is the coupled synoptic-hydrological model CLAM (Wilby *et al.* 1994). This uses Lamb's weather types (see Rumsby this volume) to generate precipitation data which is used as input to a semi-distributed catchment model. Again, just as future probabilities of weather type series can be used so can past weather series for which there is relatively good historical data in several parts of the Old World. These models may also ultimately allow us to link channel palaeohydrology with catchment conditions.

Applied palaeohydrology

Climate change largely impacts on people through the hydrological system. Since palaeohydrology is concerned with the relationship between climate-catchment and hydrological response it is central to the prediction of the probable impact of climate change, whether natural or human induced. Two examples will suffice, the first of which is the estimation of the probable maximum flood (PMF). Since the maximum flood vs basin area curve is climate dependent small changes in the incidence of high-flood (or drought) producing synoptic conditions may change the PMF (Acreman 1989). The estimation (cf. retrodiction) of historical floods has already caused the revision of PMF for the river Tyne in England (Archer 1993) and any revision of the PMF will in turn require the updating of flood-risk maps which are statutorily required in many countries (Arnell 1992). Similarly design floods may have to be revised for the calculation of dam and bridge safety factors. Palaeohydrology and particularly palaeoflood analysis is currently producing data that is of particular relevance in the classic debate over the existence of absolute maximum limits or 'natural upper bounds' to flood series for catchments of a given size (Georgiadi 1979; Georgiadi 1993; Enzel 1993).

A second area of application is in the mobilization of contaminants held in alluvial sediments. Mining in all parts of the world has produced mine wastes much of which has gone into river systems and been locked-up in alluvial and estuarine sediments. This is particularly significant where mining has been hydraulic such as in traditional methods of streaming for tin and gold and more recently in hydro-jet mining for gold. Studies of alluvial sediments in England have shown how the history of lead and zinc mining can be recorded very accurately in alluvial units (Macklin 1985; Macklin *et al.* 1992). The problem is that heavy metals may be at concentrations high enough to be toxic to aquatic life, and in some cases there may be even more toxic elements which are byproducts of mining, an example being arsenic which is released along with tin during tin streaming (Bradley 1989). Even after the mining has long since ceased remobilisation of these contaminants can significantly effect water quality. The mobilization of contaminants is a function of the rate and locations of channel change and, other factors being equal, this is closely related to the flood series as illustrated by Hooke's work on the Gila in Arizona (Hooke this volume).

As Baker (1991) put it, there would seem to remain a 'bright future for old flows' and whilst the accuracy of numbers given to these flows must always be viewed in the light of the precision and assumptions of the methods used, the act of estimation is an important part of the 'discovery of the workings of important hydrological phenomena as elucidated by their operations in the past'. The short-term future of palaeohydrology is likely to be characterized by specialist contributions to large inter and multi-disciplinary projects concerned with environmental change whilst the longer-term future will hopefully see palaeohydrology as an essential element in theories of landscape evolution and the relative role of internal and external factors in fluvial adjustment to climate change.

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